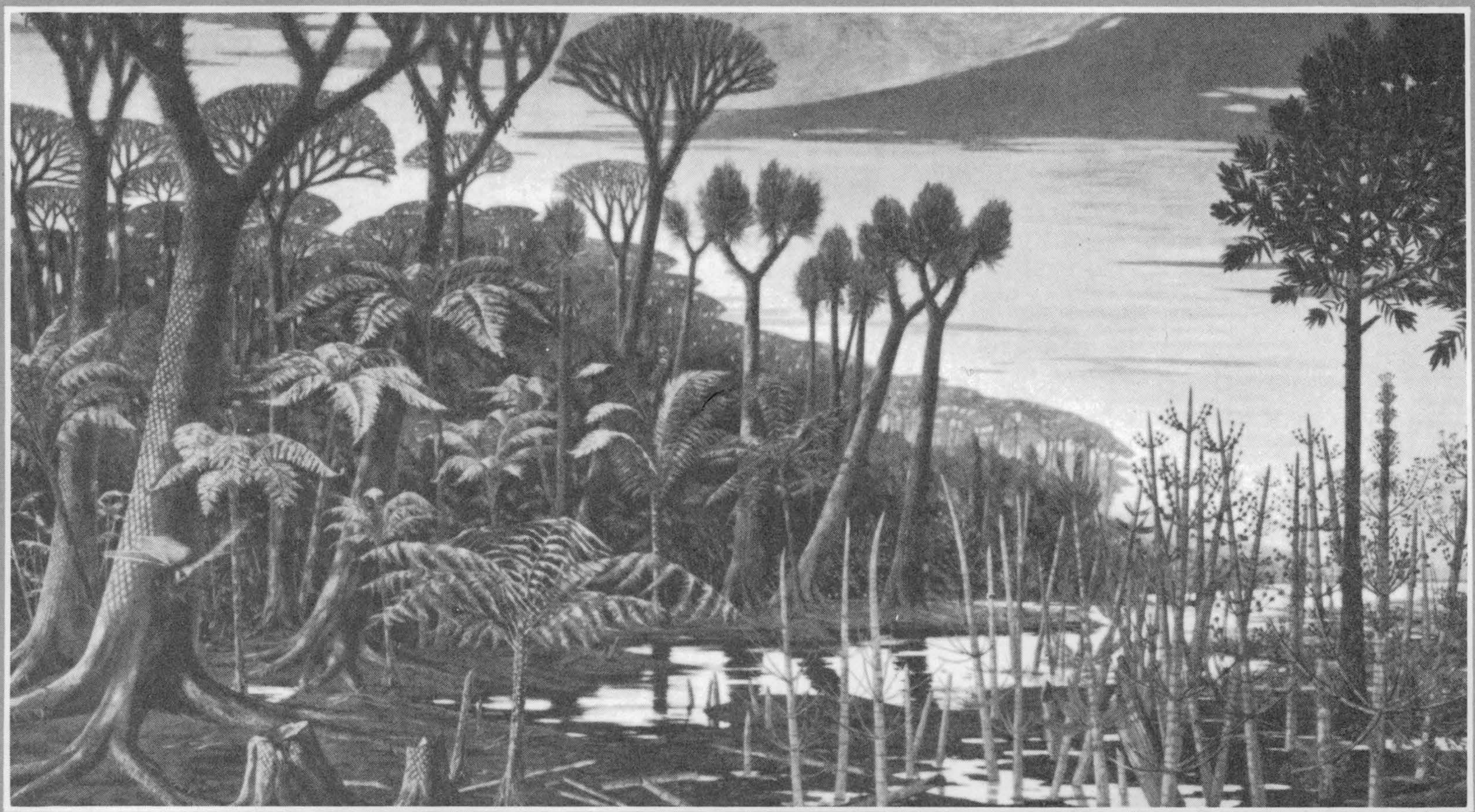
The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States



GEOLOGICAL SURVEY PROFESSIONAL PAPER 1110–M–DD

ON THE COVER

Swamp-forest landscape at time of coal formation: lepidodendrons (left), sigillarias (in the center), calamites, and cordaites (right), in addition to tree ferns and other ferns. Near the base of the largest *Lepidodendron* (left) is a large dragonfly (70-cm wingspread). (Reproduced from frontispiece in Kukuk, Paul (1938), "Geologie des Niederrheinisch-Westfälischen Steinkohlengebietes" by permission of Springer-Verlag, New York, Inc.)

The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—

M. Iowa, by Matthew J. Avcin and Donald L. Koch
N. Missouri, by Thomas L. Thompson
O. Arkansas, by Boyd R. Haley, Ernest E. Glick, William M. Caplan, Drew F. Holbrook, and Charles G. Stone
P. Nebraska, by R. R. Burchett
Q. Kansas, by William J. Ebanks, Jr., Lawrence L. Brady, Philip H. Heckel, Howard G. O'Connor, George A. Sanderson, Ronald R. West, and Frank W. Wilson
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X. Montana, by Donald L. Smith and Ernest H. Gilmour
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Z. Arizona, by H. Wesley Peirce
AA. Idaho, by Betty Skipp, W. J. Sando, and W. E. Hall
BB. Nevada, by E. R. Larson and Ralph L. Langenheim, Jr., *with a section on* Paleontology, by Joseph Lintz, Jr.
CC. California, Oregon, and Washington, by Richard B. Saul, Oliver E. Bowen, Calvin H. Stevens, George C. Dunne, Richard G. Randall, Ronald W. Kistler, Warren J. Nokleberg, Jad A. D'Allura, Eldridge M. Moores, Rodney Watkins, Ewart M. Baldwin, Ernest H. Gilmour, and Wilbert R. Danner
DD. Alaska, by J. Thomas Dutro, Jr.

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UNITED STATES DEPARTMENT OF THE INTERIOR

CECIL D. ANDRUS, *Secretary*

GEOLOGICAL SURVEY

H. William Menard, *Director*

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FOREWORD

The year 1979 is not only the Centennial of the U.S. Geological Survey—it is also the year for the quadrennial meeting of the International Congress on Carboniferous Stratigraphy and Geology, which meets in the United States for its ninth session. This session is the first time that the major international congress, first organized in 1927, has met outside Europe. For this reason it is particularly appropriate that the Carboniferous Congress closely consider the Mississippian and Pennsylvanian Systems; American usage of these terms does not conform with the more traditional European usage of the term "Carboniferous."

In the spring of 1976, shortly after accepting the invitation to meet in the United States, the Permanent Committee for the Congress requested that a summary of American Carboniferous geology be prepared. The Geological Survey had already prepared Professional Paper 853, "Paleotectonic Investigations of the Pennsylvanian System in the United States," and was preparing Professional Paper 1010, "Paleotectonic Investigations of the Mississippian System in the United States." These major works emphasize geologic structures and draw heavily on subsurface data. The Permanent Committee also hoped for a report that would emphasize surface outcrops and provide more information on historical development, economic products, and other matters not considered in detail in Professional Papers 853 and 1010.

Because the U.S. Geological Survey did not possess all the information necessary to prepare such a work, the Chief Geologist turned to the Association of American State Geologists. An enthusiastic agreement was reached that those States in which Mississippian or Pennsylvanian rocks are exposed would provide the requested summaries; each State Geologist would be responsible for the preparation of the chapter on his State. In some States, the State Geologist himself became the sole author or wrote in conjunction with his colleagues; in others, the work was done by those in academic or commercial fields. A few State Geologists invited individuals within the U.S. Geological Survey to prepare the summaries for their States.

Although the authors followed guidelines closely, a diversity in outlook and approach may be found among these papers, for each has its own unique geographic view. In general, the papers conform to U.S. Geological Survey format. Most geologists have given measurements in metric units, following current practice; several authors, however, have used both metric and inch-pound measurements in indicating thickness of strata, isopach intervals, and similar data.

FOREWORD

This series of contributions differs from typical U.S. Geological Survey stratigraphic studies in that these manuscripts have not been examined by the Geologic Names Committee of the Survey. This committee is charged with insuring consistent usage of formational and other stratigraphic names in U.S. Geological Survey publications. Because the names in these papers on the Carboniferous are those used by the State agencies, it would have been inappropriate for the Geologic Names Committee to take any action.

The Geological Survey has had a long tradition of warm cooperation with the State geological agencies. Cooperative projects are well known and mutually appreciated. The Carboniferous Congress has provided yet another opportunity for State and Federal scientific cooperation. This series of reports has incorporated much new geologic information and for many years will aid man's wise utilization of the resources of the Earth.

H William Menard

H. William Menard
Director, U.S. Geological Survey

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Iowa

By MATTHEW J. AVCIN *and* DONALD L. KOCH

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1110-M

Prepared in cooperation with the
Iowa Geological Survey

Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Iowa

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1979

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—IOWA

By MATTHEW J. AVCIN[1] and DONALD L. KOCH[1]

ABSTRACT

Carboniferous rocks crop out in abundance from southeastern to north-central Iowa. Exposures are less numerous in south-central Iowa and are extremely limited in the west-central and southwestern parts of the State. Early studies were concentrated on Pennsylvanian strata because of the economic importance of their coal deposits. The regional stratigraphy of Pennsylvanian strata had been well summarized in publications of the late 19th and early 20th centuries.

Current interest in energy sources has led to a new interest in evaluation of the State's coal resources. A concerted effort is underway to define the stratigraphy of the Pennsylvanian in greater detail and to interpret the environments of deposition as they relate to the formation and occurrence of coal beds, particularly with respect to variations in thickness and quality of the coal deposits.

Generally, Mississippian strata have received less attention than those of the Pennsylvanian. Recent studies have resulted in recorrelation of the lower part of Mississippian beds with Devonian formations. Diagenetic effects on carbonate beds of north-central Iowa have resulted in repetitive sequences of limestones and dolostones that have led to confusion and miscorrelation of several member units.

Economic products from rocks of the Mississippian and Pennsylvanian Systems are limited to coal, road-constuction materials, brick and tile, soil-conditioner materials, and gypsum. Although nearly 50 oil-gas test wells have been drilled in Carboniferous strata, mostly within the deeper part of the Forest City basin, no economic deposits have been discovered to date.

INTRODUCTION

Carboniferous rocks underlie approximately 60 percent of Iowa's 143,209 km^2 (55,941 mi^2) and are the bedrock over most of this area (fig. 1). The combined maximum thickness recorded for the units assigned to the Mississippian (fig. 2) and Pennsylvanian (fig. 3) is 585 m (1,909 ft).

Because of resistant units, Mississippian rocks are well exposed along many of the streams that flow across the outcrop belt. These natural exposures are supplemented by many quarries in the Mississippian limestones. Regional dip is into the Western Interior (Forest City) basin from the outcrop belt that trends northwest from the southeast corner of Iowa. The outcrop belt arcs around to the west in the north-central part of the State where Cretaceous units overlap Mississippian strata.

Successively younger Pennsylvanian rocks are exposed downdip toward the deeper parts of the basin and are generally more poorly exposed than are Mississippian rocks. The Des Moines series is best exposed along the Des Moines River and in the many small surface mines that roughly parallel its course. Younger Pennsylvanian rocks are exposed in cutbanks along the smaller streams and in scattered quarries in the limestones. Because thick Pleistocene deposits mantle the erosion surface on top of the Carboniferous, exposures are very limited in upland areas away from the major drainageways.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Iowa Geological Survey.

LITHOSTRATIGRAPHY

MISSISSIPPIAN

Mississippian formations constitute the bedrock in a diagonal belt 32 to 64 km (20–40 mi) wide from Lee County in the southeast corner of the State northwestward to southeastern Kossuth County in north-central Iowa (fig. 1).

The standard sections of Mississippian rock units are found along the valley of the Mississippi River from southeastern Iowa into southern Illinois and southeastern Missouri. All formational units of the Kinderhook, Osage, and Meramec Series are represented in Iowa (figs. 2, 4). Because units of the Kinderhook vary lithologically from area to area, and because several of the lower units that originally

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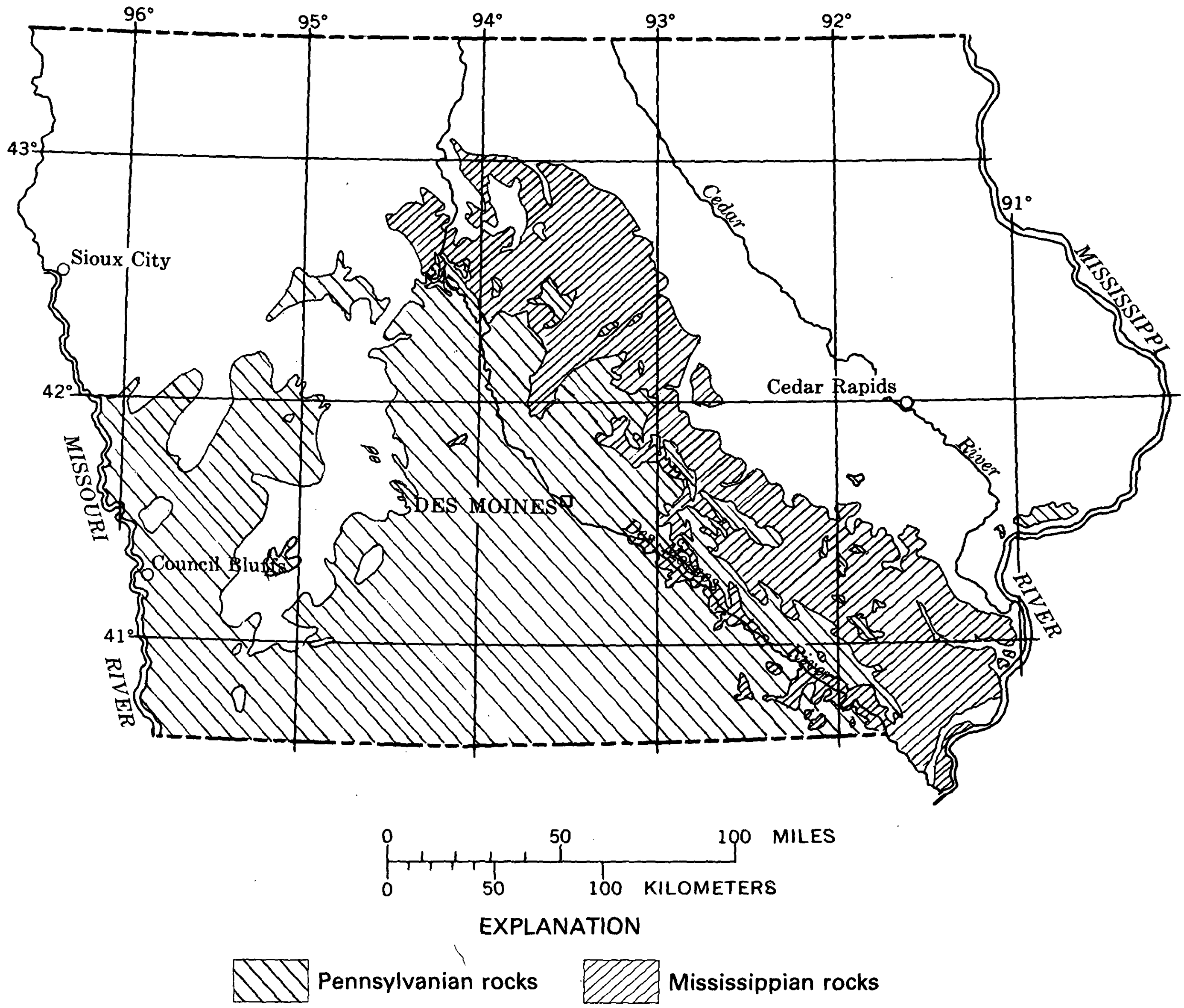

FIGURE 1.—Generalized outcrop pattern of Carboniferous rocks in Iowa.

were correlated as Mississippian are now placed in the Devonian, no standard section has been established for the Kinderhook. The type sections of the Burlington and Keokuk Formations are near those cities in southeastern Iowa. Exposures at Warsaw, Ill., nearly opposite Keokuk, Iowa, provide the type section for the Warsaw Formation. The Spergen, St. Louis, and Ste. Genevieve Formations are represented in Iowa by facies of formations farther south in the Mississippi Valley that are thicker and more uniform in lithology.

None of the Chester formations are represented in Iowa; the Chester seas probably never extended this far north.

The Mississippian-Devonian boundary is placed at the base of the North Hill Group by the Iowa Geological Survey. A discussion of this boundary in Iowa was presented by Dorheim, Koch, and Parker (1969).

Along the eastern and northeastern edge of the outcrop belt, Kinderhook strata are above formations of the Yellow Spring Group (uppermost Devonian). Along the northern boundary in Hancock and Kossuth Counties, Kinderhook strata are above the Lime Creek Formation (lower Upper Devonian). In western Iowa and eastern Nebraska, Kinderhook strata are above Middle Devonian rocks.

Series	Group	CENTRAL	SOUTHEAST
Meramec		Ste. Genevieve	Ste. Genevieve
		St. Louis	St. Louis Spergen
Osage		Warsaw	Warsaw
		Keokuk	Keokuk
		Burlington	Burlington Formation Cedar Fork Member Haight Creek Member Dolbee Creek Member
Kinderhook		Gilmore City Hampton Formation Iowa Falls Member Eagle City Member Maynes Creek Member	Hampton Formation Wassonville Member
	North Hill Group	Chapin Prospect Hill McCraney	Starrs Cave Prospect Hill McCraney

FIGURE 2.—Nomenclature of Mississippian units in Iowa.

For the most part, the boundary is a transgressive nonconformable contact.

The Mississippian System is unconformably overlain by Lower Pennsylvanian (Cherokee Group) sedimentary rocks throughout most of southeastern Iowa. Jurassic(?) (Fort Dodge Beds) and Lower Cretaceous units were deposited upon the eroded Mississippian surface in north-central Iowa. Pleistocene deposits mantle the Mississippian rocks along the outcrop belt. The following discussion of the Mississippian units of Iowa is based largely upon publications by Parker (1973) and Koch (1973).

KINDERHOOK SERIES

The Kinderhook Series in southeastern Iowa includes the North Hill Group and the superjacent Hampton Formation. A third unit, the Gilmore City Formation, is present above the Hampton in north-central Iowa. The Kinderhook is composed dominantly of carbonate rocks. Siltstone occurs in the North Hill Group and chert in the Hampton.

The North Hill Group (Workman and Gillette, 1956) includes in ascending order: McCraney Limestone, Prospect Hill Siltstone, and Starrs Cave Formation. An interval of dominantly oolitic limestone, the Chapin Formation of north-central Iowa, is equivalent to the Starrs Cave Formation of southeastern Iowa. The McCraney Limestone is a very pale orange to pale-yellowish-brown sublithographic limestone. In southeastern Iowa, it is characterized by the presence of brown medium-grained dolomite in irregular horizontally and vertically oriented planes. The basal part in southeastern Iowa consists of an oolitic crinoidal limestone, which locally contains a coarse brachiopod coquina composed almost entirely of *Chonetes*. The McCraney is present in north-central Iowa only along the eastern outcrop belt. The Prospect Hill Siltstone is a light-greenish-gray medium siltstone containing discontinuous green shaley seams in southeastern Iowa. In north-central Iowa, the formation varies from a dolomitic siltstone along the outcrop belt to a silty dolostone toward the west. The Prospect Hill contains fish teeth, brachiopods, and pelecypods. The Starrs Cave Formation is a very pale orange to pale-gray oolitic limestone, which contains fragmented crinoids, brachiopods, and corals. In north-central Iowa, the equivalent Chapin Formation is dominantly an oolitic limestone near the outcrop belt. Farther west, this interval is a dolostone, and it usually is inseparable from the superjacent Maynes Creek Member of the Hampton Formation.

The Wassonville Limestone Member is the only representative of the Hampton Formation (Laudon, 1931) in southeastern Iowa. The Wassonville consists of pale- to dark-yellowish-brown dolostone. Over most of the area of its occurrence, the Wassonville contains appreciable amounts of light-gray fossiliferous chert. In the extreme southeastern counties, the Wassonville is relatively chert free and is dominantly a dolomitic limestone.

In north-central Iowa, the Hampton Formation includes, in ascending order: Maynes Creek (equivalent to the Wassonville of southeast Iowa), Eagle City, and Iowa Falls Members. Contacts of these members are placed at a change in lithology from dolostone (Maynes Creek Member) to limestone (Eagle City Member) and from limestone to dolostone (Iowa Falls Member). The Maynes Creek is composed of dolostone and chert. Chert generally is present throughout the member near the outcrop belt. Farther west, chert generally is absent in the upper part of the member. At most localities, the upper contact is placed where chert is first observed. This result in apparent variations in thickness that often are extreme. In addition, the upper part is limestone at many localities and is then inseparable from the superjacent Eagle City Member. Where the upper contact is placed at the top of a brown dolostone, the unit is about 40 m (131 ft) thick and thins to about 17 m (56 ft) near the outcrop belt.

The Eagle City-Iowa Falls interval and the overlying Gilmore City Formation constitute the most variable carbonate sequence of the Kinderhook in north-central Iowa. The limestone interval above the Maynes Creek dolostone and beneath the Iowa

SYSTEM	SERIES	GROUP	FORMATION
Pennsylvanian	Virgil	Wabaunsee	French Creek
			Jim Creek
			Friedrich
			Grandharen
			Dry
			Dover
			Langdon
			Maple Hill
			Wamego
			Tarkio
			Willard
			Elmont
			Harveyville
			Reading
			Auburn
			Wakarusa
			Soldier Creek
			Burlingame
			Silver Lake
			Rulo
			Cedar Vale
			Happy Hollow
			White Cloud
			Howard
			Severy
		Shawnee	Topeka
			Calhoun
			Deer Creek
			Tecumseh
			Lecompton
			Kanwaka
			Oread
		Douglas	Lawrence
			Stranger
			Iatan
			Weston
	Missouri	Lansing	Stanton
			Vilas
			Plattsburg
		Kansas City	Bonner Springs
			Wyandotte
			Lane
			Iola
			Chanute
			Drum
			Quivira
			Westerville
			Cherryvale
			Dennis
			Galesburg
			Swope
			Ladore
			Hertha
		Pleasanton	undifferentiated
	Des Moines	Marmaton	Lenapah
			Nowata
			Altamont
			Bandera
			Pawnee
			Labette
			Fort Scott
		Cherokee	undifferentiated

FIGURE 3.—Nomenclature of Pennsylvanian units in Iowa.

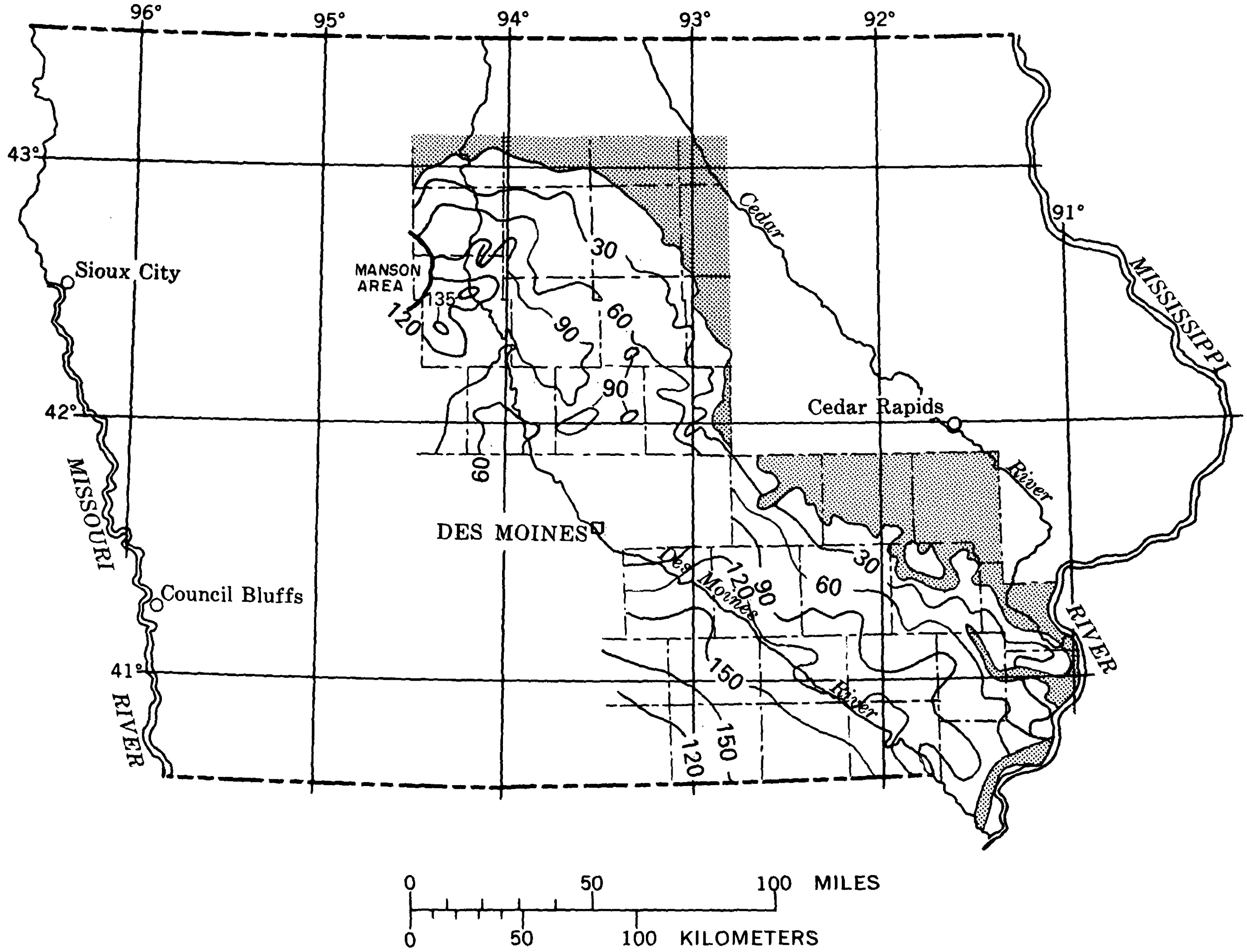

FIGURE 4.—Thickness, in meters, of Mississippian units within and peripheral to the outcrop belt.

Falls dolostone is designated as Eagle City. Along the outcrop belt, the Eagle City generally is an oolitic limestone in the upper part and a micrite in the lower part. It grades westward to micrite and dolomitic micrite containing beds of dolostone. In the subsurface section, the thickness of the Eagle City ranges from about 3.3 m (11 ft) to 17 m (56 ft). This variation in thickness is the result of differences in the vertical extent of dolomitization from one location to another.

The Iowa Falls is a dolostone interval above the Eagle City and beneath the Gilmore City Limestone; apparently, it is a dolomitized lateral equivalent of the Gilmore City as suggested by Thomas (1960). Thicknesses of as little as 3.3 m (11 ft) and as much as 20 m (66 ft) have been interpreted in the subsurface section. Again, the variations in thickness are related principally to differences in the vertical extent of dolomitization. At some locations, no interval of dolostone is present that is correlative with the Iowa Falls.

A composite thickness of 68 m (223 ft) was indicated for the Maynes Creek-Eagle City-Iowa Falls interval by Laudon (1931). As no exposure of the entire sequence was available, composite thicknesses for each unit were obtained by correlating fossil zones and lithologic types from one partial exposure to another presumed to be higher or lower in the section. Multiple facies changes from limestone to dolostone both laterally and vertically generally were unrecognized. Indeed, the contacts as they are identified today usually are gradational contacts from limestone to dolostone or from dolostone to limestone. Drill cuttings show an average of only 37 m (121 ft) for the entire thickness of the Maynes Creek-Eagle City-Iowa Falls interval near the outcrop belt. Toward the west, the average thickness increases to 58 m (190 ft).

The Gilmore City Limestone is conformable above the Hampton formation, and, like the Eagle City, it is composed dominantly of oolitic limestone near the outcrop belt, but toward the west it contains alternating beds of oolitic limestone and biofragmental (crinoidal) limestone. The lower contact generally is gradational with the Iowa Falls. However, where the Iowa Falls dolostone is "absent" it is extremely difficult to identify a lower boundary, and the boundary selected usually is arbitrary. The upper contact is unconformable at different locations with younger Mississippian units or with Pennsylvanian, Cretaceous, or Pleistocene units. Consequently, the thickness of the Gilmore City is extremely variable. Where Osage units are superjacent, the thickness of the Gilmore City is about 45 m (148 ft).

The above discussion of the Hampton and Gilmore City Formations of north-central Iowa clearly shows that the veracity of current nomenclature is questionable. In summary, the following conditions result in a complexity of geologic factors that obfuscate analysis of the stratigraphy of this part of the Mississippian section:

1. Various units have been partly or completely dolomitized, and there is no uniformity, either horizontal or vertical, in the extent of dolomitization.
2. Alternating units of limestone and dolostone are the principal basis upon which formations and members have been identified. Many subsurface sections show an unusually thick formation or a "missing" member where dolomitization has been more extensive.
3. Facies changes from oolitic limestone to biofragmental limestone and micrite within members and formations further complicate correlations.
4. Most of the natural exposures are limited to the eastern outcrop belt. At no exposure is a complete sequence observed.
5. Where complete sequences are present, they are covered by Pennsylvanian, Cretaceous, or Pleistocene strata. Thus, subsurface data must be relied upon for stratigraphic control.

Hughes (1977) recognized that the greatest thickenings of units of the Hampton Formation and of the Gilmore City Formation are mutually exclusive, which further demonstrates that formations and members are diagenetic and (or) depositional facies of each other. The recognized stratigraphic discrepancies can be resolved only through applied petrographic studies such as those of Hughes (1977) and Lloyd (1973).

OSAGE SERIES

The Osage Series in Iowa includes the Burlington Limestone (Hall, 1857), Keokuk Limestone (Owen, 1852), and Warsaw Formation (Hall, 1857). The Osage is composed dominantly of cherty limestone and dolostone containing minor beds of shale. The detailed stratigraphy of and conformable relationships between the Burlington and Keokuk and the Keokuk and Warsaw in southeastern Iowa were described by Harris and Parker (1964).

The members of the Burlington Limestone are, in ascending order: Dolbee Creek, Haight Creek, and Cedar Fork. In general, the Dolbee Creek Member (Laudon's *Cactocrinus* zone in the type area, but including his underlying zones where the member is thick) is a coarsely crystalline crinoidal limestone with a small amount of chert. *Spirifer grimesi* and fragments of bryozoans are common. This unit is thickest (5.7 m; 19 ft) in extreme southeastern Iowa and is absent in central and north-central Iowa. The Haight Creek Member (Laudon's *Physetocrinus* zone) is highly cherty and contains both dolostone and limestone beds. The base of this unit is marked by a dolostone that contains greenish-black glauconite grains or pellets. In southeastern Iowa, the Haight Creek has a rather uniform thickness of about 15 m (49 ft). Towards the north and west, this member is superjacent to Kinderhook beds, and where glauconite is absent, it is difficult to discriminate between the biofragmental limestone of the lower Burlington and the similar lithology of the subjacent Gilmore City. The Cedar Fork (Laudon's *Dizygocrinus* and *Pentremites* zones) is a coarsely crystalline crinoidal limestone similar to the Dolbee Creek. However, the *Dizygocrinus* zone commonly contains abundant soft flakes of glauconite, which gives the limestone a green color, and the *Pentremites* zone is cherty. In southeast Iowa, the Cedar Fork ranges from 3.3 to 10 m (11 to 33 ft) in thickness. In north-central Iowa, where dolomite content increases and both the carbonate and chert become darker, the member generally is not differentiated.

The Keokuk Limestone in southeastern Iowa consists of cherty crinoidal carbonate beds in the lower part ("Montrose Cherts" of Keyes, 1894) and fossiliferous cherty carbonate beds and interbedded fossiliferous shale in the upper part. The "Montrose

Cherts" are easily distinguished from the relatively noncherty Cedar Fork Member of the Burlington Limestone. Also, chert in the Keokuk usually contains abundant spicular fossil fragments that serve to distinguish the Keokuk from the dominantly crinoidal chert of the Burlington. Shale beds in the Keokuk increase in number and thickness upward, grading into the gray Warsaw shale, and dolostone beds replace the limestone. In subsurface usage, the upper boundary of the Keokuk generally is placed at the top of the uppermost carbonate bed. In extreme southeastern counties, the Keokuk averages about 25 m (82 ft) in thickness. In north-central Iowa, the Keokuk usually is undifferentiated from the Warsaw.

The Warsaw Formation is characterized by shale and argillaceous dolostone. Geodes that principally contain quartz are common in the southeastern outcrop belt. Chalcedonic chert and crystalline quartz, dominantly of geode origin, are particularly abundant in the subsurface toward the west-northwest where argillaceous dolostone is dominant over shale. The average thickness for the Warsaw is 17 m (56 ft) in southeastern Iowa. In north-central Iowa, the maximum thickness of the Keokuk-Warsaw interval is approximately 25 m (82 ft). The Warsaw is unconformably overlain by the Spergen and St. Louis Formations.

MERAMEC SERIES

The Meramec Series contains the youngest Mississippian rocks in Iowa, in ascending order: Spergen (Salem) Limestone (Ulrich, 1905), St. Louis Limestone (Engelmann, 1847), and Ste. Genevieve Limestone (Shumard, 1860). The rocks of the Meramec units are more varied than those of the Kinderhook and Osage. In southeastern Iowa, the Meramec overlies older Mississippian strata unconformably and overlaps formations as old as the North Hill Group (Kinderhook). The upper boundary constitutes a post-Mississippian erosional surface that is unconformably overlain by the Pennsylvanian Cherokee Group and by Pleistocene deposits in southeastern Iowa. Post-Mississippian—pre-Pennsylvanian erosion removed Meramec strata over much of north-central Iowa, and pre-Cretaceous and Pleistocene erosion further reduced the area of occurrence of these units. Only the St. Louis Limestone and the Ste. Genevieve Limestone are recognized in north-central Iowa.

The Spergen (Salem) Limestone is a brownish-gray micaceous dolomitic limestone that has a minor quartz sand fraction. It is restricted to southern Iowa by the pre-St. Louis unconformity. The Spergen often has been confused with the basal St. Louis. The average thickness is 6.6 m (22 ft).

The St. Louis Limestone in southeastern Iowa is dominantly a micritic limestone; it has a lower zone of brown arenaceous dolostone that in the subsurface contains a high percentage of anhydrite and gypsum. The high degree of brecciation of the carbonate rocks in the outcrop section probably is the result of dissolution of the evaporites and subsequent collapse. The average thickness is 17 m (56 ft), although thicknesses of 41.6 m (136 ft) are recorded within the evaporite areas. In north-central Iowa, sandstone generally is dominant in the lower part of the formation; the maximum recorded thickness is 18 m (59 ft).

The Ste. Genevieve in southern Iowa is composed of fine-grained limestone that has a high percentage of quartz sand. In central Iowa, the unit is represented mainly by green and red shale and thin beds of fine-grained limestone. The contact relationships with the subjacent St. Louis are unclear, and both the thickness and distribution of the unit vary greatly because of post-Mississippian erosion.

DEPOSITIONAL ENVIRONMENTS OF MISSISSIPPIAN ROCKS

The beginning of Mississippian time was a period of general emergence; no evidence of localized uplift has been found. The region was a broad plain, modified somewhat by differential erosion. Shallow restricted seas are evidenced by the distribution of the formations of the North Hill Group. Depositional regimes for the Hampton and Gilmore City include offshore marine, unrestricted shoal, restricted shoal, restricted nearshore, and tidal environments. The nearly pure biofragmental carbonate rocks of the Osage were deposited in a shallow shelf environment. The shales of the Warsaw are the termination of a clastic wedge from the east. Carbonate deposition continued into the lower Meramec. However, an evaporite facies is present in the lower St. Louis. The normal marine carbonate rocks of the upper St. Louis and the Ste. Genevieve contain an increase in sand and shale northward, culminating in a shale facies of the Ste. Genevieve in north-central Iowa. No evidence of Chester deposition has been recognized in Iowa. At the end of Mississippian time, this region was a landmass of low relief undergoing erosion and karst formation. The Mississippian surface was further modified by uplift, which continued into Pennsylvanian time.

NATURE OF MISSISSIPPIAN-PENNSYLVANIAN CONTACT

The Cherokee Group unconformably overlies units of the Meramec and Osage Series throughout most of its areal extent. However, Pennsylvanian outliers overlie progressively older Mississippian, Devonian, and Silurian units to the northeast. In some outcrops and cores, a regolith is present on top of the Mississippian. A typical exposure of this interval consists of 1.5–2.4 m (5–8 ft) of St. Louis in a quarry north of Keosauqua (SE¼ sec. 36, T. 70 N., R. 10 W., Van Buren County). The heavily iron stained irregular upper surface of the St. Louis has as much as 0.6 m (2 ft) of relief. This surface is overlain by a green clay as much as 2.4 m (8 ft) thick which contains corroded limestone "clasts" in the lower part. The limestone "clasts" decrease in size and disappear about midway up, and the clay becomes noncalcareous slightly farther up. Next is an interval of approximately 6 m (20 ft) of dark-gray silty shale of Pennsylvanian age, which in turn is overlain by about 45 cm (18 in.) of highly pyritic coal. The contact between the dark-gray shale and the green clay is sharp and occasionally shows erosional reworking of the green clay into the superjacent dark-gray shale. Where this regolith attains a thickness of 4.5–7.5 m (15–25 ft) in the subsurface, the central part of the clay is usually dark red.

Although the contact relationships definitely indicate that the surface of the Mississippian was eroded, its present configuration, which is a composite of original topography and post-Pennsylvanian deformation, gives no hint of its configuration at the beginning of Pennsylvanian deposition. Preliminary data suggest that the Mississippian surface was slightly undulatory and of low relief, with locally prominent positive features.

PENNSYLVANIAN

Rocks of Pennsylvanian age occupy an area of 51,000 km^2 (20,000 mi^2) in the southwestern third of Iowa, and they form the bedrock over a significant part of this area. The Pennsylvanian System in Iowa is divided, in ascending order, into the Des Moines, Missouri, and Virgil Series (fig. 3).

DES MOINES SERIES

Rocks of the Des Moines Series crop out in a broad northwest-trending band on both sides of the Des Moines River and form scattered outliers along the northeast margin of this band. Exposures along the southern part of the Des Moines River serve as the type area for the Des Moines Series. The maximum reported thickness for this series is approximately 275 m (900 ft), in the subsurface of southwestern Iowa.

The Des Moines Series is divided into two groups, the Cherokee and the overlying Marmaton. The Cherokee Group unconformably overlies Mississippian rocks of the Osage and Meramec Series and is dominated by a sequence of dark shale and siltstone. Sandstone, lighter colored shale, coal, limestone, and black phosphatic shale, in decreasing order of abundance, complete the suite of lithologies represented in the Cherokee Group. Although the Cherokee reaches a maximum thickness of approximately 230 m (750 ft) in southwest Iowa, no formational subdivisions are differentiated. However, informal names have been applied to some persistent limestones and coals in the upper part of the group, particularly in south-central and southeastern Iowa.

Recent work in southeastern Iowa indicates that a significant part of the lower Cherokee Group may be Atokan in age, and an outlier exposed near Davenport on the Mississippi River is even older. No attempt has been made at this time to redefine the boundaries of the Cherokee Group either lithostratigraphically or chronostratigraphically.

The 45-m (145-ft)-thick Marmonton Group in Iowa has been divided into 7 formations; 14 members are recognized. Although shale and sandstone are dominant, persistent limestones account for eight of the named members and permit subdivision of the Marmaton. Several coals are recognized and have been named. However, only the Mystic Coal has significant reserves.

MISSOURI SERIES

Rocks of the Missouri Series are divided into three groups, in ascending order, the Pleasanton, Kansas City, and Lansing. The series is approximately 93 m (305 ft) thick.

The Pleasanton Group is poorly understood in Iowa; it is defined as the interval between the base of the overlying Hertha Limestone and the unconformity at the top of the subjacent Des Moines Series. Over much of the State, the unconformity is not apparent, but Cline (1941, p. 70) placed the unconformity at the base of the Chariton Conglomerate in Appanoose County. The thickness of the Pleasanton Group is variable but is believed to reach a maximum of approximately 12 m (40 ft) in the subsurface. Shale and minor sandstone are dominant. The interval also contains several thin limestones, one of which is named, and one named coal.

The Kansas City Group is the thickest of the three groups of the Missouri Series at approximately 66 m (215 ft) and conformably overlies the Pleasanton Group. Fourteen formations have been defined in the alternating sequence of shales and limestones. Similarly, four of the limestone and two of the shale formations have been divided into alternating limestone and shale members. The shale formations range in thickness from 1.5 to 4.5 m (5 to 14.5 ft). The limestone formations range in thickness from 0.6 to 8.5 m (2 to 27.5 ft).

The conformable superjacent Lansing Group continues the alternating series of limestone and shale found in the Kansas City Group. The two limestone formations and one shale formation have a maximum total thickness of 15 m (50 ft).

VIRGIL SERIES

The rocks of the Virgil Series are divided into three groups, in ascending order, the Douglas, Shawnee, and Wabaunsee. The maximum thickness for the series is about 150 m (500 ft).

The Douglas Group in Iowa is a highly variable unit consisting of 5.5–25 m (18–80 ft) of shale, siltstone, and locally limestone. Although four formations are officially recognized in Iowa, these units are very difficult to recognize, except locally. Hershey and others (1960) included the Pedee Group with the Douglas and indicated that the basal unconformity cannot be found.

The Shawnee Group continues the alternating sequence of limestone and shale shown by the Kansas City and Lansing Groups. Seven formations are recognized, which have a total maximum thickness of 55 m (180 ft). The unit is conformable with the underlying Douglas Group. All but one of the formations, the Calhoun, have been divided into members, on the basis of alternating limestone and shale. Thirty-six members have been recognized; the limestone formations contain at least seven members, and the shale formations, three members. The shale formations range in thickness from 1 to 7 m (3 to 22 ft), and the limestone formations, from 3.5 to 11 m (11 to 35 ft).

The Wabaunsee Group has a maximum thickness of about 65 m (210 ft) and conformably overlies the Shawnee Group. Although this group is divided into 25 formations on the basis of an alternating sequence of limestone and shale, shale is dominant. Secondary lithologies, in order of abundance, are siltstone, sandstone, and coal; no members are recognized, but the coals have been informally named. Limestone formations are relatively thin, having an average thickness of 1 m (3 ft) and a range of 0.3–3.5 m (1–12 ft) in thickness. Thicknesses of shale formations range from 1 to 30 m (3 to 100 ft). The group thins toward the northeast.

The top of the Virgil Series coincides with the beginning of an extensive break in the depositional record in Iowa. An angular unconformity marks the contact between the Pennsylvanian and overlying units. Rocks of questionable Jurassic age overlie the Pennsylvanian in several areas in the vicinity of Fort Dodge. Although this thin sequence of red and green shale covers only about 90 to 130 km^2 (35–50 mi^2), it is important because of the industry based on the associated gypsum deposits. The remainder of the Pennsylvanian is overlain either by Cretaceous or Quaternary clastic deposits. The Cretaceous discontinuously overlies the Pennsylvanian in the western third of the State and is in turn overlain by the Quaternary, which overlies the Pennsylvanian wherever the Cretaceous is absent.

FACIES AND DEPOSITIONAL ENVIRONMENTS OF PENNSYLVANIAN ROCKS

The Pennsylvanian section in Iowa is less well understood than are rocks of equivalent age in nearby States, particularly Kansas and Illinois. This lack of understanding results from a paucity of exposures due to a thick Quaternary cover, from local structure, from rapid lateral and vertical facies changes, and from the limited number of geophysical logs available from drill holes that penetrate the entire Pennsylvanian sequence. Currently, the Iowa Geological Survey is attacking part of the problem by means of a coring program in the Des Moines Series to evaluate the coal resources of the State. The data obtained from this program have served as the basis for the Des Moines Series part of this discussion. Similarly, an earlier study of highway construction materials by Hershey and others (1960) provided the data for the discussion of the Missouri and Virgil Series.

In the Cherokee Group, the rapid facies changes result from the interaction of several factors, including original topography, an overall transgression that persisted through the Des Moines Series, and repeated widespread eustatic fluctuations of sea level. In the southeastern part of the State, the oldest Pennsylvanian sedimentary rocks are terrestrial to marginal marine and are dominated by dark-gray shale, siltstone of various shades of gray, small-scale channel sandstone, and interbedded siltstone and sandstone. Successive cyclic units indicate a gradually increasing marine influence, as shown by the progressive appearance of *Lingula*, nodular septarian limestones, light-gray to gray-green shale containing calcareous marine fossils, bedded marine fossiliferous limestone, and phosphatic black shale. Concomitantly, the dark-gray shale and darker siltstone decrease in dominance and eventually disappear, and the size of the channel sands increase. Throughout the Cherokee, clastic sediments were transported into the basin from two directions. In the southeastern part of the basin, transport was from the northeast, whereas in the northern part of the basin, transport was from the north.

The first appearance of the various lithologies was not synchronous over the entire area but varied locally depending on original topography of the site of deposition and the proximity of the shoreline. The marine influence was felt earlier in the west and southwest than in the southeast. Similarly, the Cherokee Group thickens to the west, the most dramatic thickening being in the lowermost part of the unit. These lowermost beds may be older in the west; however, at present, their age is undetermined.

In the upper part of the Cherokee Group, original topography no longer exerted much influence because of the filling of topographic lows on the original surface. Contemporaneously, the overall transgression had progressed far enough so that laterally persistent marine limestone and phosphatic black shale began to form, and the sequence of lithologies began to resemble the cyclothems of the Eastern Interior Basin. A good discussion of the Eastern Interior Basin cyclothems may be found in Hopkins and Simon (1975, p. 173). Like the cyclothems of the Eastern Interior Basin, the upper Cherokee cyclothems of Iowa rarely contain all the lithologies attributed to the idealized sequence. The one outstanding difference between the upper Cherokee Group cyclothems and those of the Eastern Interior Basin, is that the units in Iowa, particularly the coals, generally are much thinner.

The general trend toward open marine conditions continued upward into the Marmaton Group. Because of the surface of low relief that existed during deposition of the Marmaton, the transgressive-regressive cycles produced by eustatic sea-level changes formed thin but laterally continuous shallow-water marine deposits, particularly limestone and shale. Minor deeper water sedimentation is indicated by the presence of phosphatic, slaty black shale, according to Heckel (1977). The sandstone, thick siltstone, and nonfossiliferous shale represent prodeltaic and lower delta-plain deposits. Although the Marmaton is predominantly clastic, insufficient data are available to define the direction of transport, but a northeasterly source is currently favored. Only one coal, the Mystic, reaches a significant thickness—75 cm (2.5 ft)—in the Marmaton Group, despite the fact that subaerial conditions existed several times; the thinness of the coal beds perhaps indicates insufficient time or an unfavorable environment for the formation of peat.

During deposition of the Missouri Series, marine conditions predominated, the characteristics of each cycle being determined by the water depth during the transgressive phase and the position of the shoreline at maximum regression. The Pleasanton Group is poorly defined in Iowa. It includes shale and siltstone and some sandstone, and it contains the Exline Limestone and the Ovid coal. These units indicate a minor transgression during which the incoming waters remained relatively shallow, followed by a regressive phase during which the shoreline migrated through Iowa and the Ovid coal was deposited.

In the overlying Kansas City and Lansing Groups, limestone and marine shale are dominant, and several of the cyclothems (such as the Swope Limestone) show characteristics of the "Kansas cyclothem" as described by Heckel (1977). However, not all the transgressions reached sufficient depth to produce phosphatic black shale, and there appears to have been an irregular upward trend of decreasing water depth during the transgressive phase. Concomitantly, an increase in the deposition of shallow-water clastic materials took place, and subaerial conditions prevailed at least once, in the Chanute Shale.

This trend toward decreasing water depth continued into the base of the Virgil Series, where the Douglas Group cannot be divided because of the suppression of limestone which elsewhere serves as a basis for recognition of members. The Douglas Group, like the Pleasanton Group, represents a minor eustatic event, during which the water remained shallow, and regressive prodeltaic shale was dominant.

The Shawnee Group reflects the earlier pattern of the Kansas City Group and is a dominantly marine section containing widespread thick limestone beds and several phosphatic black shale beds. The black shales indicate several deep-water transgressive events separated by shallower water transgressive and regressive but generally marine deposits.

Rocks of the Wabaunsee Group record a return toward shallowing water, possibly as a result of basin filling. The Severy Shale, which is the lowermost formation of the Wabaunsee Group, has a phosphatic black shale indicating deep marine conditions, but the black shale almost immediately overlies a coal, which illustrates a reversion to an upper Cherokee depositional pattern. Above this sequence there is a marked increase in nearshore sediments, particularly prodeltaic shale, and a general decrease in limestone deposits.

With the exception of a series of thin isolated units exposed near Fort Dodge, which have been tentatively assigned to the Jurassic, no record of deposition exists from the top of the Virgil Series to the base of the Dakota Group of the Cretaceous.

ECONOMIC PRODUCTS

PRODUCTS FROM MISSISSIPPIAN ROCKS

Mississippian rocks are a significant source of industrial minerals in Iowa. About 25 percent of the stone used for road construction and maintenance is derived from Mississippian strata, principally the Hampton, Burlington, and Gilmore City Formations, in 23 counties. A substantial amount of the stone production is for agricultural lime. Although most of the stone is from surface quarries, six underground mines are in operation.

Gypsum is mined from the lower St. Louis at a single location in south-central Iowa. The gypsum is used as a retarder by cement plants in Des Moines, and some of it is pelletized (together with finely crushed limestone and bentonite) for use as a soil conditioner.

PRODUCTS FROM PENNSYLVANIAN ROCKS

Coal mining started in several areas of Iowa in the early 1840's (figs. 5, 6). The coal was used locally for home heating and for smithing operations. Production increased steadily as coal replaced wood fuel both for the home and the Des Moines

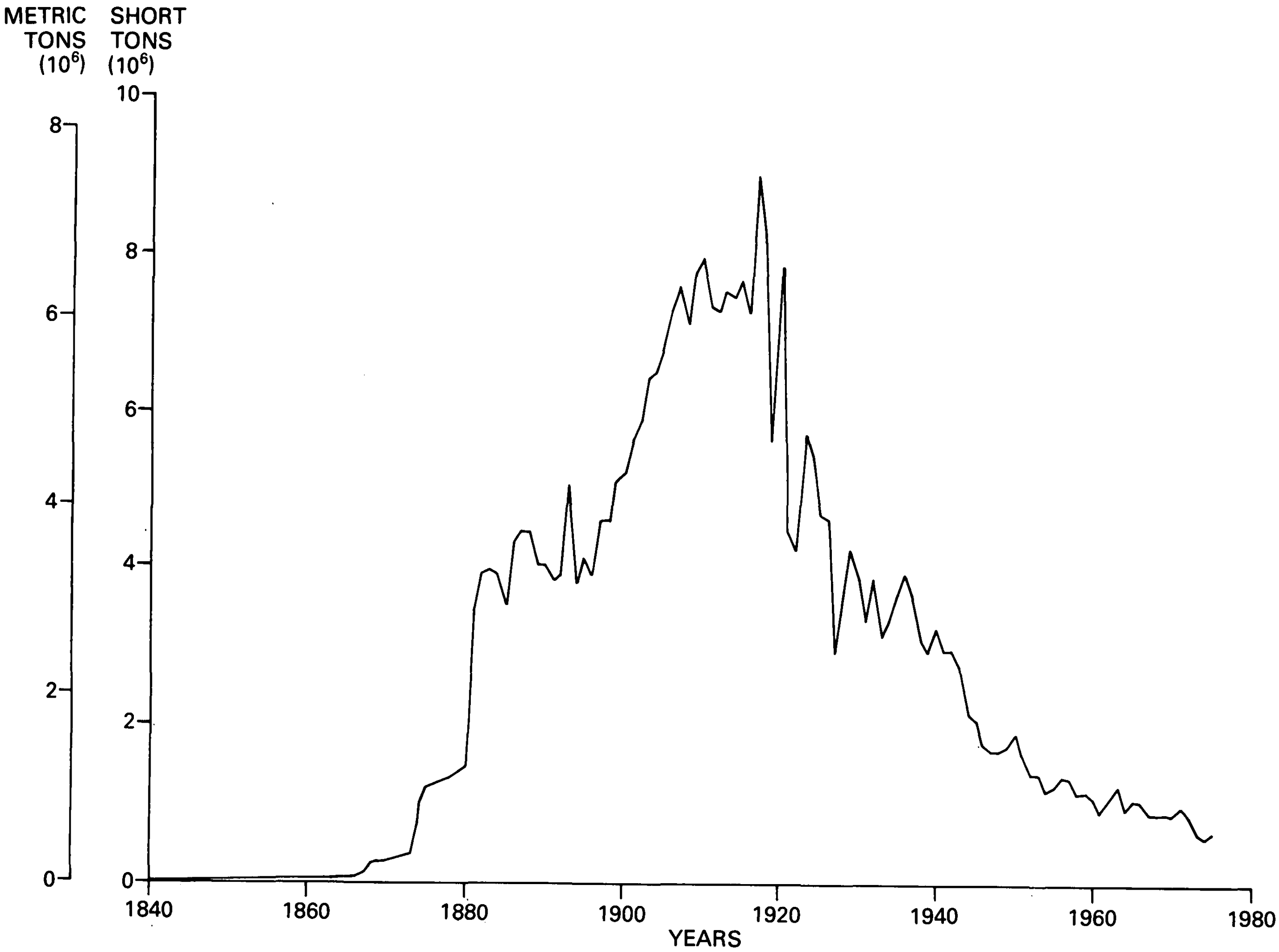

FIGURE 5.—Yearly coal production in Iowa.

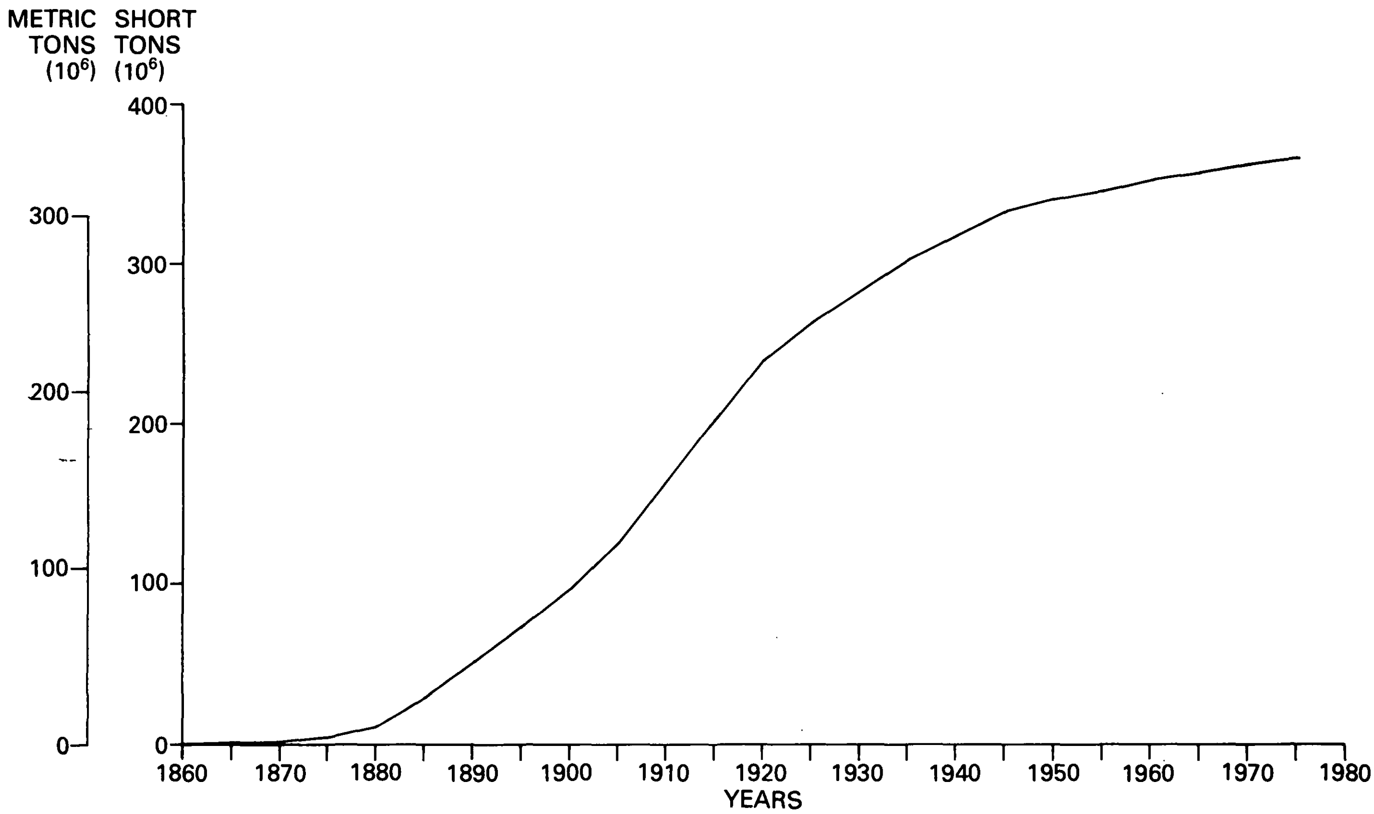

FIGURE 6.—Cumulative coal production in Iowa.

riverboats, but it did not top 100,000 tons per year until the late 1860's. About this time, the coal industry in Iowa began to accelerate because of the coming of the railroads, and it continued to expand until the early 1900's. The State Mining Inspectors office was established in 1880 with the responsibility of collecting production data. The unprecedented increase in coal production between 1880 and 1881 probably is related more to inaccurate prior records than to an actual increase in tonnage.

By the early 1900's, the industry was beginning to peak, although it had its best year in 1917 in response to war conditions. In 1900, the population of Iowa had reached 2,232,000; it grew only moderately thereafter to the present level of approximately 2,870,000. Agriculture has been the dominant economic endeavor; industrial development slowed when it became obvious that the State had insufficient natural resources to support large industry, removed as it was from the main markets to the east. These trends created a ceiling on the development of Iowa coal, and a long-term decline was assured by events outside the State.

The principal pressure on the industry was competition for its traditional markets from out-of-state coal, which was cheaper, either because of easier mining conditions or better applied technology combined with economies resulting from larger operations and thicker coal beds. In addition, as oil and gas became readily available they began to compete with coal. Without a large industrial base to serve as a safe market, the industry slowly slid back to production levels comparable with those of the 1860's and early 1870's. The industry today supplies small amounts of coal on short-term contracts and the spot market to local utilities.

In spite of its inherent disadvantages, namely, high sulfur content and high mining cost, Iowa coal production should be rejuvenated in the near future. This prediction is based on the large remaining reserves in the State, rising mining costs elsewhere as easily mined coal is exhausted, and increasing shipping costs attendant on moving large volumes of coal and the necessary upgrading of roadbeds. Therefore, the cumulative production curve should show a steepening; the only remaining question is when?

In addition to coal, Pennsylvanian strata yield limestone and clay. Limestone is produced in 13 counties in southwestern Iowa, primarily as crushed stone. Most of the production is used as bituminous and concrete aggregate, roadstone, and road base, although significant quantities are utilized as agricultural limestone and in the manufacture of cement.

Clay is utilized in the manufacture of tile, brick, and cement. Accurate production information for either clay or limestone cannot be obtained because in many of the counties the industries are so small that publication of production figures would result in the release of information considered proprietary by the industry.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Missouri

By THOMAS L. THOMPSON

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1110-N

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Division of Geology and Land Survey

Historical review and summary of
areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in Missouri

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—MISSOURI

By THOMAS L. THOMPSON[1]

ABSTRACT

Carboniferous rocks crop out over approximately two-thirds of the State of Missouri, although much of this region is partially covered by either Pleistocene drift and loess (northern part) or residual material from deeply weathered bedrock. Mississippian (lower Carboniferous) strata are predominantly limestone, although some shale is present in eastern and southwestern Missouri. Mississippian limestone is an economically important resource to the State and has been the host rock for lead and zinc. Pennsylvanian (upper Carboniferous) rocks consist of thin beds of limestone, shale, and sandstone; limestone and shale beds, although thin, are very widespread laterally persistent units. Coal is an increasingly important resource associated with Pennsylvanian strata, and many of the limestone beds are thick enough to be quarried in some regions of the State. In restricted parts of central Missouri, Pennsylvanian clay has been mined for its refractory properties.

INTRODUCTION

Rocks of Carboniferous age in North America have been divided into two systems, the Mississippian and Pennsylvanian. These systems were proposed in a review of Carboniferous strata of North America by Williams (1891), who described the "Mississippian series" of the upper Mississippi River valley (cited by Williams, 1891, p. 136, to have been originally named the "Mississippi limestone series, or Mississippi group" by Winchell 1872), and the "coal measures or Pennsylvanian series" of the Appalachian province of Pennsylvania. Although not immediately adopted by all North American Carboniferous stratigraphers, these names received general acceptance soon after the turn of the century. By the period 1935–1940, they seem to have been accepted by most, if not all, North American Carboniferous stratigraphers, and the general correlation of Pennsylvanian for upper Carboniferous and Mississippian for lower Carboniferous became common usage. Missouri contains at least part of the type areas for all four of the formally recognized series in the Mississippian System (fig. 1) and includes the type area for one of the five recognized midcontinent Pennsylvanian series.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Missouri Department of Natural Resources, Division of Geology and Land Survey.

GEOLOGIC SETTING

The dominant factor in the present distribution of Paleozoic strata in Missouri is the Ozark "dome" (or "uplift"). Rocks dip away from the region of maximum uplift (in east-central Missouri, the St. Francois Mountains) in all directions, and the amount of strata eroded decreases away from the center. Uplift and erosion before Carboniferous deposition (but apparently after Late Devonian deposition) produced an arcuate distribution of pre-Carboniferous formations. Basal Carboniferous rocks rest on strata ranging from Early Ordovician to Late Devonian in age; strata are progressively younger away from the uplift center.

The basal contact of the Carboniferous is unconformable, with the possible exception of that in northeastern Missouri, where the uppermost late Upper Devonian Louisiana Limestone underlies basal Carboniferous strata (Hannibal Shale) with little or no unconformity. The hiatus at the basal Carboniferous contact increases rapidly westward, however, as the Louisiana Limestone disappears and the Hannibal thins markedly. The region of Devonian-Carboniferous transition is restricted to that along the Mississippi River valley in Marion, Pike, and Ralls Counties, Mo.

Except for a sandstone (Indian Cave Sandstone) in northwestern Missouri of questionable Permian

[1] Missouri Department of Natural Resources, Division of Geology and Land Survey, Rolla, MO 65401.

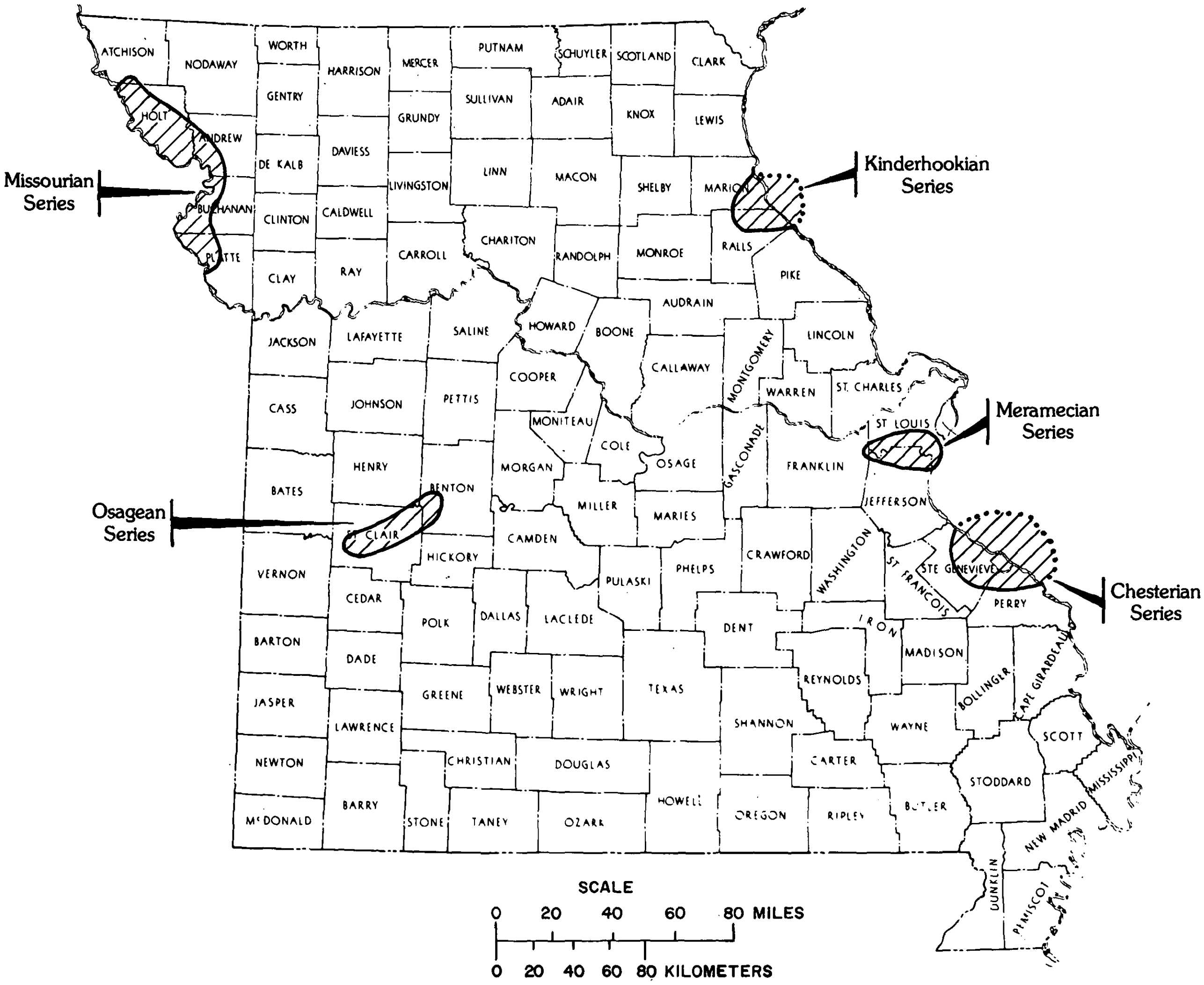


FIGURE 1.—Type areas of Mississippian and Pennsylvanian Series.

age, upper Carboniferous rocks constitute the youngest consolidated strata in Missouri except for the Cretaceous embayment region of southeastern Missouri. Thus, sediments overlying the Carboniferous consist primarily of Pleistocene glacial and other deposits north of the Missouri River and residual products derived by in situ weathering of bedrock south of the Missouri River.

Structural events were relatively minor throughout early Carboniferous time until late Meramecian and early Chesterian deposition. Increased clastic content in eastern Missouri and removal of Meramecian and upper Osagean strata by erosion before initial deposition of Chesterian strata (Hindsville Limestone) in western Missouri indicate late Mississippian uplift and associated activity. After Mississippian deposition, uplift and erosion removed considerable amounts of Mississippian strata before the start of Pennsylvanian deposition. Chesterian strata have been preserved in downdropped fault blocks in southeastern and southwestern Missouri. Elsewhere, erosion removed strata down to middle Mississippian and, in the Ozark uplift region south of the Missouri River, to Lower Ordovician strata.

Most of the Ozark "mountains" lie inside the belt of Mississippian outcrop. In western Missouri, the Mississippian marks the location of a broad, highly dissected upland (Springfield Plateau); the sharp eastern margin (often termed the "Eureka Springs escarpment") of this plateau forms a distinctive topographic feature above the Ordovician outcrops to the east. This plateau is less distinct north of Springfield because it has been highly dissected by tributaries of the Missouri River system.

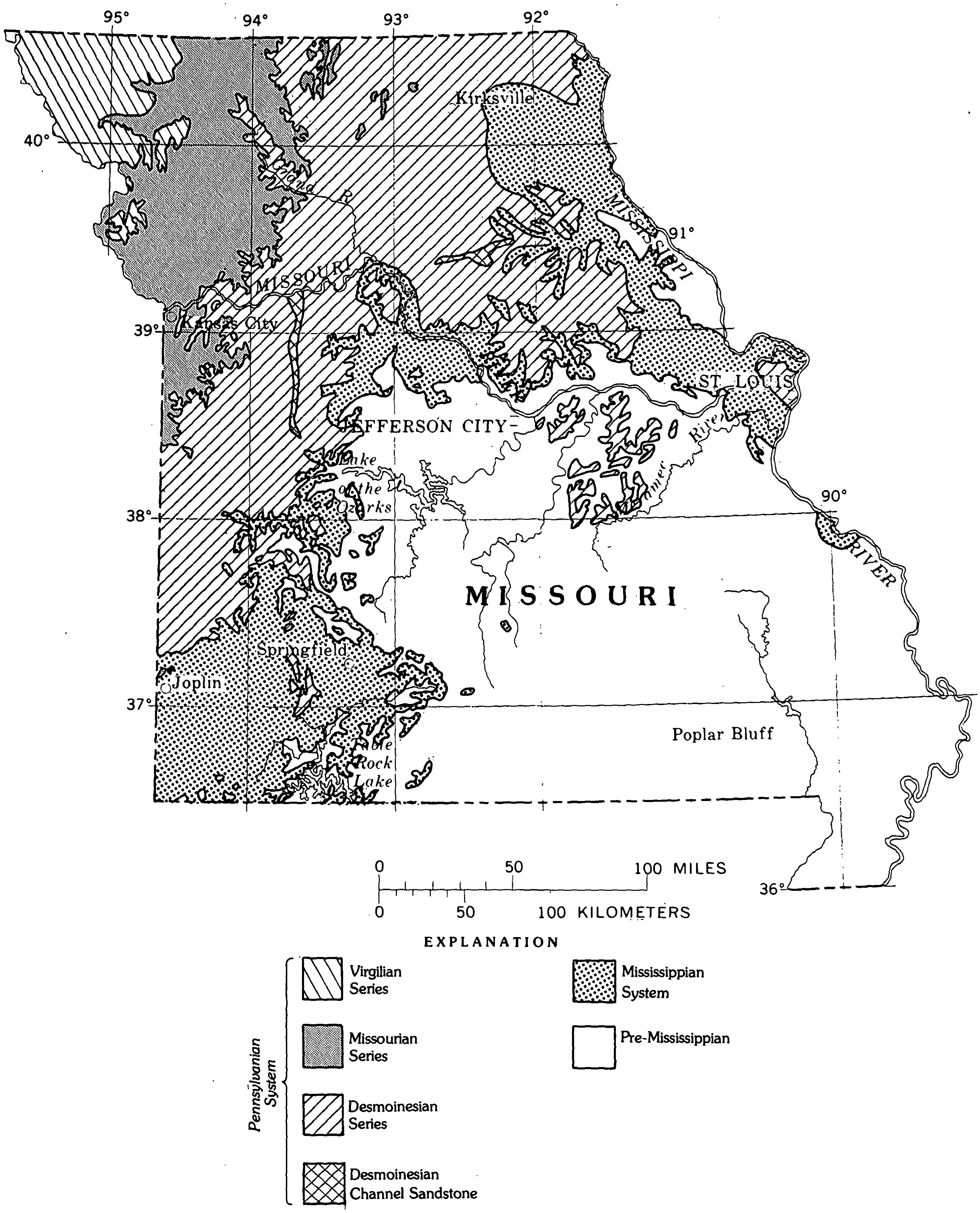
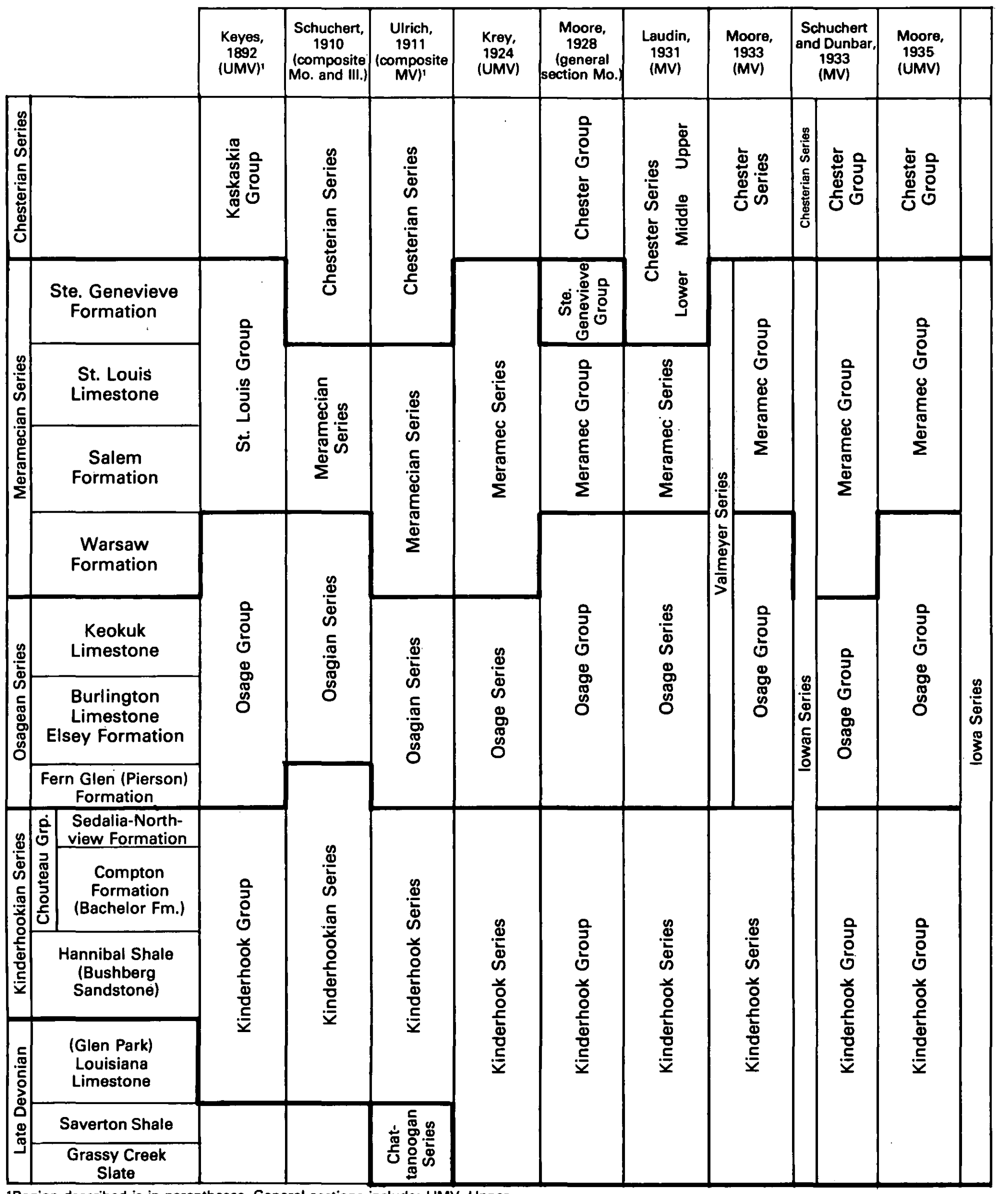


FIGURE 2.—Distribution of outcropping Mississippian and Pennsylvanian strata.

Carboniferous strata crop out over almost two-thirds of the State (fig. 2). Rocks south of the Missouri River valley (approximately the 39th parallel) are generally well exposed except where covered by loess (for example, in areas immediately adjacent to major river valleys) and (or) residual materials produced by weathering of the carbonate terrane. Outcrops are less common north of the Missouri River valley owing to cover by glacial drift deposited during the Kansan and Nebraskan episodes of Pleistocene ice advance. Here, best exposures are found where streams have cut down through the drift to re-expose the Paleozoic strata.

MISSISSIPPIAN SYSTEM

Rocks of the Mississippian System have been divided into four series: Chesterian Series (Worthen, 1866); Meramecian Series (Ulrich, 1904); Osagean (or Osagian) Series (Williams, 1891); and Kinderhookian Series (Meek and Worthen, 1861). Weller and others (1948, chart 5) stabilized this nomenclature as the "standard stratigraphic section" for the Mississippian System in North America (a condition lacking in nomenclature of Pennsylvanian stratigraphy). Figure 3 depicts the conceptual changes of the four Mississippian series



[1]Region described is in parentheses. General sections include: UMV, Upper Mississippi River Valley; MV, Mississippi River Valley.

FIGURE 3.—Development of nomenclature of Mississippian series in Missouri and surrounding midcontinent region.

from 1892 to the present. (See Thompson and Anderson, 1976, for synonymies of Mississippian nomenclature published for strata in Missouri.) Locations for the type areas of the four Mississippian series are shown on figure 1.

Although the four-part division of the Mississippian is more or less standard for North America, inability to agree on the placement of the Osagean-Meramecian boundary (whether it is above or below the Warsaw Formation or even whether such a boundary exists) led to a three-part division of the Mississippian in some States; Valmeyeran Series (Weller and Sutton, 1933; fig. 3) replaced Osagean and Meramecian in Illinois and Indiana. Kansas (Jewett and others, 1968) recognized "Lower Mississippian" and "Upper Mississippian" Series (fig. 3), the former composed of the Kinderhookian and Osagean Stages, the latter of the Meramecian and Chesterian Stages. However, this is nomenclatural variation and does not change the overall concept of time-stratigraphic correlation of the Mississippian System in the midcontinent region.

Rocks of the Mississippian age crop out extensively across Missouri (fig. 2). They are or have been economically important as a limestone resource where exposed and as host rock for lead and zinc

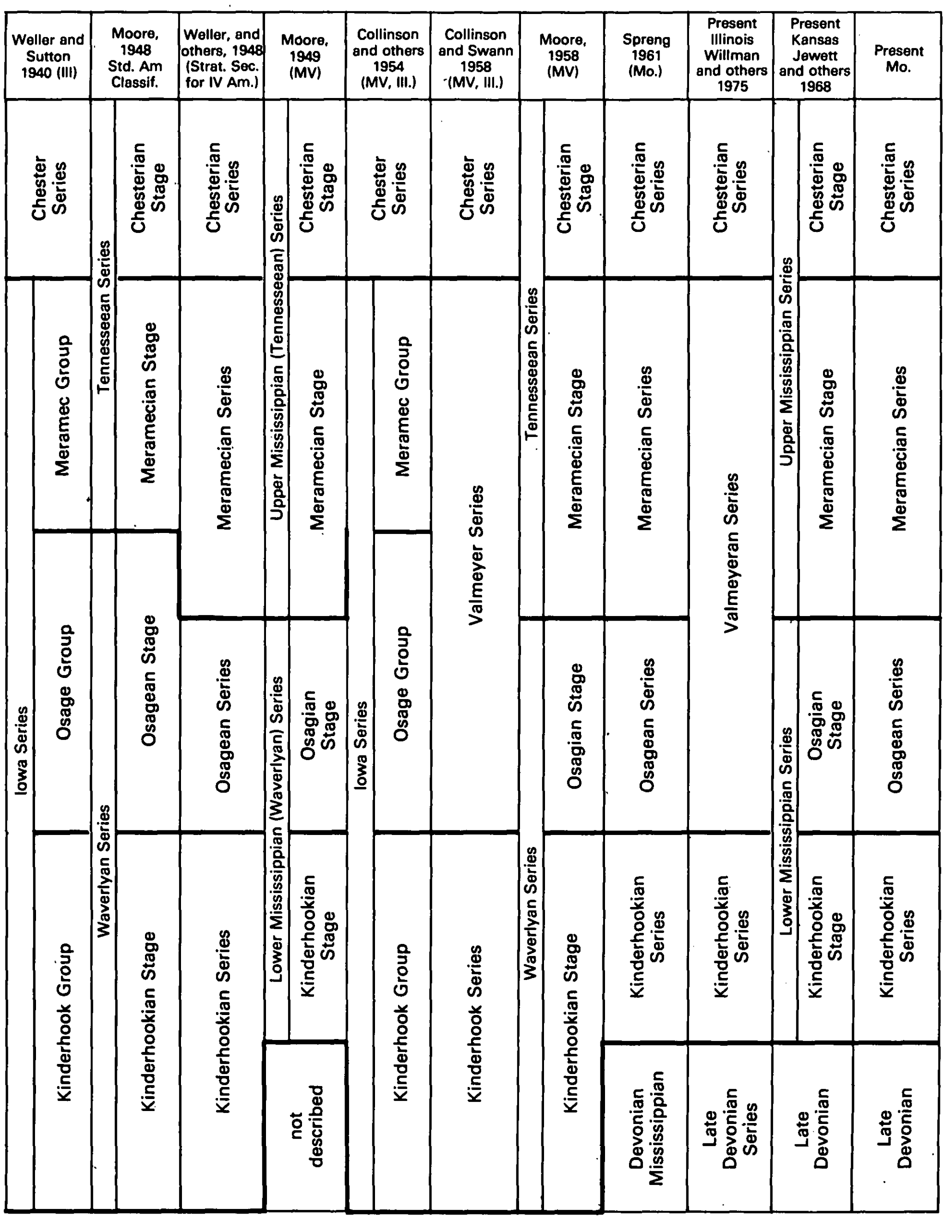


FIGURE 3.—Continued

deposits in the presently inactive Tri-State mining district of Missouri-Kansas-Oklahoma.

KINDERHOOKIAN SERIES

The "Kinderhook group" was named by Meek and Worthen (1861, p. 288) for strata exposed in the east bluffs of the Mississippi River valley immediately north of the small town of Kinderhook, Pike County, Ill. Formations identified within the Kinderhookian Series in Missouri include

Western and southwestern	*Northeastern*
Chouteau Group:	Chouteau Limestone
Northview Formation	
Sedalia Formation	
Compton Formation	Hannibal Shale
Bachelor Formation	

Originally this series also included strata later determined to be Late Devonian in age (Grassy Creek and Saverton Shales, Louisiana Limestone) and other rocks now assigned to the Osagean Series (Fern Glen Formation: see Weller, 1909, and Schuchert, 1910). As late as 1948 (Weller and others), the Grassy Creek, Saverton, and Louisiana formations were still identified as Kinderhookian in age. Scott and Collinson (1961) determined that the Devonian-Mississippian boundary was between the Louisiana Limestone and the overlying Hannibal Shale, finally defining the base of the Kinderhookian Series as now recognized. One of the best exposures of Kinderhookian strata in the type region (fig. 1) is a section along the Burlington Northern Railroad track immediately south of the town of Hannibal, Marion County, Mo., where the entire Hannibal Shale is exposed (the entire Kinderhookian interval in this region), along with the underlying Louisiana Limestone (Koenig and others, 1961, p. 44–46, stop 11).

The basal Kinderhookian formation in most of the Missouri outcrop area, other than the northeastern (Hannibal Shale) and southeastern (Bushberg Sandstone) regions, is the Bachelor Formation (Mehl, 1960, 1961), a very thin (5–10 cm), distinctive, persistent light-green to tan calcareous-cemented quartz sandstone that overlapped all pre-Mississippian strata during middle Kinderhookian time. This sandstone and overlying thin shale lie directly on strata of Ordovician, Silurian, Devonian, and even early Kinderhookian age (Bushberg Sandstone of southeastern Missouri). Conodonts recovered from the Bachelor (Thompson and Fellows, 1970) indicate basically the same age (that is, the same conodont zone, fig. 4) for the Bachelor throughout its outcrop from northwestern Arkansas to east-central Missouri. Late Devonian conodonts have commonly been reworked into the middle Kinderhookian fauna, even where the Bachelor rests on Lower Ordovician rocks, indicating that a widespread Upper Devonian sea existed before pre-Mississippian regression. The quartz sand of the Bachelor was most likely derived from Upper Devonian sandstone (Sylamore Sandstone), a nearshore facies of the Chattanooga Shale (Freeman and Schumacher, 1969).

Older Kinderhookian is represented only in the limited outcrop region of the Hannibal Shale on the western margin of the Illinois basin. Most of Missouri was still above sea level until middle Kinderhookian, when relatively sudden inundation resulted in the deposition of the Bachelor Formation. Stability soon followed, and the basal carbonate facies of the Chouteau Group (Compton facies) and its attendant crinoid fauna were deposited over most, if not all, of the Missouri region.

Throughout Missouri, Kinderhookian strata are predominantly carbonate rocks, either limestone or dolomitic limestone, except in the type area in northeastern Missouri where introduction of argillaceous material from the east produced the Hannibal Shale. In places, chert is present in the carbonate rocks, but not in the amounts found in some of the overlying beds of Osagean limestone. North of the Ozark uplift area (generally in the subsurface north of the Missouri River), Kinderhookian shales continue westward, resting on shale and limestone

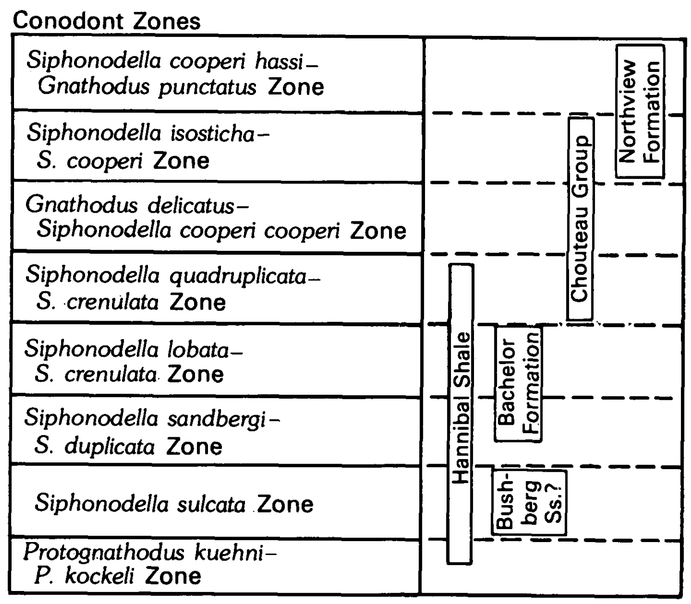


FIGURE 4.—Kinderhookian conodont zones (from Collinson, Rexroad, and Thompson, 1971) showing relative ages of Kinderhookian formations in Missouri.

of Late Devonian age. These Kinderhookian (Hannibal) shales continue in the subsurface to northwestern Missouri (Wells, 1960), where they rest on Upper Devonian shale.

Kinderhookian strata other than the basal Hannibal Shale are composed chiefly of limestone and dolomitic limestone (Chouteau Group) except for a northwest-trending "basin" of dolomitic silt and shale (Northview Formation, fig. 5) as much as 25 m thick in the vicinity of Springfield in southwestern Missouri (Clark and Beveridge, 1952; Thompson and Fellows, 1970). The Compton Formation, a limestone beneath the Northview, can be traced into Arkansas and Oklahoma and into the subsurface of eastern Kansas; Manger and Shanks (1976) recognized it as a member of the St. Joe Limestone in Arkansas. Statewide, the Kinderhookian carbonate rocks include fossiliferous limestone (with or without chert), oolitic limestone, silty and (or) argillaceous limestone, dolomitic limestone (with or without chert), dolomite, and even lithographic limestone. Interpretation of depositional environments is locally complicated by facies changes, as several of the above-mentioned lithologic types occur close together both horizontally and vertically.

The Chouteau in its type region in central Missouri, first described by Shallow (1855), is, from the base to top (fig. 6): fossiliferous limestone (Compton facies, units 3 and 4); lithographic limestone interfingering with silty, dolomitic limestone (unit 5); lithographic limestone ("Chouteau limestone", unit 6); silty, dolomitic, cherty limestone (Sedalia facies, unit 7); and dolomitic limestone (unit 8). The entire group is not more than 15 m

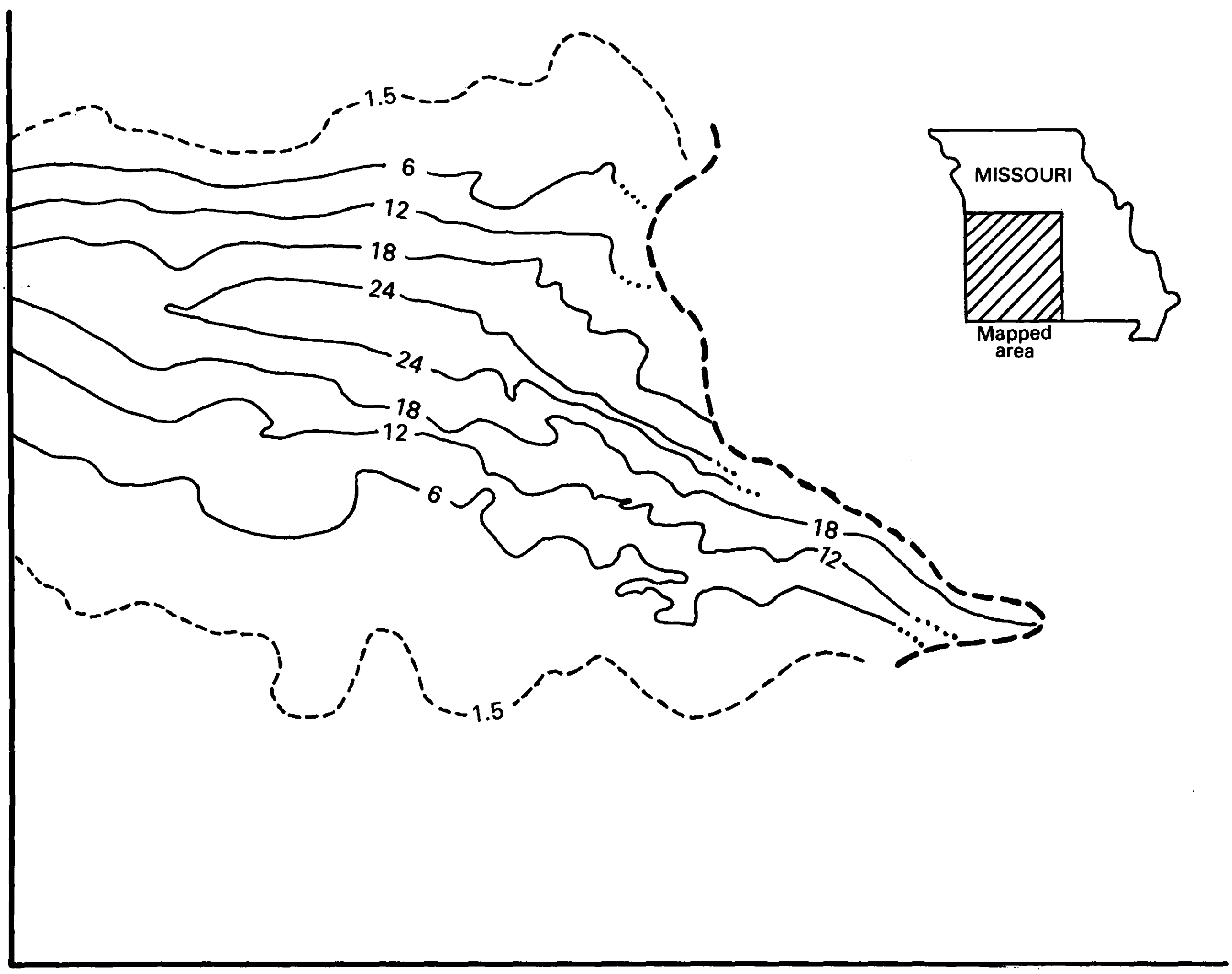


FIGURE 5.—Isopach map of the Northview Formation, southwestern Missouri (modified from Beveridge and Clark, 1952). Isopachs in meters.

thick. An isopach map of the Chouteau Group in southwestern, west-central, and central Missouri (fig. 7) shows the "Northview basin" and the increasing thickness of Chouteau strata to the northeast and north into central Missouri.

An unpublished study of the Chouteau Group of Missouri (Frey, 1967) described five carbonate facies in detail and presented a cross section of Chouteau strata from southwestern to northeastern Missouri (fig. 8). In Frey's interpretation of Chouteau strata, the Chouteau consists of the Compton, Sedalia, and Northview Formations; no "Chouteau limestone" was specifically identified by name. Frey did recognize two distinct facies, other than transitional facies, that had not been previously named formally. The "dolomitic limestone facies" (unit D, fig. 8) has been referred to in older literature as the "Chouteau limestone" in central Missouri. Facies unit E is a medium to coarsely crystalline fossiliferous limestone in northeastern Missouri and west-central Illinois; it is quite unlike the other Chouteau facies, except possibly for the Compton facies (unit F). The facies E grades into the upper part of the Hannibal Shale in northeastern Missouri, whereas the Compton facies (F) appears to be at the base of the Chouteau interval throughout its outcrop.

OSAGEAN SERIES

The "Osage group" was named by Williams (1891, p. 169) without further definition or discussion as part of his "Mississippian series." Later geologists differed in their definition of this group. Keyes (1892), Weller (1914), and Laudon (1948) included the Warsaw Formation as the upper unit of the Osagean, whereas Weller (1898), Moore (1928), and many other geologists (Weller and others, 1948) placed the top of the Osagean at the base of the Warsaw, the top of the underlying Keokuk Limestone. It was this controversy over the placement of the Osagean-Meramecian boundary that led to the establishment of the Valmeyeran Series in Illinois (Weller and Sutton, 1933) to include strata previously assigned to both former series.

The type area of the Osagean Series is the Osage River valley near Osceola, St. Clair County, west-central Missouri (fig. 1). However, in this region, Osagean exposures are poor, and only part of the series is represented. Osagean formations recognized in Missouri include, in descending order: Keokuk Limestone, Burlington Limestone, Elsey Formation ("lower Burlington" of eastern Missouri), and the Reeds Spring and Pierson Formations (Fern Glen Formation of eastern Missouri).

Strata of the Osagean Series are more than uniform in lithology than those of the Kinderhookian, and the formations are thicker. Except for a thin (2–3 m) shale or shaly limestone in the lower part near St. Louis in east-central Missouri (part of the Fern Glen Formation; Thompson, 1975), the Osagean is almost without exception a fossiliferous limestone or a cherty limestone. Lower Osagean limestones can be very cherty, and those in southwestern and east-central Missouri (Reeds Spring and Elsey Formations, upper part of the Fern Glen and lower part of the Burlington) commonly consists of more than 50 percent chert in at least part of the sequence. Upper Osagean limestones (Burlington and Keokuk Limestones) are extremely fossiliferous, the Burlington Limestone being more than 90 percent fossil debris throughout much of its outcrop. Upper Osagean limestones, more than 50 m thick, do contain some chert, but as 1 to 3-m zones separated by 10 to 15-m intervals of chert-free limestone.

MERAMECIAN SERIES

The "Meramec group" was named by Ulrich (1904, p. 110) for exposures and quarries in the "Meramec Highlands" adjacent to the Meramec River, St. Louis County, Mo. (fig. 1). Some early geologists excluded Warsaw and (or) Ste. Genevieve strata from this series, but Weller and others (1948) defined this series as it is now recognized. Meramecian formations in Missouri include, in descending order: Ste. Genevieve Formation, St. Louis Limestone, Salem Formation, and Warsaw Formation.

Meramecian rocks are, with the exception of the Warsaw and lower Salem of east-central Missouri, like the underlying Osagean, a thick sequence of fossiliferous limestone. Only near the top (Ste. Genevieve Formation) does the clastic content become significant, as an increasing amount of quartz sand indicates the beginning of instability that culminated in post-Mississippian–pre-Pennsylvanian uplift.

To the east, in the Michigan and Illinois basins, a sequence of deltaic siltstones and shales (Borden Siltstone: see Lineback, 1969) built up from the east toward the Mississippi River valley region. During early Meramecian time, this clastic material spilled westward over the Osagean carbonate bank that was forming west of the Illinois basin,

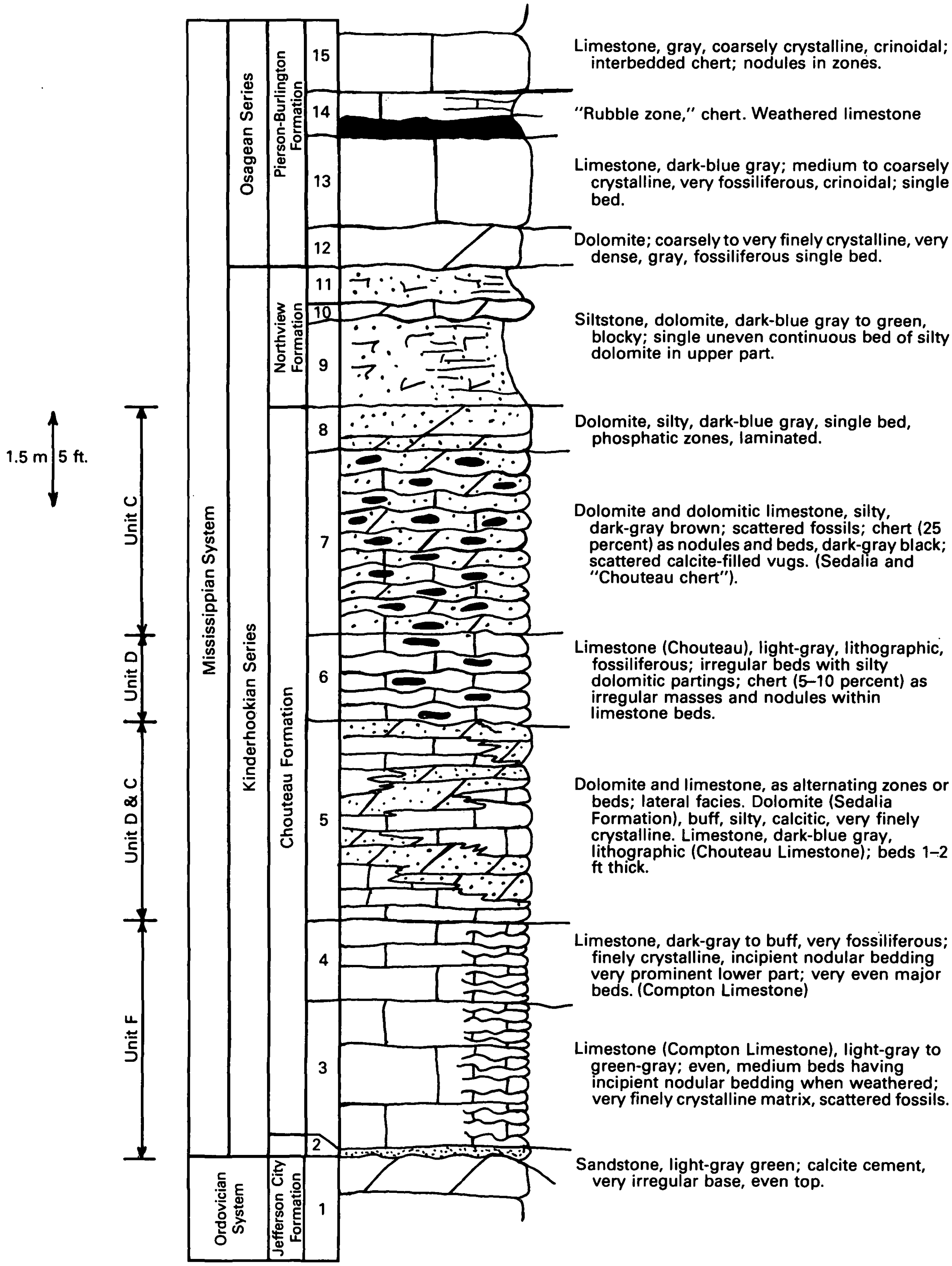


FIGURE 6.—Stratigraphic column of "Chouteau Group" section in the vicinity of Sedalia, Mo., within the type region of the Sedalia Formation (location at *A′* on fig. 8). Unit letters designate facies shown on figure 8.

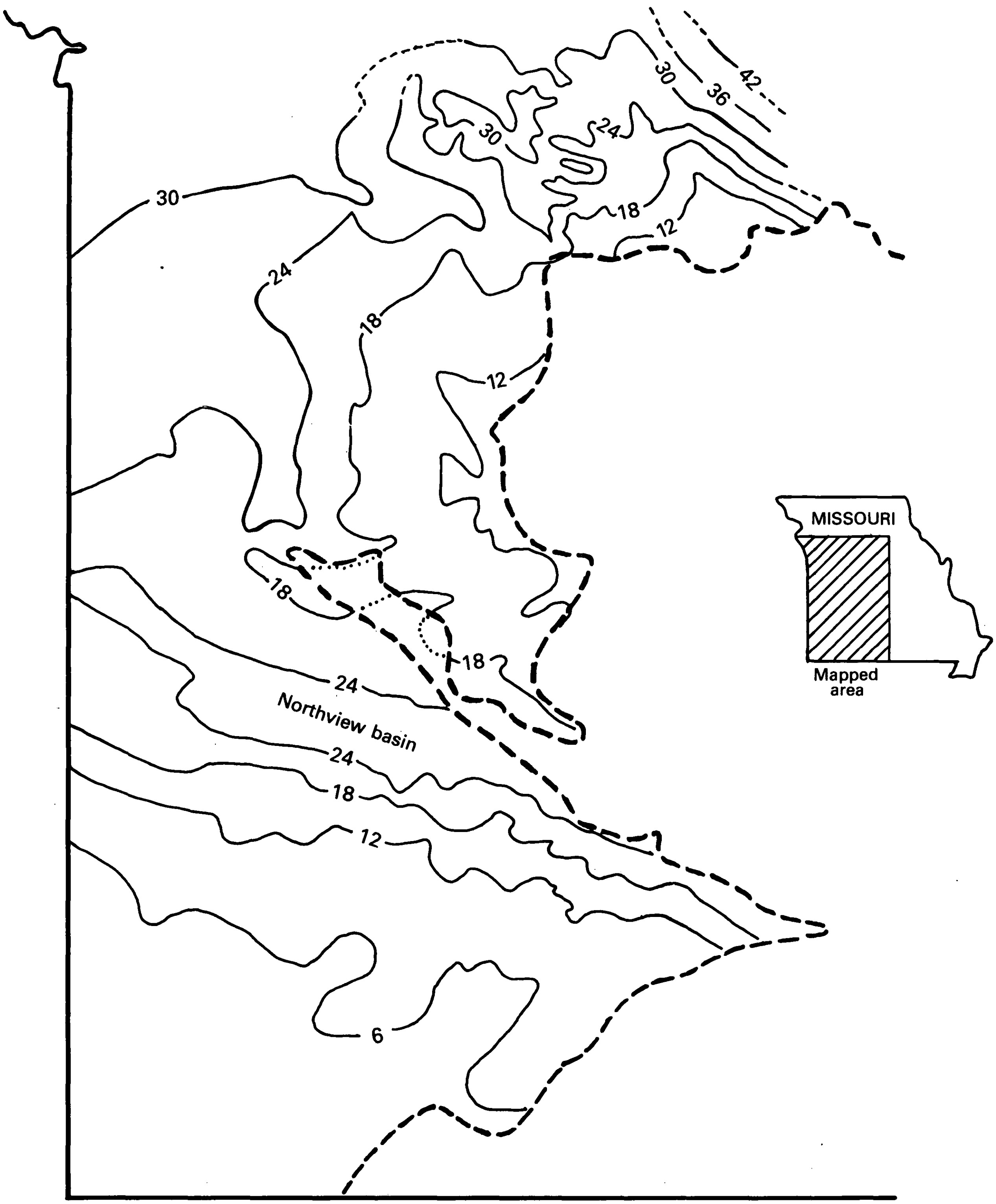


FIGURE 7.—Isopach map of the Chouteau Group, southwestern and western Missouri (modified from Beveridge and Clark, 1952). Isopachs in meters.

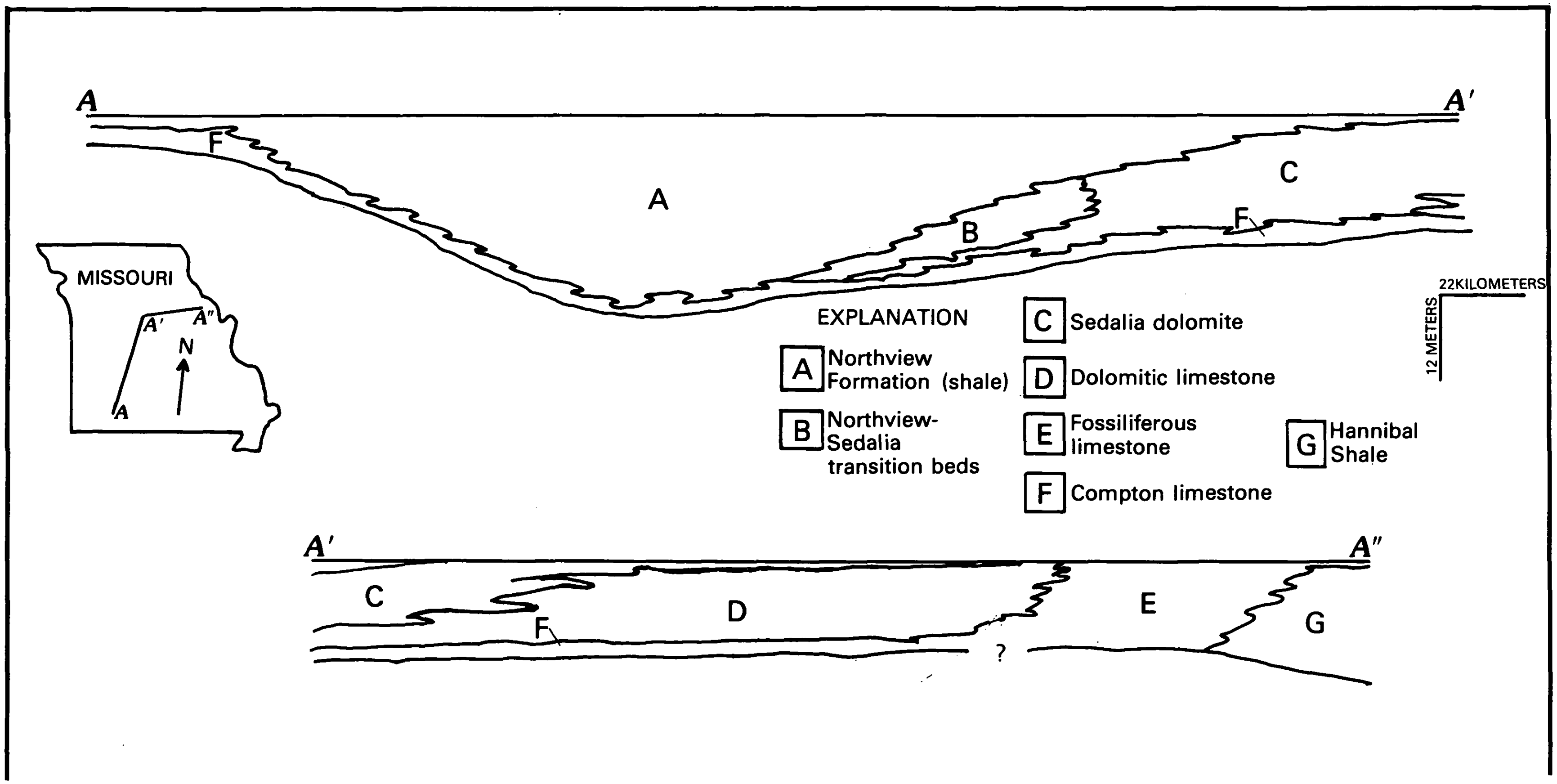


FIGURE 8.—Stratigraphic cross section of Kinderhookian strata from southwestern (*A*) to northeastern (*A''*) Missouri, illustrating carbonate facies of the Chouteau Group (modified from Frey, 1967).

stopping carbonate deposition by killing the predominantly echinoderm fauna responsible for a large part of the Osagean carbonate sediment and depositing shale and siltstone of the Warsaw Formation (a condition similar to that during deposition of the Hannibal Shale of Kinderhookian age). The Warsaw shales grade westward into calcareous shale and limestone, and finally in western Missouri, the sequence is again an unbroken one of fossiliferous limestone that shows little evidence of unconformity at the base of the Warsaw. The Keokuk-Warsaw contact is marked only by a thin (1–2 m) oolite (Short Creek Oolite Member of Keokuk Limestone).

The stable depositional conditions that produced a growing carbonate bank throughout Kinderhookian and Osagean time continued virtually unabated until upper Meramecian, when depositional environments changed, indicated by anhydrite beds and increased clastic rocks (St. Louis and Ste. Genevieve Formations) along with inconsistent limestone types of the St. Louis Limestone. Ste. Genevieve strata reflect more closely those of the overlying Chesterian Series and have in fact been recognized as such by Swann (1963).

CHESTERIAN SERIES

Worthen (1866) named the "Chester group" from a sequence of sandstones and limestones exposed in the bluffs of the Mississippi River valley in the vicinity of Chester, Randolph County, Ill. (south of St. Louis, Mo., fig. 1). These strata represent a shifting environment that produced an alternating sequence of 10 formations now recognized in southeastern Missouri (five sandstone and five limestone): Vienna Limestone, Tar Springs Sandstone, Glen Dean Limestone, Hardinsburg Formation, Golconda Formation, Cypress Formation, Paint Creek Formation, Yankeetown Sandstone, Renault Formation, and Aux Vases Sandstone. Conditions had changed considerably from the stable conditions of limestone deposition of the preceding series.

In addition, a Chesterian sequence is present in southwestern Missouri (Thompson, 1972) that has little depositional relationship to that in eastern Missouri; it is more closely related to Chesterian strata in Oklahoma and Texas. This limestone and shale sequence (Hindsville Limestone and Fayetteville Formation) appears to have been deposited in a depositional basin separate from that of the eastern sequence, as the difficulty of correlation by conodonts indicates (Thompson, 1972).

Outcrops of Chesterian rocks in southwestern Missouri are very restricted, preserved as downdropped fault blocks surrounded by Osagean strata. Obviously, Chesterian strata covered a much larger part of southwestern Missouri at one time, but the maximum extent is not known. Outcrops in eastern Missouri are restricted to the Mississippi River bluffs in northern Perry and southern Ste. Genevieve Counties.

SUMMARY OF MISSISSIPPIAN SYSTEM

Mississippian rocks in Missouri indicate depositional stability from earliest Kinderhookian into late Meramecian time. Thus, during two-thirds of Mississippian time, the region was covered by a warm shallow sea in which a broad carbonate shelf formed that extended throughout the western interior region (Missouri, Iowa, Kansas, northern Arkansas, and northeastern Oklahoma). Exceptions were the influxes of shale and silt into northeastern and east-central Missouri from the east that are represented by the basal Kinderhookian Hannibal Shale and the basal Meramecian Warsaw Formation. Both these units grade westward into limestone.

A northwest-trending trough of silty shale and dolomitic siltstone (Northview Formation) was rapidly deposited during late Kinderhookian time in part of southwestern Missouri, but the source of this material has not yet been identified. Deposition of the shale and siltstone did no more than create a pause in the limestone deposition in the region.

Limestones of the Mississippian System are composed largely of disaggregated echinoderms and other bioclastic debris. The percentage of echinoderm material varies, and other fossil groups are represented to various extents. Some limestone units are high in matrix material; others are very low, composed almost entirely of crinoid debris (Burlington Limestone).

The Mississippian carbonate shelf extended a short distance into Arkansas, where it abruptly terminated. In less than 8 km the composite Kinderhookian-lower Osagean section in northern Arkansas decreased from more than 25 m to less than 0.5 m (Thompson and Fellows, 1970). The southern end of the shelf facies grades within a short distance into a sediment-poor ("starved") limestone facies, where the rate of deposition was much slower than that of the shelf.

An uncommon situation exists in east-central and southeastern Missouri from the Meramec River valley just west of St. Louis southward. Osagean strata

(Fern Glen and Burlington) appear to be similar to those in western and southwestern Missouri (Pierson to Burlington sequence), but except for a very thin basal Mississippian sandstone (Bachelor Formation and possibly the Bushberg Sandstone), Kinderhookian carbonate rocks are missing; and no evidence is seen of prior deposition or erosion of these carbonate rocks. Perhaps Kinderhookian time in eastern and southeastern Missouri was a period of "still-stand."

Pre-Mississippian strata range from Early Ordovician (Canadian Series) to Late Devonian (Famennian Stage) in age. This surface reflects post-Devonian–pre-Mississippian erosion that left Upper Devonian strata in northern Missouri and scattered places in southwestern Missouri (Grassy Creek and Chattanooga shales) but removed all strata down to Lower Ordovician in western and southwestern Missouri.

The Kinderhookian-Osagean boundary is represented by a hiatus throughout all but possibly the extreme southwestern part of the State (Thompson and Fellows, 1970). This short period has been recognized by a fauna of late Kinderhookian conodonts absent from the Mississippi River valley sections. The base of the Osagean Series, as identified by conodonts (Thompson and Fellows, 1970; Thompson, 1975), is consistent and well represented throughout most of the State.

The Osagean-Meramecian boundary is transitional in all but the east-central part of Missouri, where the abrupt change from Keokuk limestone to Warsaw shale marks this boundary. In western Missouri, this boundary is identified as the top of a thin (1–3 m) oolitic limestone bed separating the Keokuk and Warsaw. Little, if any, time is lost across this "boundary."

The Meramecian-Chesterian boundary in eastern Missouri has traditionally been placed between the Ste. Genevieve Formation and the overlying Aux Vases Sandstone, identified on the basis of crinoids (Swann, 1963). However, lithologically, the basal part of the Ste. Genevieve, where the first clastic sediments appear, better reflects the conditions attributable to a series boundary. Conodonts do not indicate a faunal change between Ste. Genevieve and Chesterian strata but do indicate a distinct change between St. Louis and Ste. Genevieve rocks.

MISSISSIPPIAN-PENNSYLVANIAN BOUNDARY

The Mississippian-Pennsylvanian boundary in Missouri is represented by a hiatus of variable but, in many places, considerable magnitude. Lower and (or) Middle Pennsylvanian strata rest on rocks ranging from Early Ordovician (Canadian) to Late Mississippian (Chesterian) in age. However, not all the inferred removal took place in post-Mississippian time; some took place in pre-Mississippian and also during Mississippian time. This hiatus contrasts sharply with the nearly transitional Mississippian-Pennsylvanian boundary within the second tier of counties south of the Missouri-Arkansas border in northwestern Arkansas (Washington to Independence Counties).

The youngest identified Mississippian rocks in southwestern Missouri constitute the middle Chesterian Fayetteville Formation (Thompson, 1972). No exposures of upper Chesterian strata (equivalent to Pitkin and Imo rocks in Arkansas, fig. 9) have been identified in southwestern Missouri. However, Thompson (1970) identified uncommon strata near an old coal pit in extreme southwestern Missouri as early Morrowan (Early Pennsylvanian) in age; the strata consist of marine sedimentary rocks that correlate, by conodonts, directly with the Prairie Grove Member of the Hale Formation in the type region of the Morrowan Series in Washington County, Ark. This correlation leads to the possibility that the transitional Mississippian-Pennsylvanian seas may have extended north into southwestern Missouri.

Current and unpublished studies by the author describe late Morrowan and Atokan strata in the west-central part of Missouri that are preserved within low regions of an undulating erosional surface on middle Mississippian (Meramecian and Osagean) limestones. Thus, considerable erosion took place before the appearance of Morrowan seas in this region. Because the Chesterian formations of southwestern Missouri rest on the Osagean Keokuk Limestone (the entire Meramecian and upper Osagean are missing), much of the pre-Pennsylvanian erosion might have been pre-Chesterian as well.

Late Morrowan and Atokan sediments were deposited in extreme western Missouri before initial deposition of the more widespread Desmoinesian sandstones. The total present extent of pre-Desmoinesian strata is quite limited. The overlying continental Desmoinesian sandstones overlapped the Atokan shales eastward over the Mississippian erosional surface, extending well beyond the known margin of Atokan deposition. Between the top of the Mississippian and basal Pennsylvanian sandstones is the "basal Pennsylvanian conglomerate,"

Pennsylvanian System	Atokan Series	Atokan Formation (Winslow Formation)	
	Morrowan Series	Bloyd Formation	Kessler Limestone Member Dye Shale Member Woolsey Member Brentwood Limestone Member
		Hale Formation	Prairie Grove Member Cane Hill Member
Mississippian System	Chesterian Series	Imo Formation Pitkin Formation Fayetteville Formation Batesville Sandstone Hindsville Limestone	

FIGURE 9.—Upper Mississippian (Chesterian Series) and Lower Pennsylvanian (Morrowan Series) formations of northwestern Arkansas.

composed of well-rounded boulders and pebbles of Mississippian chert within a quartz sand matrix (Graydon Formation).

Over much of the basal Pennsylvanian-outcrop region, the oldest Pennsylvanian sedimentary rocks are those of the Desmoinesian Cherokee Group (including the sandstone mentioned above). These sedimentary rocks are generally of a continental nature, a complex of distributary channel sandstones, interbedded shales and coals, and widespread sheet sandstones. In central and eastern Missouri, Pennsylvanian clay, shale, and sandstone were deposited on a karst surface of Lower Ordovician (Canadian) dolomites. These strata (Cheltenham Formation), previously considered Atokan in age (Searight and Howe, 1961), are now regarded as "unassigned" (Anderson and others, 1979). The lowest recognizable Pennsylvanian beds overlying these clay and shale beds are basal members of the Marmaton Group. Cheltenham strata are questionably correlated with Cherokee strata elsewhere.

PENNSYLVANIAN SYSTEM

Pennsylvanian strata of North America are not as easy to relate on a continent-wide basis as are those of the Mississippian because of the distinct differences in lithologic sequences between major depositional regions (such as the Pennsylvania coal basin, Illinois basin, and western interior coal basin of Kansas, Iowa, Oklahoma, Nebraska, and Missouri). Therefore, several separate classifications exist, each applicable to a particular region. As described by Moore and others (1944, chart 6) the "Mid-Continent Region" Pennsylvanian System consists of five series, originally identified as formations, bounded by major unconformities within the Pennsylvanian sequence. They are, in descending order: Virgilian Series (Moore, 1932), Missourian Series (Keyes, 1894), Desmoinesian Series (Keyes, 1894), Atokan Series (Taff and Adams, 1900; Lampasas Series, in Moore and others, 1944) and Morrowan Series (Adams and Ulrich, 1905). These unconformities often are marked by the presence of bodies of sandstone resting as sheets and (or) channels on an eroded surface of the previous series.

Pennsylvanian stratigraphic units in the midcontinent region (western-interior coal basin) are distinctive in their extreme thinness, despite great lateral persistence. The same formations—even members of formations—extend over Kansas, Nebraska, Iowa, and Missouri (and some into Oklahoma). A single shale bed 1 m thick can be identified from western Kansas to northern Missouri, and one of the coal beds (Crowebrug) has been correlated from Kansas to Pennsylvania (Hopkins and Simon, 1975, p. 187).

In this complex stratigraphic sequence, a great number of formations have been recognized—69 presently in Missouri, alone—composed of more than 100 formally recognized members. Because the Pennsylvanian depositional history was one of cyclic sequences (Weller, 1930; Moore, 1930, 1931), formations have been combined into groups by cyclic similarities. Originally proposed as formations, each of the nine formally named groups possesses peculiar and distinct attributes that distinguish it from others. The groups are bounded by unconformities of regional distribution. For clarity, and to best describe both physical and economic conditions, Pennsylvanian strata in Missouri will be detailed to the group level (fig. 10).

MORROWAN AND ATOKAN SERIES

Adams and Ulrich (1905) defined the "Morrow formation" from outcrops in Washington County, Ark. Although identified as the "Morrow series" by Moore (1932), Henbest (1962a,b) formally defined the Morrowan Series as presently recognized.

Taff and Adams (1900) named the "Atoka shale" from a thick sequence exposed in Atoka County,

Virgilian Series
Wabaunsee Group Shawnee Group Douglas Group
Missourian Series
Pedee Group Lansing Group Kansas City Group Pleasanton Group
Desmoinesian Series
Marmaton Group Cherokee Group
Atokan Series
Morrowan Series

FIGURE 10.—Stratigraphic framework for Pennsylvanian System in Missouri, to group level.

Okla. Unfortunately, this exposure is essentially unfossiliferous, and the inability to determine accurately the relationship between type Atokan strata and those of the type Morrowan in Arkansas has been the basis of a controversy over the validity of the Atokan Series as a time-stratigraphic (chronostratigraphic) unit (see Shaver and Smith, 1974). Moore and others (1944) defined this sequence as the Lampasas Series, from a section exposed in Texas. Dunn (1976) and Webster (1969) identified this interval as the Derryan Series, named by Thompson (1942) from a section in New Mexico. However, the relationship of Derryan strata to type Morrowan is also unclear because Thompson did not recognize Morrowan in his sections but identified all pre-Desmoinesian strata as "Derryan." Atokan Series is still used in the midcontinent region for lack of a more clearly defined series.

Distribution of Pennsylvanian strata older than Desmoinesian in Missouri is not well known; strata are generally discontinuous and in some areas are limited to remnants preserved in collapse or fault structures (fig. 11). Before 1970, no rocks of Morrowan age had been positively identified in Missouri. Sandstone capping outliers of Chesterian (Upper Mississippian) strata in extreme southwestern Missouri (formerly identified as the Hale Formation; Searight and Howe, 1961; Thompson, 1972) has recently been identified as Mississippian in age by E. Glick (Wedington Sandstone Member of the Fayetteville Formation (oral commun., 1975).) Thompson (1970) recovered a conodont fauna of definite Morrowan age from one isolated outcrop preserved as a block within a fault zone in southwestern Missouri; he correlated these rocks with the early Morrowan Prairie Grove Member of the Hale Formation in the type Morrowan region of northwestern Arkansas.

Beneath the shale of the Riverton Formation in west-central Missouri is a thin (1–3 m) calcium carbonate-cemented quartz sandstone (hereafter termed "calcareous sandstone") that lies directly on the eroded surface of Mississippian limestone. This sandstone was included within the Riverton by Searight (1959). However, it has yielded a conodont fauna that indicates a Morrowan age. Conodonts recovered from the shale of the overlying Riverton Formation indicate an Atokan age.

The Burgner Formation (Searight and Palmer, 1957), named from a core taken in a sink structure near Joplin, Mo., was dated by fusulinids (Thompson, 1953) as mid-Atokan in age. In the type Burgner Core, limestone of the limestone, shale, and coal sequence yielded a conodont fauna younger than that recovered from the calcareous sandstone

<table>
<tr><td rowspan="7">Pennsylvanian System</td><td rowspan="2">Desmoinesian Series</td><td rowspan="2">Cherokee Group</td><td rowspan="2">Krebs Subgroup</td><td>Warner Formation</td></tr>
<tr><td>Hartshorne ? Formation</td></tr>
<tr><td rowspan="4">Atokan Series</td><td rowspan="4" colspan="2"></td><td>Riverton Formation</td></tr>
<tr><td>Burgner Formation</td></tr>
<tr><td>McLouth Formation (subsurface only)
- - - - - - - ? - - - - - - -</td></tr>
<tr><td>Cheltenham Formation
- - - - - - - - - ? - - - - - - - - -</td></tr>
<tr><td>Morrowan Series</td><td colspan="2"></td><td>Hale Formation</td></tr>
</table>

FIGURE 11.—Lower Pennsylvanian sequence in western Missouri, as presented by Searight and Howe (1961).

beneath the Riverton shales in west-central Missouri but definitely older than that recovered from the Riverton shales. Burgner strata are interpreted to lie approximately at the Morrowan-Atokan boundary and to be of very early Atokan age in at least the upper part. The Burgner, like the Riverton in the Joplin area, is found primarily within collapse or filled-sink structures in Mississippian limestones. The limestone of the Burgner Formation is definitely marine, as is the calcareous sandstone, and thus its presence indicates Morrowan to early Atokan seas in southwestern Missouri (Thompson, 1970) and into west-central Missouri.

The Riverton Formation (fig. 11), beneath the basal Krebs Subgroup of the Cherokee Group, has been preserved in several sink structures around Joplin. Riverton strata are also known north of Joplin in west-central Missouri beneath the basal formation of the Krebs Subgroup (Warner Formation). The pre-Pennsylvanian surface in west-central Missouri is broadly undulating. On the highest parts, the sandstone of the basal Krebs Warner Formation rests directly on Mississippian strata. The Riverton strata and underlying Morrowan calcareous sandstone are increasingly thick toward the lowest parts of the undulating surface. The calcareous sandstone is generally present only in the deepest parts, is more restricted than the overlying shale, and has been identified in only one small area (fig. 12). In the region of Joplin, beds of the Riverton Formation (coal, shale, and sandstone) have been distorted and dip steeply into circular collapse structures; this position indicates that the Riverton beds were let down into the structures after they were deposited. Much of the lead and zinc recovered from the Missouri part of the Tri-State mining district was obtained along the walls of these circular structures.

Searight and Howe (1961) regarded the Riverton as late Atokan in age (fig. 11). Conodonts recovered from the basal part of the Riverton at two localities (Thompson, unpub. data), one an open sink structure near Joplin, the other an outcrop in west-central Missouri, clearly indicate an Atokan age for the Riverton Shales; that is they are younger than the Burgner Formation. From this, the relationship of pre-Desmoinesian formations of western Missouri is shown in figure 13.

FIGURE 12.—Areas of known Morrowan and Atokan strata. ×, lower Morrowan strata; ○, Riverton and Burgner Formations near Joplin, Missouri, Atokan in age; □, Riverton Formation (Atokan) and underlying Morrowan calcareous sandstone.

DESMOINESIAN SERIES

The "Des Moines formation" was named by Keyes (1894, p. 82) to "represent the lower Coal Measures, or the marginal deposits of the Upper Carboniferous." The type locality is along the Des Moines River in central Iowa. Originally, the "Des Moines series" (Keyes, 1896) extended to the base of the present Kansas City Group and included the section now named the Pleasanton Group (lower Missourian Series). Moore (1932) restricted the "Des Moines series" to its present limits, which includes only the Cherokee and Marmaton Groups. The name was combined to "Desmoinesian" by Moore (1948).

Desmoinesian strata are divided into two named groups, the Cherokee and overlying Marmaton. They are quite distinct from each other, the Cherokee composed almost entirely of shale, clay, sandstone, and coal beds, the Marmaton, of alternating limestone and shale and of minor sandstone bodies associated with the thicker beds of shale.

Desmoinesian Series	Warner Formation
Atokan Series	Riverton Formation Burgner Formation "calcareous sandstone" "correlative of Prairie Grove Member of Hale Formation"
Morrowan Series	

FIGURE 13.—Proposed sequence of Lower Pennsylvanian formations in western and southwestern Missouri.

Cherokee Group.—Most of the minable coal beds in Missouri are found in the Cherokee Group. Haworth and Kirk (1894, p. 105) named the "Cherokee formation" for strata of the "lower division of the Lower Coal Measures" from exposures in Cherokee County, Kans. Depositional conditions were complex, most sedimentary rocks being sandstone (occurring as shallow channels (of a distributory nature) and broad sheets), siltstone, clay (underclay), overlying coal beds, and a few thin beds of limestone. Lower Cherokee strata contain thick coal-bearing shales between and contemporary with distributary-channel and widespread sheet sandstones. Sandstone is less abundant in upper Cherokee strata, where clay and shale are dominant.

Marmaton Group.—Marmaton strata are an alternating sequence of limestone and shale; several of the limestone beds are thick enough for quarrying. A coal bed (Summit coal) is present above the basal limestone but is minable only in a few restricted areas. Haworth (1898, p. 92) named the "upper division of the Lower Coal Measures" the "Marmaton formation" from strata exposed along the Marmaton River in eastern Kansas. Originally including Pleasanton strata in the upper part, Moore (1932) restricted the group to its present definition. Many early geologists called this sequence the "Henrietta formation," or "Henrietta group" (named by Keyes, 1898), but Greene and Searight (1949) formally abandoned this name in favor of Marmaton.

Marmaton strata are an alternation of two sedimentary cycles: a sequence of thin limestone-shale-limestone units, and a thick shale (in many places associated with channel sandstone bodies and coal beds). Four of each of these basic cycles constitute the Marmaton. The top of the group is marked by major channel sandstone of the basal Missourian Pleasanton Group.

MISSOURIAN SERIES

Keyes (1894, p. 82) named the "Missouri formation" to correspond to the "upper Coal Measures." The type locality is in the bluffs of the Missouri River valley in northwestern Missouri. Originally defined to include all Pennsylvanian strata above the Desmoinesian, Moore (1932) restricted the definition to include beds between two important unconformities, one at the base of the Pleasanton, the other (top) between the Pedee and overlying Douglas Groups.

The Missourian Series consists of four groups: the basal Pleasanton Group, which is an unconformable continental sandstone and shale sequence on the underlying marine Desmoinesian (Marmaton) strata, and the predominantly marine Kansas City, Lansing, and Pedee Groups.

Pleasanton Group.—The "Pleasanton shales" (Haworth and Bennett, 1896, p. 44) were named for a sequence of shale and sandstone cropping out in the vicinity of Pleasanton, Linn County, eastern Kansas. Although almost entirely a clastic unit, a conspicuous limestone (Exline in Missouri, Checkerboard in Kansas) about one-fourth to one-half meter thick, is persistent in the lower part of the group. One important aspect of Pleasanton strata is in the impressive channel sandstones exposed in west-central (Warrensburg Sandstone Member) and north-central (Moberly Sandstone Member) Missouri, representing remnants of a once-extensive drainage system in early Missourian time.

Kansas City Group.—The "Kansas City formation" (Hinds and Greene, 1915, p. 23) is a sequence of alternating beds of limestone and shale. Formations are defined on the basis of cycles of limestone-shale-limestone formations alternating with shale or shale-limestone-shale formations. In all, 17 formations have been named in the Kansas City Group, of which 5 are shale units not divided into members. Several limestone members are thick (3 m) and have been extensively quarried in the Kansas City region.

Complex cyclic sedimentation is represented in some of the Kansas City formations. Sequences of distinctive associations (Swope Formation) are repeated almost exactly in younger successions in the Virgilian Series (Shawnee Group).

Lansing and Pedee Groups.—The Lansing and Pedee Groups (originally part of the Kansas City Group) continue the limestone-shale-limestone sequence. The Pedee Group was removed by pre-Virgilian erosion in some places north of Kansas City; this removal left a Missourian-Virgilian unconformity. The "Lansing formation" was first identified by Hinds (1912, p. 7), and the Pedee was named by Moore (1932, p. 88); the type localities are near the town of Lansing, Leavenworth County, Kans., and along the "Pedee Branch in vicinity of Weston, Missouri," in Platte County (Moore, 1936, p. 137).

VIRGILIAN SERIES

Moore (1932) named the "Virgil series" for strata near the town of Virgil, Greenwood County, Kans. Condra (1949, p. 11) located the type area as "on Verdigris River from west of Madison to Virgil and southeastward to central Wilson County." Virgilian strata are divided into three groups, the Douglas, Shawnee, and Wabaunsee. The formations of the last two are complex, and cyclic, some consisting of "megacyclothems," a certain repetitive cycle of cyclothems. Heckel (1968, 1975, 1977) and Heckel and Baeseman (1975) have described in detail the environmental sequences identified by specific rock types within these cyclic sequences.

Douglas Group.—Named for Douglas County, Kans., by Haworth (1898, p. 92, 93), the "Douglas formation" was proposed as a group by Moore (1932). The base of the group is marked by a channel sandstone (Tonganoxie Sandstone Member of Stranger Formation) in parts of northeastern Kansas and northwestern Missouri. The remaining Douglas strata are predominantly shale but contain a few thin limestone members.

Shawnee Group.—Haworth (1898, p. 93) named the "Shawnee formation" from strata exposed in Shawnee County, Kans. Moore (1932) redefined these strata as the Shawnee Group, composed of a sequence of "megacycles," each reflecting a complex cycle of sedimentation. Limestone beds predominate in this sequence, although some beds of thick shale are also present. Characteristic of these strata are certain units of thin shale that are laterally persistent over much of the midcontinent region.

Wabaunsee Group.—Named from strata in Wabaunsee County, Kans. (Prosser, 1895, p. 689), the "Wabaunsee formation" was redefined by Moore (1936) as the Wabaunsee Group. It is distinguished from the Shawnee Group by a larger relative percentage of shale and relatively more sand and sandstone in the shales. As stated by Moore (1936, p. 201), "A distinctive feature of the Wabaunsee group is the character of the cyclic sedimentary succession, which shows regularly alternating nonmarine and marine units in which a grouping of cyclothems in megacyclothems is not evident. This serves especially to set the Wabaunsee beds apart from those of the Shawnee group."

SUMMARY OF PENNSYLVANIAN SYSTEM

Whereas Mississippian strata represent a period of stable shallow marine deposition in the Missouri region, Pennsylvanian strata indicate a period of continuous change. Earliest Pennsylvanian sedimentary rocks (Morrowan Series) are known from one outcrop in southwestern Missouri, which may represent the extension of Morrowan seas of northwestern Arkansas into Missouri. Remnants of Atokan strata are primarily shales, which indicate less stable, more continental conditions. Strata of the lower Desmoinesian Cherokee Group are primarily sandstone and shale, which reflect a distributary, deltaic to coastal-plain environment; the fluctuating, very shallow swampy environment of the "Coal Measures" proper is best seen in the upper Cherokee Group.

Pennsylvanian strata of the upper Desmoinesian (Marmaton Group) and above, with the exception of the Pleasanton strata, generally consist of alternating limestone and shale. Conditions alternated from shallow marine to nonmarine continental. Marine phases consist of limestone and thin shale; these alternating with the continental phase, of thick shale, sandy shale, and (or) sandstone. Several coal beds formed during this continental phase are thick enough to be mined. However, most of the economically important coal beds are within the Cherokee Group.

Thus, in contrast with the stable conditions of Mississippian time, Pennsylvanian environments of deposition ranged from shallow open marine to complete regression to low-relief continental (deltaic and coastal plain).

MINERAL RESOURCES

Mineral production from Carboniferous strata in Missouri has consisted primarily of limestone, refractory clay, shale, lead, zinc, and coal. Few significant amounts of petroleum and natural gas have been recovered from these strata, although there is current interest in the "tar sands," or heavy oil in the lower Pennsylvanian sandstones of western Missouri.

MISSISSIPPIAN SYSTEM

Limestone is the single most important economic product derived from Mississippian strata in Missouri today. Both dimension and crushed and broken stone are produced from open-pit and drive-in mines. The limestone is used principally for construction aggregate, marble and dimension stone, agricultural limestone (aglime), riprap, filler, and the manufacture of cement and lime. Throughout the Mississippian outcrop region, quarries are common, and many are associated with past highway construction. Large permanent quarries exist where special products are produced, such as cement or dimension stone, and in urban areas, where the demand for construction aggregate is high. Several underground operations have been converted to warehouse storage. In the past, lead and zinc were recovered from Mississippian Formations in the Tri-State district of southwestern Missouri and the adjoining areas of Kansas and Oklahoma, but this activity has ceased in the last 10 years.

Lead and zinc were mined extensively in the Tri-State district (Missouri-Kansas-Oklahoma) from the late 19th century to the early 1970's, when low prices and decreasing quality or grade of ores forced closing of the mining and milling operations. Production of significant amounts had ceased in the early 1950's in Missouri. Peak production, in 1916, was valued at $41,681,000 and consisted of 155,527 tons of recoverable zinc and 30,827 tons of recoverable lead (Wharton and others, 1969).

KINDERHOOKIAN SERIES

Limestone of the Chouteau Group is quarried primarily for use as aggregates in bituminous surfacing, roadstone (base and surfacing), aglime, and riprap. It is generally dolomitic and (or) cherty and does not always meet the specifications for concrete aggregate. In western Missouri, the region of greatest thickness, Chouteau carbonate rocks were quarried adjacent to railroads, and several very large quarries were developed before 1950. Where less dolomitic, as in southwestern Missouri, Kinderhookian limestones are quarried with the overlying Osagean strata.

OSAGEAN SERIES

Osagean limestones are used extensively throughout their outcrop area as construction aggregate and aglime and for cement and lime manufacture. The Burlington Limestone is quarried in southwestern Missouri (Springfield) for lime manufacture and in northeastern Missouri (Hannibal) for the manufacture of cement. Several chert-free zones in Osagean formations are quite widespread, and those near construction projects are quarried. As stated, Osagean strata were the principal source beds for lead and zinc ores in the Tri-State district that included southwestern Missouri.

MERAMECIAN SERIES

Meramecian limestones, like those of the Osagean Series, are extensively quarried throughout their outcrop area for use as construction aggregates and in cement and lime manufacture. The Salem limestone is mined in southeastern Missouri (at Ste. Genevieve) for lime, and the St. Louis limestone is quarried in the St. Louis region for cement. The Mississippi Lime Co. at Ste. Genevieve is the largest lime plant in the United States. The Warsaw Formation is mined and quarried in southwestern Missouri (Carthage) for dimension stone, known commercially as "Carthage marble."

Lead and zinc in the Tri-State region have been recovered from Meramecian limestones. Several large solution structures (sinkholes) in Meramecian Warsaw limestone were filled with lower Pennsylvanian shales and sandstones. Galena and sphalerite were mined from the wall rock of these sink structures; many of the excavations are more than 200 feet in both depth and width.

CHESTERIAN SERIES

Chesterian strata, which have a very small outcrop area in the State, contain only a few limestone quarries in southwestern Missouri (Hindsville Limestone) and limestone and sandstone quarries along the Mississippi River in southeastern Missouri. The limestone and sandstone have been used primarily for riprap.

PENNSYLVANIAN SYSTEM

Pennsylvanian strata are commercially important sources of coal, limestone, shale, and refractory clay. Several thick beds of shale in central and western Missouri have been mined extensively for use in the manufacture of structural clay products (brick, tile, lightweight aggregate). Coal from the Cherokee Group continues to be used as fuel in Missouri's steam-powered electric-generating plants. Several beds of limestone in the Marmaton, Kansas City, and Shawnee Groups are of sufficient thickness to be sources of construction aggregates, aglime, and cement manufacture. East-central Missouri is one of the major refractory clay (fireclay) producing areas in the United States.

DESMOINESIAN SERIES

Cherokee Group.—Most minable coal beds in Missouri are within the Cherokee Group. Recent emphasis on replacing petroleum-based fuels has led to increased study of coal resources of Missouri. Evaluation of Missouri coal fields is currently underway by the Missouri Department of Natural Resources, Division of Geology and Land Survey, to refine reserve estimates.

The major refractory clay resources are pre-Marmaton but whether they are equivalent to the Cherokee or are older (pre-Desmoinesian) is questionable.

Marmaton Group.—The principal resources of the Marmaton Group are limestone and coal. Several limestone beds, chiefly the Higginsville and Myrick Station, are quarried where they are of sufficient thickness. The principal coal beds that are mined locally are the Summit, Lexington, and Mulberry.

MISSOURIAN SERIES

Pleasanton Group.—Thick beds of shale in the Pleasanton Group in northwestern Missouri are a major source of raw material for the manufacture of structural clay products. The shale is quarried near Chillicothe for the manufacture of brick.

Kansas City Group.—The principal resource of Kansas City Group is limestone for construction aggregates and aglime. Several limestone members are quarried in the Kansas City region of northwestern Missouri (Bethany Falls, Winterset, Argentine). The Bethany Falls is quarried at Kansas City for cement manufacture. A large number of drive-in mines have been developed in the Bethany Falls limestone, and many of these have been converted to warehouse and office space.

Lansing and Pedee Groups.—A few limestone beds have been quarried in northwestern Missouri from Lansing (Spring Hill, Stoner) and Pedee (Iatan) strata, although the strata are primarily shale. Shale of the Weston Formation (Lansing Group) is quarried north of the Kansas City region for the manufacture of lightweight aggregate.

VIRGILIAN SERIES

Douglas and Shawnee Groups.—The Amazonia, Plattsmouth, and Ervine Creek limestones are the major source of construction aggregates and aglime in the extreme northwestern part of the State. The units are thin, and many quarries have been opened.

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The Mississippian and Pennsylvanian Systems in the United States—Arkansas

By BOYD R. HALEY, ERNEST E. GLICK, WILLIAM M. CAPLAN,
DREW F. HOLBROOK, *and* CHARLES G. STONE

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Arkansas

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—ARKANSAS

By Boyd R. Haley, Ernest E. Glick, William M. Caplan,[1] Drew F. Holbrook,[1] and Charles G. Stone [1]

ABSTRACT

Rocks of Carboniferous age are exposed in the northwestern third of Arkansas; more than two-thirds of the exposed rocks are of Pennsylvanian age. The contact between Devonian and Carboniferous rocks in most of Arkansas must be paleontologically determined. The area that is now Arkansas received shallow- to deep-water marine deposits until the start of the Ouachita orogeny (Middle Pennsylvanian). Most of the Middle and all of the Upper Pennsylvanian rocks have been removed by erosion. Sill-like lenses of soapstone-serpentine of possible Late Pennsylvanian age are found in Middle Ordovician rocks. Beds of volcanic tuff are present in Upper Mississippian rocks. Hydrothermal quartz veins of Late Pennsylvanian age are abundant in the Ouachita Mountains. Commercial quantities of coal, natural gas, antimony, barium, manganese, mercury, zinc, lead, dimension stone, limestone, slate, and tripoli are present in Carboniferous rocks.

INTRODUCTION

The outcrop area of Carboniferous rocks shown in figure 1 contains rocks of Devonian and Mississippian age and of Early and Middle Pennsylvanian age. The contact between Devonian and Mississippian rocks, as paleontologically determined, is in a rock unit that cannot be lithologically separated in most of Arkansas; hence, Devonian rocks are included in figure 1.

Rocks of Carboniferous age are exposed in the northwestern third of Arkansas; more than two-thirds of the exposed rocks are of Pennsylvanian age. Carboniferous rocks are known to be present to the east under younger rocks in the Mississippi Embayment and to the south under the younger rocks in the Gulf Coastal Plain (fig. 1).

The unit shown in figure 2 as of Morrowan age contains fossils reported to be of youngest Mississippian age in its lowermost parts in areas east of long 92°45′ W. Changing the lithologic and thus the geographic position of the contact between rocks of Mississippian and Pennsylvanian age as shown in figure 2 or on the "Geologic Map of Arkansas" (Haley and others, 1976) did not and does not seem feasible in view of the available evidence.

The area that is now Arkansas received shallow- to deep-water marine and continental deposits until the start of the Ouachita orogeny (Middle Pennsylvanian). The rocks were then folded, faulted, and severely eroded until sediments deposited during Cretaceous and Eocene times covered them to about their present extent. Even though the presently exposed Carboniferous rocks have been almost continuously eroded since late Paleozoic time, they are remarkably fresh where exposed along streams, lakes, roads, and in excavations.

GEOLOGIC INVESTIGATIONS

Early explorers and scientists made references to the rocks in the area that is now Arkansas as early as 1817. However, David D. Owen during the years 1857 to 1860 was the first to publish reports of reconnaissance studies pertaining in part to the rocks of Carboniferous age. Since that time, so many people have worked on the rocks of this age that space does not allow recognition of their reports.

Table 1 lists the names and the publications (concerning Arkansas) of those people who have made major contributions to geologic knowledge about the Carboniferous rocks of the State. These geologists have described rock units and have established nomenclature for these units; most of them have provided geologic maps that have been or will be the basic standard for all subsequent geologic investigations. This table serves as a list of references for this report.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomen-

[1] Arkansas Geological Commission, Little Rock, Ark. 72204.

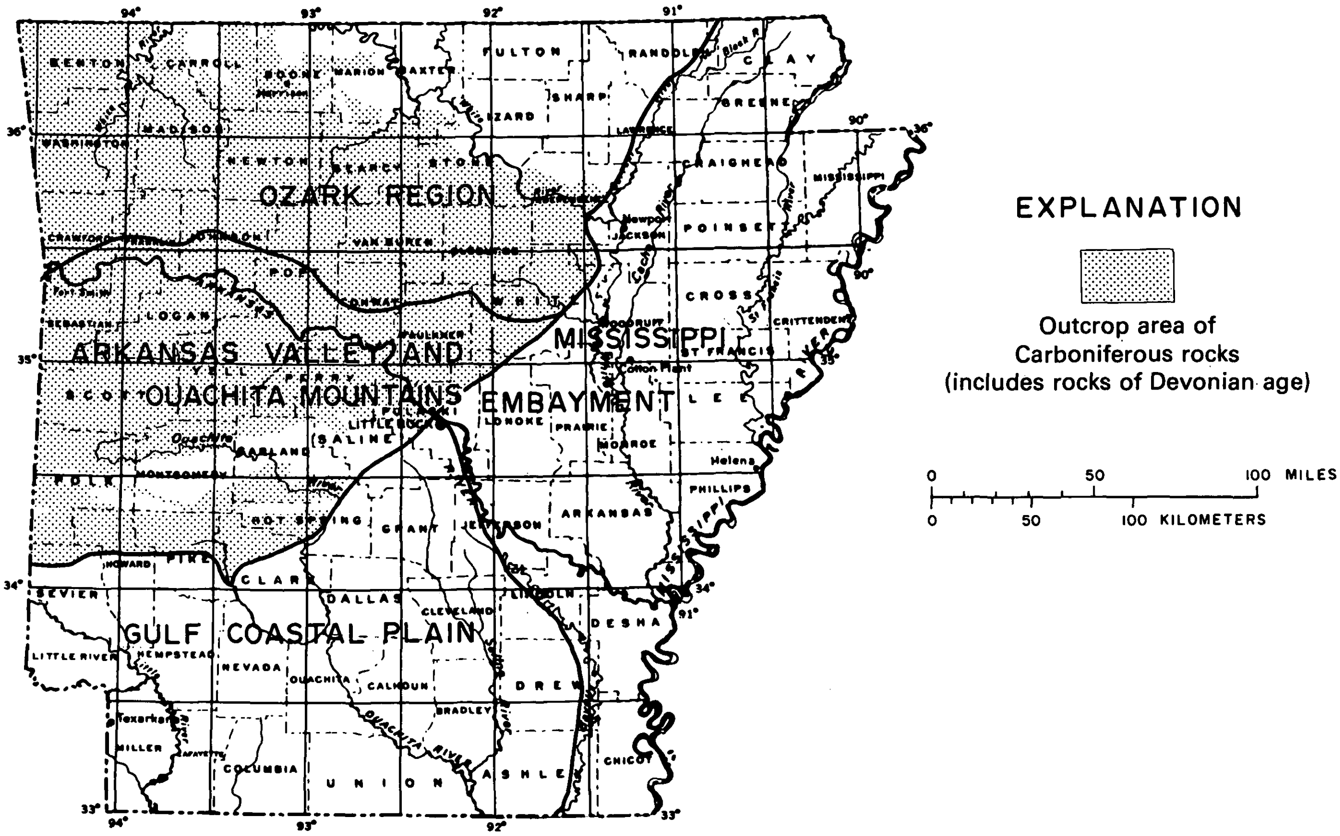

FIGURE 1.—Index map of Arkansas showing physiographic regions and outcrop of Carboniferous rocks.

TABLE 1.—*Major geologic contributions pertaining to the Carboniferous rocks of Arkansas*

Reference	Area of investigation
Adams, G. I., and Ulrich, E. O., 1905, Description of the Fayetteville quadrangle [Arkansas-Missouri]: U.S. Geol. Survey Geol. Atlas, Folio 119, 6 p.	Ozark region.
Collier, A. J., 1907, The Arkansas coal field. With reports on the paleontology by David White and G. H. Girty: U.S. Geol. Survey Bull. 326, 158 p.	Arkansas Valley.
Glick, E. E. (with Haley, B. R., and others), 1976, Geologic map of Arkansas (see Haley and others, 1976)	Ozark region.
Griswold, L. S., 1892, Whetstones and the novaculites of Arkansas: Arkansas Geol. Survey Ann. Rept., 1890, v. 3, 443 p.	Ouachita Mountains.
Haley, B. R., and others, 1976, Geologic map of Arkansas: Reston, Va., U.S. Geol. Survey, scale 1:500,000.	Ozark region, Arkansas Valley, and Ouachita Mountains.
Henbest, L. G., 1953, Morrow Group and lower Atoka Formation of Arkansas: Am. Assoc. Petroleum Geologists Bull. v. 37, no. 8, p. 1935–1953.	Ozark region.
Hendricks, T. A., and Parks, B. C., 1950, Geology of the Fort Smith district, Arkansas: U.S. Geol. Survey Prof. Paper 221–E, p. 67–94.	Arkansas Valley.
Hopkins: T. C., 1893, Marbles and other limestones: Arkansas Geol. Survey Ann. Rept. 1890, v. 4, 443 p.	Ozark region.
McKnight, E. T., 1935, Zinc and lead deposits of northern Arkansas: U.S. Geol. Survey Bull. 853, 311 p.	Do.
Miser, H. D., 1922, Deposits of manganese ore in the Batesville district, Arkansas: U.S. Geol. Survey Bull. 734, 273 p.	Ozark region.
Miser, H. D., 1929, Geologic map of Arkansas: Little Rock, Arkansas Geol. Survey.	Arkansas Valley and Ouachita Mountains.
Miser, H. D., and Purdue, A. H., 1929, Geology of the De Queen and Caddo Gap quadrangles, Arkansas: U.S. Geol. Survey Bull. 808, 195 p.	Ouachita Mountains.
Purdue, A. H., 1907, Description of the Winslow quadrangle [Arkansas-Indian Territory]: U.S. Geol. Survey Geol. Atlas, Folio 154, 6 p.	Ozark region.
Purdue, A. H., and Miser, H. D., 1916, Description of the Eureka Springs and Harrison quadrangles [Arkansas-Missouri]: U. S. Geol. Survey Geol. Atlas, Folio 202, 22 p.	Do.
Purdue, A. H., and Miser, H. D., 1923, Description of the Hot Springs district [Arkansas]: U.S. Geol. Survey Geol. Atlas, Folio 215, 12 p.	Ouachita Mountains.
Reinemund, J. A., and Danilchik, Walter, 1957, Preliminary geologic map of the Waldron quadrangle and adjacent areas, Scott County, Arkansas: U.S. Geol. Survey Oil and Gas Inv. Map, OM–192, scale 1:48,000.	Arkansas Valley and Ouachita Mountains.
Stone, C. G., (with Haley, B. R., and others), 1976, Geologic map of Arkansas (*see* Haley and others, 1976)	Do.

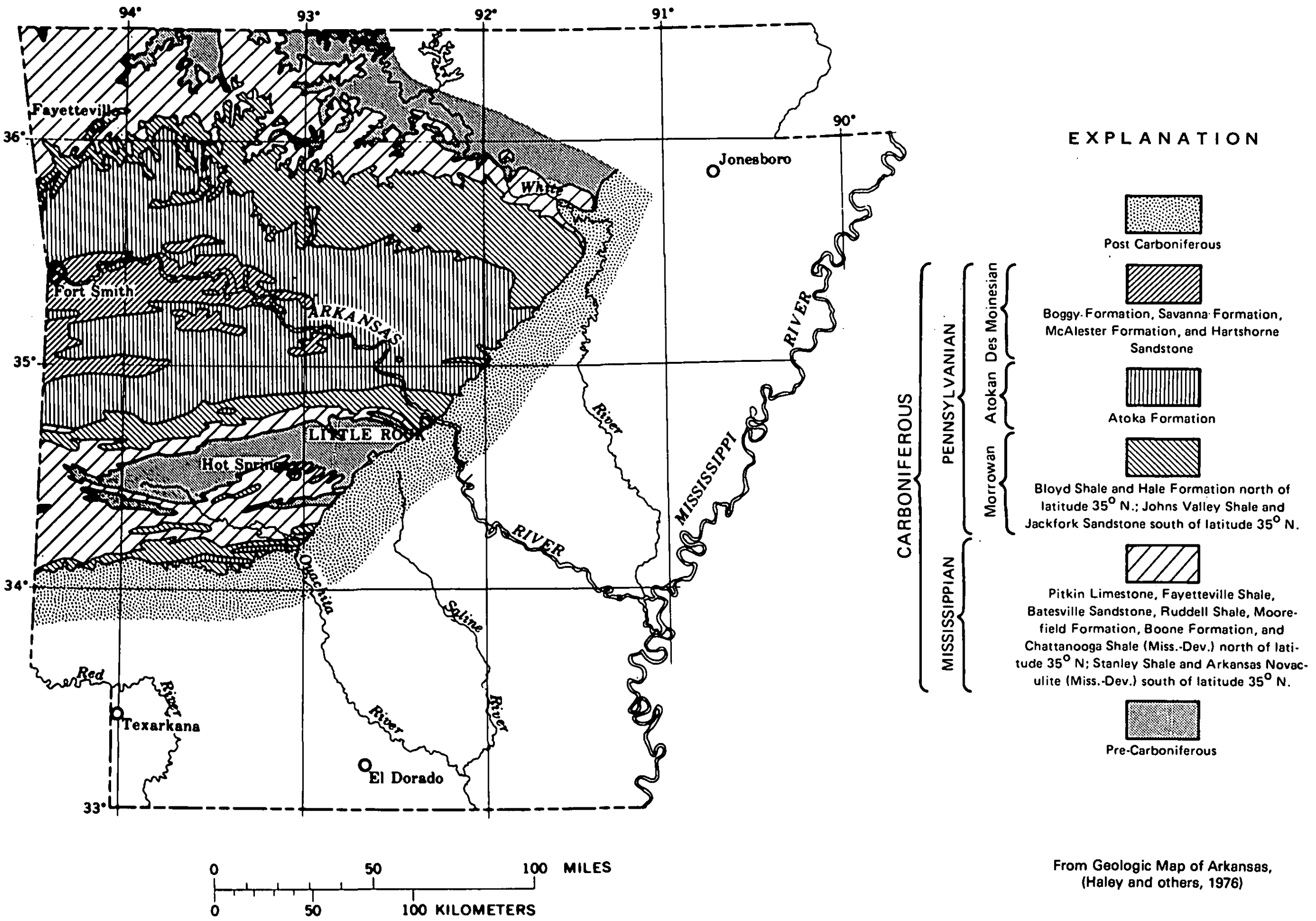

FIGURE 2.—Outcrop map of Carboniferous rocks in Arkansas.

clature used here conforms with the current usage of the Arkansas Geological Commission.

GEOLOGIC SETTING

By ERNEST E. GLICK

The paleontological base of Mississippian rocks is in the Middle Division of the Arkansas Novaculite in the Ouachita Mountains and in the Chattanooga Shale in the Ozark region. North of the limit of the Chattanooga Shale, a thin sandstone at the base of the Mississippian sequence rests on formations of Ordovician, Silurian, or Devonian age. This unit of sandstone, generally less than 0.3 m thick, is thought to be a lag sand or marine regolith that accumulated during late Devonian time. The presence of Devonian conodonts in its lower part and Mississippian conodonts in its upper part in some areas indicates that it was partly reworked during Early Mississippian time.

Deposition in Arkansas was relatively continuous during early and part of middle Carboniferous time, regional breaks being suggestive of sea-level changes until the advent of the Ouachita orogeny in Middle Pennsylvanian time. Growth faults had the greatest magnitude along a northeast-trending zone in west-central Arkansas. The displacement across this zone of faults was large enough to influence the rate and type of deposition, and generally these faults mark the boundary between the area of shelf deposition and the area of trough deposition. The growth-fault system in northeastern Arkansas is thought to be on the western side of the graben formed by the Reelfoot rift system. The fault system in west-central Arkansas may be on the northern side of a rift system that has an extent and existence still subject to conjecture. The time at which the Ouachita orogeny began is difficult to establish, but some of the rocks of Des Moinesian age have sedimentary and structural features that may have resulted from the initial compressional pulse of the orogeny. The orogeny was active by Late Pennsylvanian time and lasted until Late Permian time. All of Arkansas was uplifted, the core area of the Ouachita Mountains

Age				Ozark Mountains and Arkansas Valley		Maximum thickness in meters	Ouachita Mountains		Maximum thickness in meters
Carboniferous	Pennsylvanian	Middle	Des Moinesian	Missing			Missing		
				Boggy Formation		30			
				Savanna Formation	Paris Coal Bed Charleston Coal Bed	595			
				McAlester Formation	Upper Hartshorne Coal Bed Lower Hartshorne Coal Bed	610			
				Hartshorne Sandstone		105			
			unconformity	unconformity					
			Atokan	Atoka Formation		3,952	Atoka Formation	Upper	1,525
								Middle	1,675
								Lower	5,485
		Early	Morrowan	Bloyd Shale	Trace Creek Member Kessler Limestone Member Dye Shale Member Woolsey Member Brentwood Limestone Member	191	Johns Valley Shale		455
				Hale Formation	Prairie Grove Member	94	Jackfork Sandstone		1,830
					unconformity				
					Cane Hill Member	219			
				unconformity					

System	Series	Subseries	Stage	Ozark Mountains and Arkansas Valley	Maximum thickness in meters	Ouachita Mountains	Maximum thickness in meters
				unconformity			
Carboniferous	Mississippian	Late	Chesterian	Pitkin Limestone	141	Stanley Shale	2,585
				Fayetteville Shale; Wedington Sandstone Member	227		
				Batesville Sandstone; Hindsville Limestone Member	52		
				unconformity			
				Ruddell Shale	30		
			Meramecian	Moorefield Formation	60	Hatton Tuff Lentil; Hot Springs Sandstone Member	
				Boone Formation; Short Creek Oolite Member	130		
		Early	Osagean	Boone Formation		Upper Division Arkansas Novaculite	38
			Kinderhookian	St. Joe Limestone Member			
				unconformity		Middle Division Arkansas Novaculite	160
Devonian				Chattanooga Shale	20		
				unconformity			
				Clifty Limestone and Penters Chert	24	Lower Division Arkansas Novaculite	125

FIGURE 3.—Correlation chart of Carboniferous rocks in Arkansas.

being elevated more than 9,100 m along the anticlinorium. Compressive features die out northward from the recumbent folds and low- and high-angle thrust faults in the core area to the gentle folds in the Ozark region. During the orogeny, erosion almost kept up with the uplift so that few, if any, high mountains were formed.

After the orogeny and before the transgression of the Cretaceous seas, the rocks in the southern and eastern part of Arkansas were probably eroded to a plane; however, cuestas, hogbacks, and synclinal mountains were formed and continue to be positive features on the landscape. Erosion by the transgressive Cretaceous sea further beveled the plane on top of the Carboniferous rocks and then these rocks were covered by sediments of Cretaceous age. A thin strip of Carboniferous rocks, extending northeastward from Little Rock was later covered by sediments of Tertiary age.

LITHOSTRATIGRAPHY

The stratigraphic divisions of the Carboniferous rocks in Arkansas are listed and correlated in figure 3. The unconformities listed between some of the rock units in northern Arkansas represent only the major fluctuations of sea level. Many unconformities thought to represent lesser fluctuations are not shown.

In the northwestern part of the Ozark region, the contact between the Mississippian and Pennsylvanian rocks has been placed at the base of a conglomerate at the base of the Cane Hill Member of the Hale Formation. Lithologically and paleontologically, the contact is considered to be unconformable. In the northeastern part of the Ozark region, where the Cane Hill is much thicker, the contact has been placed at the base of the conglomerate and has been considered to be unconformable. However, fossils correlatable to only the youngest of European Mississippian forms have been reported from above the basal conglomerate. On the basis of the age of these fossils, some authors have divided the present Cane Hill Member into the Cane Hill Formation of Pennsylvanian age and the underlying Imo Formation of Mississippian age. The proposed Imo Formation, however, is not a mappable or a recognizable lithologic unit, and its top can be established only by the existing upper limit of diagnostic fossils of Mississippian age. Thus, in the northeastern part of the Ozark region, the position and the nature of the contact between Mississippian and Pennsylvanian rocks has yet to be determined.

In the Ouachita Mountains, the contact between Mississippian and Pennsylvanian rocks is placed at the base of the Jackfork Sandstone and is considered to be conformable. The exact paleontological boundary has not been determined, and some of the rocks in the lowermost Jackfork may be of Mississippian age.

In the Ozark region and most of the Arkansas Valley, the Mississippian rocks consist of shale, limestone, cherty limestone, and a lesser amount of sandstone. All are thought to have been deposited in a shallow-water marine environment, only the Batesville and Wedington Sandstones being deposited nearshore.

In the Ouachita Mountains, the Mississippian rocks consist mostly of shale, some graywacke, and subordinate amounts of sandstone and novaculite. All are thought to have been deposited in a deep-water marine environment. The change from shallow-water to deep-water deposition in west-central Arkansas is thought to have been abrupt across a growth fault system shown as the Johns Valley fault system in figure 4.

In the Ozark region and the western part of the Arkansas Valley, rocks of Morrowan age consist of sandstone and shale in the Hale Formation and limestone and shale in the Bloyd Shale. All are thought to have been deposited in a shallow-water marine environment except the Woolsey Member of the Bloyd Shale, which is continental in origin.

In the Ouachita Mountains, the Morrowan rocks consist primarily of sandstone and shale in the Jackfork Sandstone and shale and subordinate amounts of sandstone and conglomerate in the Johns Valley Shale. All are thought to have been deposited in deep water.

Rocks of Atokan age consist of shale and lesser amounts of sandstone and siltstone, and a few coal beds in the upper part of the formation. Nearly all of the Atoka was deposited in an environment that was essentially shallow-water marine in the Ozark region and deep-water marine in the Ouachita Mountains. The southward transition of shallow- to deep-water deposition was gradational and persisted during all but the later part of Atokan deposition (fig. 4).

Rocks of Des Moinesian age consist of shale and subordinate amounts of sandstone and siltstone and a few beds of limestone and coal. Most of the rocks were deposited in a shallow-water marine environment; the rest were deposited above sea level.

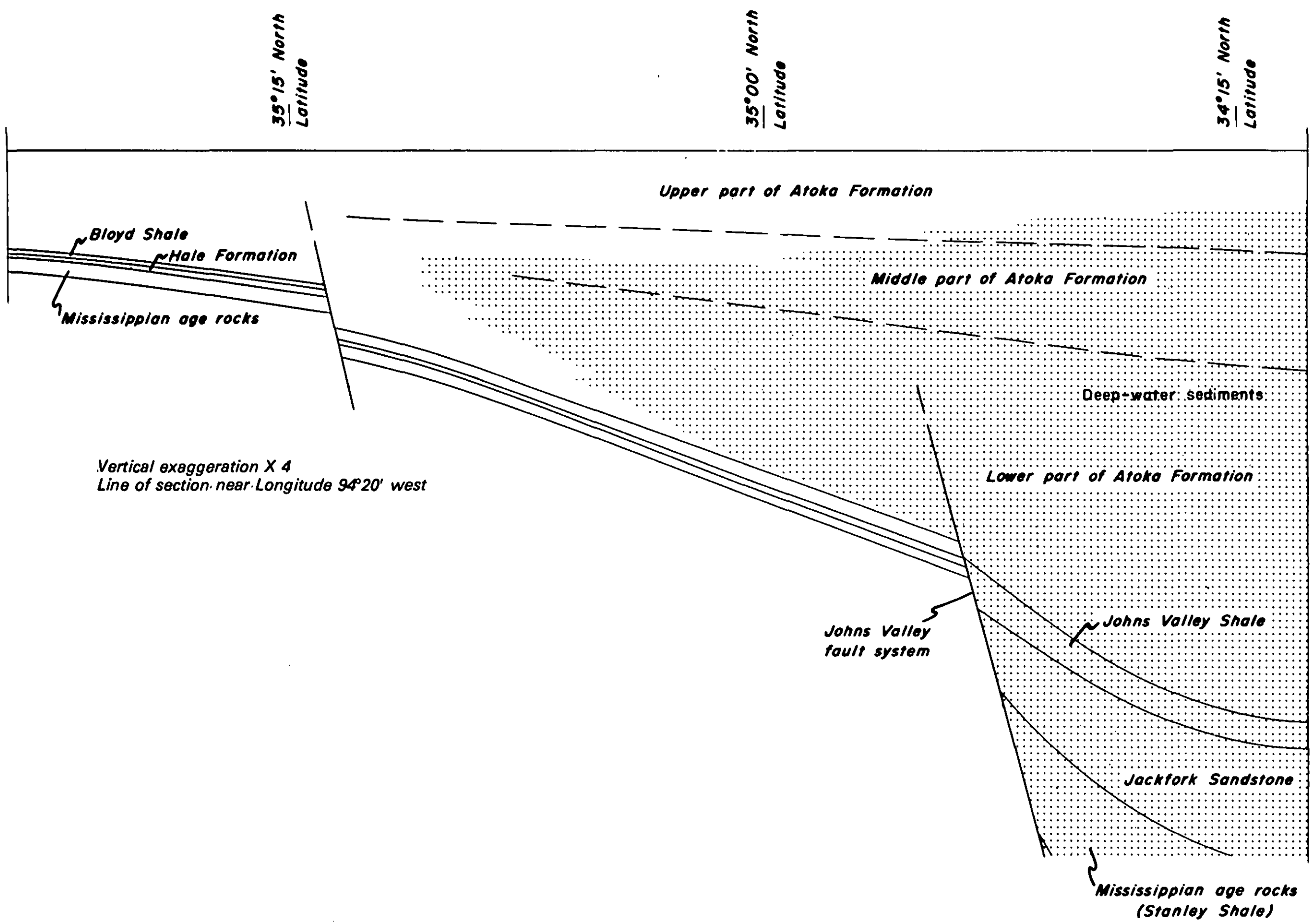

FIGURE 4.—Diagrammatic section of Atokan and older Carboniferous rocks in western Arkansas.

IGNEOUS ROCKS, METAMORPHIC ROCKS, AND VEINS

By CHARLES G. STONE

The only known igneous rocks of possible Late Pennsylvanian age in Arkansas are some soapstone-serpentine sill-like lenses in the Womble and Bigfork Formations about 15 miles west of Little Rock. The serpentine is the autometamorphosed product of an ultrabasic (peridotite) intrusive that was injected into these Ordovician rocks prior to folding processes of the Ouachita orogeny (mostly late Carboniferous).

Beds of volcanic tuff are present in the lower part of the Stanley Shale. They are thickest and most numerous near Hatton in Polk County. In this vicinity, beds of the Hatton Tuff are present in the lower 120 m of the Stanley; some of the individual beds are as much as 30 m thick. A typical bed of tuff grades upward from a coarse-grained crystalline tuff to a fine-grained vitric tuff, and some of the grading may represent reworking by turbidity currents. Potash feldspar, sodic plagioclase, perthite, and quartz are prominent minerals in the crystal tuff and ash, and shards are more common in the vitric tuff. The number and thickness of the beds and the grain size of the tuff decrease north and east of Hatton; therefore, the volcanic source of the tuff is thought to have been to the south or southwest. Radiometric ages on the tuff range from 293 ± 15 m.y. to 322 ± 26 m.y. before present. The older date conforms with the early Chesterian age of the conodont assemblages collected from the interbedded shale.

Beds of tuff composed of sodic plagioclase, volcanic rock fragments, and other materials are present in the uppermost part of the Stanley in Polk, Montgomery, and Garland Counties. Fossils collected from rocks near the tuff beds are of late Chesterian age.

Beds of bentonite (hydromica-montmorillonite), probably derived from volcanic ash falls, are indi-

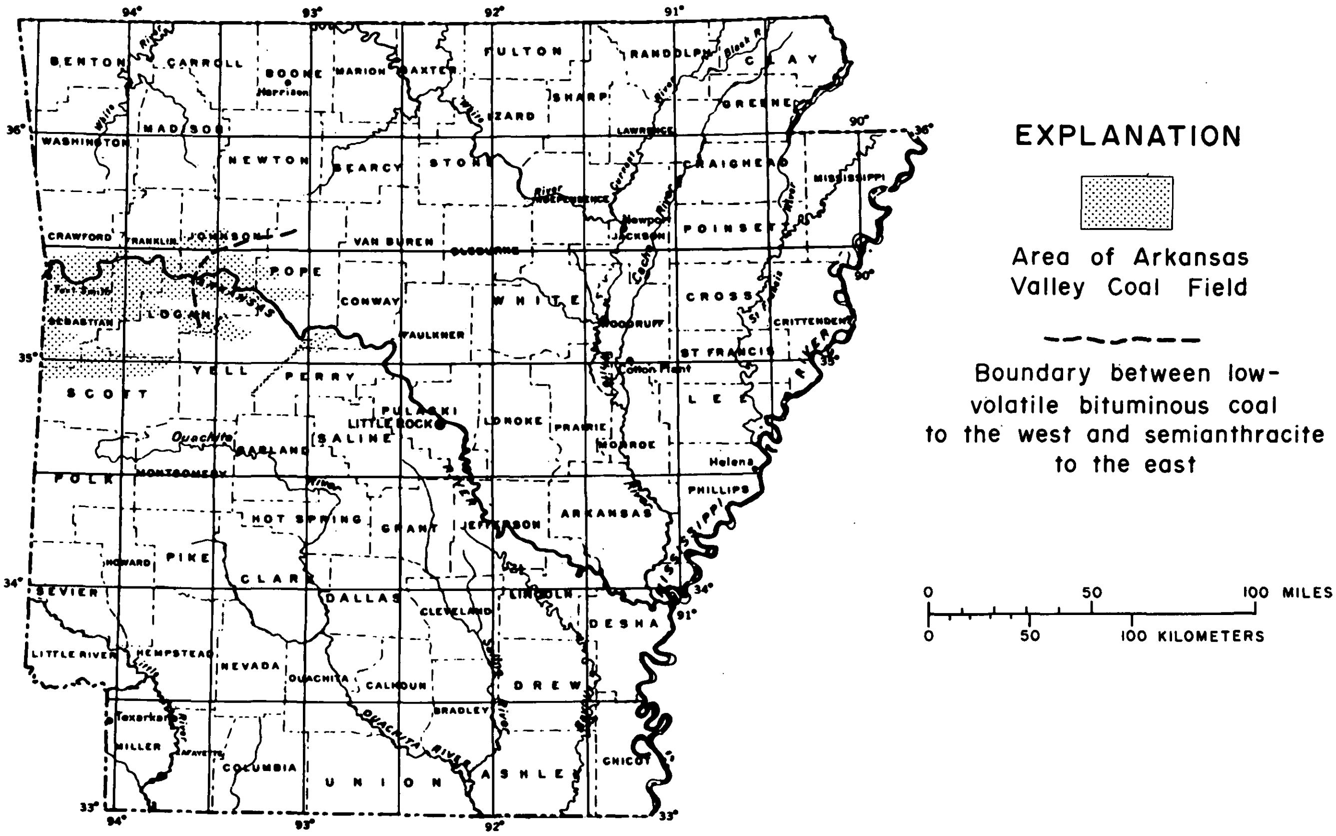

FIGURE 5.—Arkansas Valley coal-field map.

cated by studies of well cuttings from the Bloyd Shale and the lower part of the Atoka Formation in the Arkansas Valley.

A belt of low-rank regional metamorphism (chlorite stage) has affected the Mississippian and Lower Pennsylvanian rocks of the Ouachita Mountains, the area of highest intensity being centered near Little Rock. Hydrothermal quartz veins often containing minor quantities of dickite, adularia feldspar, and chlorite are common in these rocks. Most of the regional metamorphism and the emplacement of the quartz veins are thought to be related to late deformation during the Ouachita orogeny.

ECONOMIC PRODUCTS

COAL

Coal beds are present in the Bloyd Shale in northwestern Arkansas and in the upper part of the Atoka Formation in west-central Arkansas, but nearly all commercial production of coal has been from coal beds in the McAlester and Savanna Formations. The coal in the Arkansas Valley coal field (fig. 5) ranges in rank from low-volatile bituminous to semianthracite. The Baldwin coal bed in the Bloyd Shale is less than 0.4 m thick and has been mined only to supply local demand in Washington County.

The coal beds in the Atoka Formation are less than 0.6 m thick and were mined to supply local demand; however, some of the coal is now being used in the manufacture of dry-cell batteries and as an additive to charcoal briquets.

The Lower Hartshorne coal bed, near the base of the McAlester Formation, ranges in thickness from a featheredge to 1.8 m. It is the most extensive, most mined, and most economically important coal bed in Arkansas. Most of the mined coal was used for steam-generating plants, home heating, and the zinc industry, but present production is used almost exclusively by the metallurgical industry.

The Upper Hartshorne coal bed, about 18.2 m above the base of the McAlester Formation, ranges in thickness from a featheredge to 0.7 m in the southwest part of the Arkansas Valley coal field. It has not been mined.

The Charleston coal bed near the base of the Savanna Formation is less than 0.8 m thick, and in recent years the mined coal has been used as a metallurgical coal.

The Paris coal bed, near the middle of the Savanna Formation, is less than 0.8 m thick. Most of the mined coal was used for steam generating and home heat; however, it could be used as a metallurgical coal.

The distribution of the estimated 2.2 billion short tons of coal in the Arkansas Valley coal field is shown in figure 6. The total production of coal in short tons through 1975 is as follows: Franklin County 13,509,527; Johnson County 17,992,813; Logan County 9,981,039; Pope County 9,313,459; Scott County 958,057; Sebastian County 56,827,873; other counties or small mines 1,114,992; and a State total of 103,697,760.

OIL AND NATURAL GAS

By William M. Caplan

Commercial quantities of oil have not been produced from Carboniferous rocks in northern Arkansas. Oil is present in sandstone and shale of Pennsylvanian age and in concretions and black shale of Mississippian age in northern Arkansas. A solid bitumen deposit is present in sandstone of the Jackfork Sandstone in Scott County. Natural gas has been produced in commercial quantities from the Carboniferous rocks of northern Arkansas since 1902. With few exceptions, the gas is low in sulfur, high in methane, low in nitrogen, and has a heating value of about 1,000 Btu per cu ft (0.028 m^3).

The first well to produce commercially was completed in Sebastian County from the Atoka Formation, and this formation continues to be the most prolific source of natural gas of Carboniferous age. (see table 2).

From 1902 to 1944, 19 gas fields were discovered in the Atoka Formation. In 1944, gas was discovered in the Bloyd Shale in Johnson County, and in 1949, gas was discovered in the Hale Formation in Franklin County.

The first commercial production of gas of Mississippian age was from the Wedington Sandstone, where it was penetrated by a well drilling Franklin County. Commercial quantities of natural gas were first produced in 1962 from the Boone Formation in Crawford County, and additional commercial-quality gas was discovered in 1974 in Johnson County. The Boone production wells also produce gas from the Atokan or Morrowan rocks or both. The gas from the Pennsylvanian and pre-Mississippian rocks in northern Arkansas is similar enough to be commingled in the wells, but the gas from the Boone in some areas is an exception because of a relatively high hydrogen sulfide content.

Commercial deposits of natural gas of Carboniferous age in Arkansas were thought originally to be related entirely to structure, but the importance of stratigraphic and stratigraphic-structural traps as geologic settings has become increasingly significant as more is learned about the region. Exploratory wells drilled on structural prospects alone may miss the buried stream channels and deltaic deposits containing natural gas in trends normal or near normal to the strike of the structural feature.

As of January 1, 1977, the 10 counties of northern Arkansas have 60 gas fields, containing 1,223 wells. Natural gas from rocks of Carboniferous age is or could be produced in 58 of these fields irrespective of any deeper production there.

Cumulative gas production to January 1, 1977, in northern Arkansas is 1,801,722,499 Mcf ($50,989 \times 10^6 m^3$). Remaining reserves are estimated to be 1,510,523,000 Mcf ($42,748 \times 10^6 m^3$). Both figures pertain mainly to natural gas in the rocks of Carboniferous age.

Areas underlain by commercial quantities of natural gas in the Atoka Formation and the Morrow Series are shown in figure 7. Areas in which gas is produced from rocks of Mississippian age are not specifically shown in figure 7.

MINERALS

By Drew F. Holbrook

METALLIC MINERALS

ANTIMONY

The antimony district is a belt 4 km wide and 17.6 km long extending across northern Sevier County in southwestern Arkansas. The antimony deposits are present in the tightly folded rocks of the Stanley Formation; the fissure veins tend to be parallel to bedding. The ore bodies are small and lenticular and are known to have a maximum thickness of 1.3 m, a maximum length of 51.7 m, and a maximum width of 22.8 m.

The ore bodies are mainly stibnite-bearing quartz veins in which crystals and crystal masses of stibnite are present in comb quartz. Other primary minerals are native antimony, chalcopyrite, galena, pyrite, sphalerite, jamesonite and zinckenite. Only galena, sphalerite, and pyrite are present throughout the district.

Secondary minerals include cervantite, which is the most abundant, stibiconite, bindheimite, anglesite, cerussite, azurite, malachite, and smithsonite. The gangue minerals are quartz, ankerite, calcite,

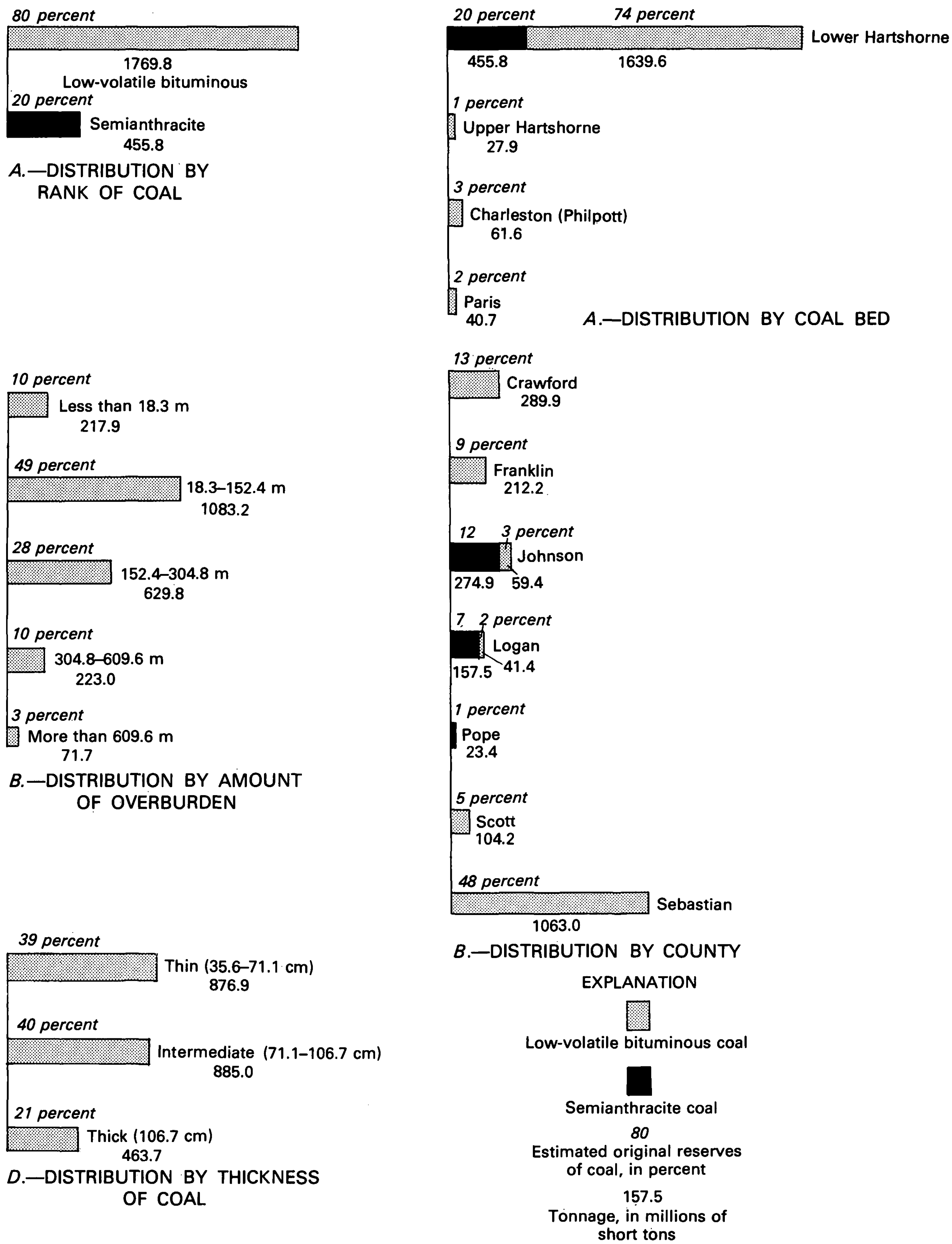

FIGURE 6.—Distribution of estimated original reserves of low-volatile bituminous coal and semianthracite in Arkansas.

TABLE 2.—*Natural-gas wells of economic and historic significance in the Carboniferous rocks of Arkansas*

Company and well	Location (Sec., T., R.)	County	Date completed	Total depth (ft (m))	Production depth (ft (m))	Pay zone	Initial potential in mcfgd ((m³) per day)	Remarks
Choctaw Oil and Gas Co. No. 2 Duncan.	1, 4N., 31W.	Sebastian	Mar. 1902	1,125 (343)	863–873 (263–266)	Atoka	550 (15,565)	Discovery well of Mansfield field and first producer of natural gas in Arkansas.
					1,125 (343)	----do ----		
Arkansas Louisiana Gas Co. No. 1 S. M. Hudson.	15, 10N., 24W.	Johnson	Dec. 1944	6,135 (1,870)	3,402–3,650 (1,037–1,085)	Kessler and Brentwood.	3,500 (99,050)	Drilled in Clarksville field; the first commercial Bloyd producer in Arkansas.
Arkansas Louisiana Gas Co. No. 1 Ralph S. Barton.	27, 9N., 28W.	Franklin	Sept. 1949	6,650 (2,027)	4,850–4,930 (1,478–1,503)	Hale	7,000 (198,100)	Discovery well of Cecil field; the first commercial Hale producer in Arkansas.
Arkansas Western Gas Co. No. 1 F. A. Parsley.	15, 10N., 28W.	----do ----	Oct. 1951	3,896 (1,188)	3,348–3,367 (1,020–1,026)	Morrow	6,000 (169,800)	The first commercial producer from the Wedington or pre-Pennsylvanian rocks in Arkansas.
					3,782–3,814 (1,153–1,163)	Wedington	2,000 (56,000)	
Carter Oil Co. No. 1 L. B. Chronister.	3, 8N., 19W.	Pope	Oct. 1956	5,211 (1,588)	3,917–4,030 (1,194–1,228)	Atoka	(71,800) (2,031,940)	Drilled in Moreland field; it had the largest initial potential of any well drilled in north Arkansas.
					4,469–4,514 (1,362–1,376)	Hale	6,700 (189,610)	
Arkansas Western Gas Co. No. 1 A. Goad.	20, 14N., 29W.	Washington	April 1959	650 (198)	565 (172)	Wedington	460 (13,018)	Discovery well of Brentwood field, but was not completed as a producer. Drilling for additional reserves started in 1977.
Beard Oil Co. No. 1 R. A. Evans.	32, 9N., 11W.	Cleburne	July 1959	3,868 (1,179)	1,956–1,998 (596–609)	Atoka	430 (12,169)	Discovery well of Quitman field; established the easternmost commercial producer in the Arkansas Valley. Field was abandoned in 1964.
Stephens Production Co. No. 1 D. L. Fontaine	32, 9N., 30W.	Crawford	June 1962	6,161 (1,878)	5,755–5,777 (1,754–1,761)	Boone	2,450 (69,335)	Drilled in Kibler field; the first commercial Boone producer in Arkansas.
					5,922–5,952 (1,805–1,815)	Penters	7,500 (212,250)	
Arkansas Western Gas. Co. No. 1 Federal ES 5262.	12, 11N., 25W.	Johnson	Dec. 1973	3,543 (1,080)	1,815–1,850 (553–564)	Hale	6,000 (169,800)	Drilled in Batson field. This well opened the first field that produced Boone gas commercially from more than one well.
					2,578–2,622 (786–799)	Boone	7,600 (215,080)	
					2,654–2,686 (809–819)	----do ----		

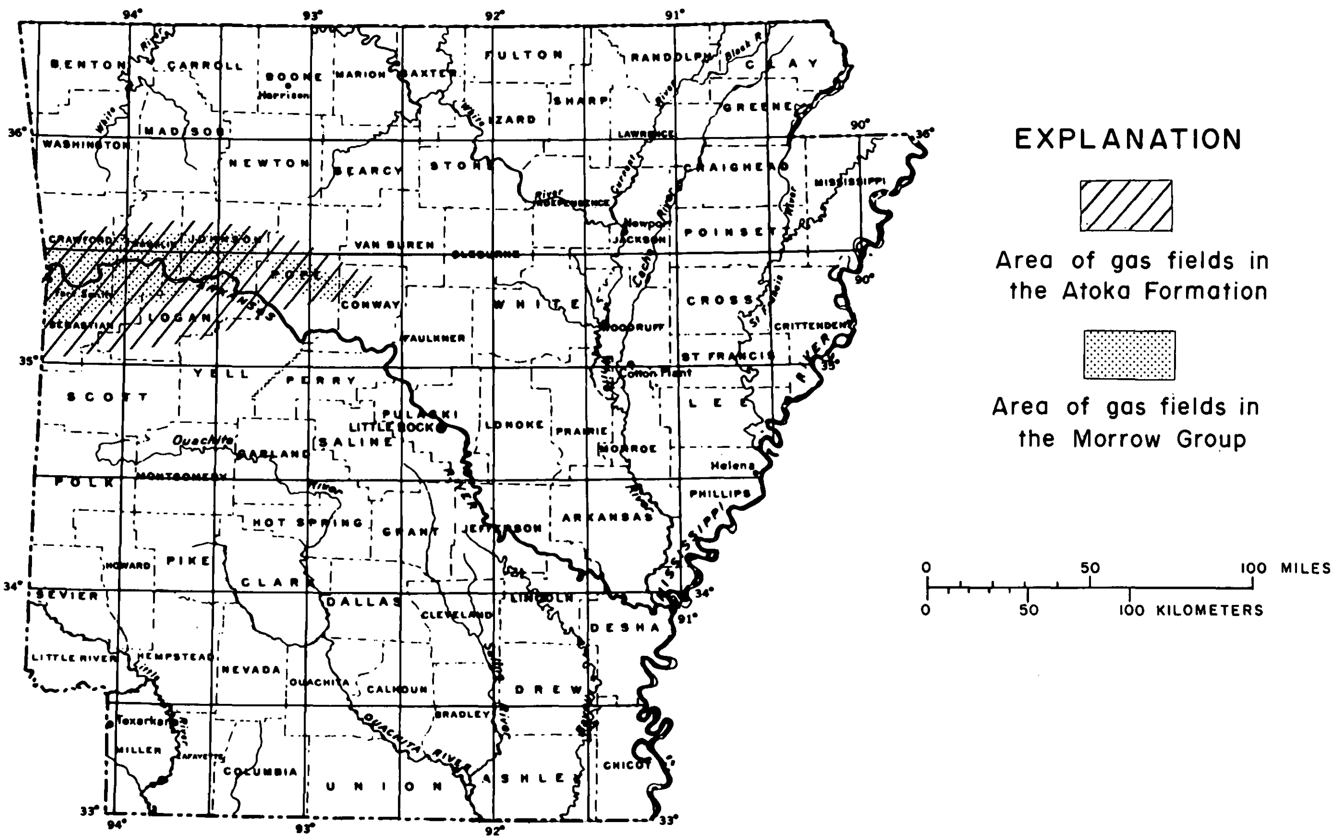

FIGURE 7.—Gas-field map of Arkansas.

siderite, chlorite, and dickite. Stibnite and the other primary minerals were deposited from hypogene solutions that moved along thrust-fault planes or subparallel fractures. The antimony mineralization probably took place after most of the quartz veins had been emplaced.

Production of antimony during the period 1873 to 1947 was sporadic and generally coincided with exceptionally high market prices. The total reported amount of antimony produced during the 70 years is only 589 short tons.

BARIUM

Barite (barium sulfate) has been produced from the Magnet Cove district in Hot Spring County and from the Pigeon Roost Mountain and Fancy Hill districts in Montgomery County. The barite deposits are in the lowermost part of the Stanley Formation. Generally the base of the barite deposit is within 6 m of the top of the Arkansas Novaculite. Most of the barite deposits are less than 30.4 m thick, but a few are as thick as 91.2 m.

The individual barite deposits contain several types of ore, which, in order of abundance, are as follows: (1) gray to dark-gray finely crystalline ore; (2) gray granular to dense ore; (3) nodular ore, as much as 5 cm in diameter; and (4) the least common type, a dark-gray to black granular ore.

Barite has been continuously produced in Arkansas since 1944. Through 1976, ten million short tons has been produced, almost all of which came from the Magnet Cove district. All the barite produced is used as a weighting agent in drilling mud.

MANGANESE

Manganese is present in Carboniferous rocks in Garland, Hot Springs, Saline, and Pulaski Counties; however, the larger deposits are in Montgomery and Polk Counties in an east-trending belt about 11.2 km wide and 56 km long. The manganese deposits are present in the Lower, Middle, and Upper Divisions of the Arkansas Novaculite. The ore bodies are small, the largest estimated to contain 50,000 short tons. The manganese ore is present as nodules, pockets, and short irregular veins ranging in thickness from a few centimeters to 1.2 m.

The ore bodies consist mainly of psilomelane, pyrolusite, manganite, and wad; two or more of these minerals are intimately mixed in the same deposit. Other minerals associated with the manganese minerals are limonite, goethite, native copper, cuprite, chrysocolla, turquoise, malachite, pyrite, lithiophorite, wavellite, variscite, dufrenite, barite, and quartz.

Manganese is present in the St. Joe Member of the Boone Formation in a few places in Izard, Searcy, and Stone Counties. Manganese was mined intermittently in western Arkansas from 1885 to 1959; production was mainly during a few years of high market prices and totaled only 6,000 short tons.

MERCURY

The mercury district is a belt 9.6 km wide and 48 km long extending eastward from Howard County to Clark County in southwestern Arkansas. The mercury deposits are about equally divided in the sandstone of the uppermost Stanley Formation and the lowermost Jackfork Sandstone. Most of the ore bodies are small, the largest being 30.4 m long, 9.1 m wide, and 36.5 m deep. Cinnabar is present as a finely crystalline ore coating on fracture surfaces, coarsely crystalline ore filling larger fractures, and sparingly, finely crystalline ore filling pore spaces in the sandstone. Other primary minerals are dickite, quartz, pyrite, siderite, stibnite, barite, and calcite. Secondary mercury minerals and iron oxides are present near the surface.

Active mining in the mercury district took place mostly during the years 1931 to 1946 and has been only intermittent since then. The total amount of mercury produced is estimated to be 12,500 flasks (946,250 lb).

VANADIUM

Vanadium ore is in local concentrations within large areas of argillic alteration near Wilson Springs in Garland County. Within such areas, fenite feldspathic breccias and metamorphosed sedimentary rocks have been altered and mineralized. Rocks of the Mississippian part of the Arkansas Novaculite may have been mineralized in the Wilson Spring area and also in similar ore concentrations near Magnet Cove in Hot Spring County. The vanadium rarely occurs as discrete vanadium minerals but generally as a vicarious element in several rock-forming minerals and their alteration products.

ZINC AND LEAD

Zinc and lead deposits are present in flat-lying carbonate rocks in the Ozark region and in folded shale and sandstone in the Ouachita Mountains.

In the Ozark region, the zinc and lead deposits are mostly in Newton County, a few being present in Searcy, Marion, and Boone Counties. Lead deposits are found in the Batesville Sandstone, but most of the lead and zinc deposits are in the Boone Formation. The ore deposits are in fracture zones, along joint systems, in irregularly shaped masses parallel to bedding of the host rock, and in some fault zones.

The ore bodies are primarily sphalerite and galena together with some chalcopyrite, pyrite, and marcasite. Secondary minerals are smithsonite, calamine (hemimorphite), cerussite, and traces of malachite, azurite, and aurichalcite. The galena is very low in silver, and the sphalerite contains small amounts of iron and cadmium.

From 1902 to 1962, the reported amount of zinc concentrates is 29,900 short tons. From 1907 to 1959, the reported amount of lead concentrates is 2,000 short tons.

In the Ouachita Mountains, small deposits of lead and zinc are present in mineralized quartz veins in the Stanley Formation in Sevier County, in the Stanley and the Arkansas Novaculite in Hot Spring County, and in the Jackfork Sandstone in Pulaski County. Galena, sphalerite, and chalcopyrite are the principal minerals. Small amounts of silver were present in the galena in all areas, larger amounts being associated with freiburgite in Pulaski County.

The amount of zinc and lead produced from the mines in these counties is unknown.

NONMETALLIC MINERALS

DIMENSION STONE

Dimension stone is being produced from sandstone in the Hartshorne Sandstone in Logan, Franklin, Pope, and Johnson Counties and in the Batesville Sandstone in Stone and Independence Counties. Black marble has been produced from the Pitkin Limestone and the Fayetteville Shale in Stone and Independence Counties, and gray marble, from the Boone Formation in Independence County.

LIMESTONE

High-calcium limestone is quarried from the Boone Formation in Independence County. It is in contiguous beds having an aggregate thickness of 33.7 m, and the quarried product is converted to lime. Limestone is also quarried from the Boone Formation and the Pitkin Limestone; the product is used as aggregate stone for construction purposes.

SLATE

Slate has been quarried from the Stanley Formation in Garland County, but its poor weathering characteristics precluded its use as an exterior finish. Slate is quarried from the Stanley in Montgomery County, where it is crushed and ground for use as roofing granules.

TRIPOLI

Tripoli is a microcrystalline finely particulate friable form of silica. Deposits of tripoli are present in the Boone Formation in Benton, Washington, and Madison Counties, the largest being in Benton County. The tripoli is the siliceous remnants of weathered calcareous chert layers. Deposits of tripoli are also present in the Upper Division of the Arkansas Novaculite in Pulaski, Garland, Montgomery, and Polk Counties, but most of it has been produced from Garland County. The tripoli is formed by the weathering of calcium carbonate from the novaculite.

The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— Nebraska

By R. R. Burchett

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1110-P

Prepared in cooperation with the Nebraska Geological Survey, Conservation and Survey Division Institute of Agriculture and Natural Resources, The University of Nebraska-Lincoln

Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Nebraska

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—NEBRASKA

By R. R. BURCHETT [1]

ABSTRACT

Nebraska is underlain by rocks of Mississippian and (or) Pennsylvanian age in all but the extreme northeastern part of the State. Only some of the Upper Pennsylvanian strata are exposed; outcrops of these rocks are limited to southeastern Nebraska. The thickness of the Pennsylvanian rocks ranges from a featheredge in northeastern Nebraska to slightly more than 625 m (2,050 ft) in the southeastern part of the State, in the deeper part of the Forest City basin. Outcropping Pennsylvanian rocks have a combined thickness of 244 m (800 ft) and are composed principally of thin to thick beds of shale, sandstone, and limestone. Several thin coal seams occur in both the exposed and deeply buried sequences of Pennsylvanian strata. In ascending order, the Pennsylvanian System in Nebraska is divided into the following series: Morrow(?), Atoka, Des Moines, Missouri, and Virgil. Of these, only the Missouri and Virgil Series are exposed. Pennsylvanian rocks in Nebraska, especially the Missouri Series and the lower part of the Virgil Series, consist of repeated sequences of marine and nonmarine beds. Some of the marine beds contain abundant invertebrate fossils. Shale associated with coal seams contains plant fossils.

Formation of Nebraska's structural features began before and continued through Carboniferous time. Several stratigraphic units recognized in the Forest City basin do not extend over the Nemaha arch, which borders on the west.

Several quarries have been opened into Pennsylvanian limestone and shale. Other important economic products are oil, gas, and water. Although coal was mined for local use at several places, it has not been mined for several years.

INTRODUCTION

Nebraska, in the northern midcontinent part of the United States, is underlain by rocks of Mississippian and (or) Pennsylvanian age in all but the northeastern corner of the State. (See figs. 1 and 2.) However, only some of the Upper Pennsylvanian strata are exposed, and the outcrops of these rocks are restricted to valley sides in the southeastern part of the State (fig. 3). For this reason, the following description of Carboniferous rocks in Nebraska is limited almost entirely to those deposited during Late Pennsylvanian time.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Nebraska Geological Survey, Conservation and Survey Division.

HISTORY

The initial recognition and investigation of Carboniferous strata in the northern midcontinent region took place more than a century ago (Owen, 1852, p. 133–138, Sections 20M–40M; Marcou, 1864; Prosser, 1897, p. 12–16; Merrill, 1924, p. 773). However, the first noteworthy report on Pennsylvanian stratigraphy in the Missouri-Kansas-Nebraska region was that by Broadhead (1873). Other important early work was reported by Meek and Hayden (1859), Swallow (1866), Meek (1872), Barbour (1903), Condra (1903), Woodruff (1906), Prosser (1902), Hinds and Greene (1915), and Tilton (1920).

Initially, the State Geological Surveys of the region worked independently in the study, naming, and mapping of geologic units. This independent activity caused duplication of names, confusion in correlation, and demonstrated a need for interstate study. It was followed by cooperative surveys across State lines, by regional correlation, and by the establishment of nomenclature on a priority basis. However, the thick Pleistocene deposits in the northern part of the region and the folding and faulting in the southern midcontinent region made it difficult to run traverses between outcrops and very difficult to correlate the formations in the more or less buried structures. However, by detailed lithologic and faunal study of the outcrops, by logging deep wells drilled for oil and gas, and by test drilling to obtain

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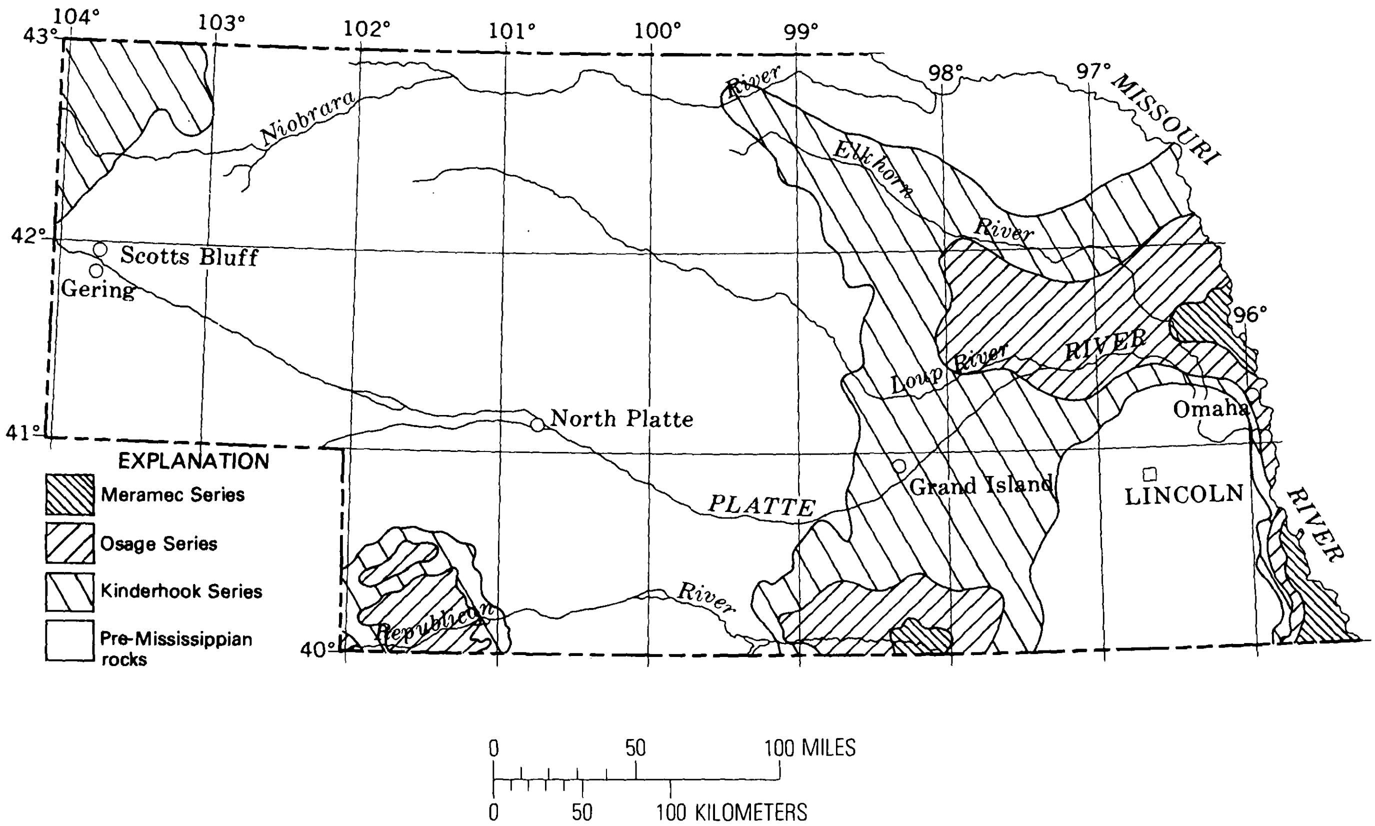


FIGURE 1.—Distribution of the Mississippian System in Nebraska (from Carlson, 1963).

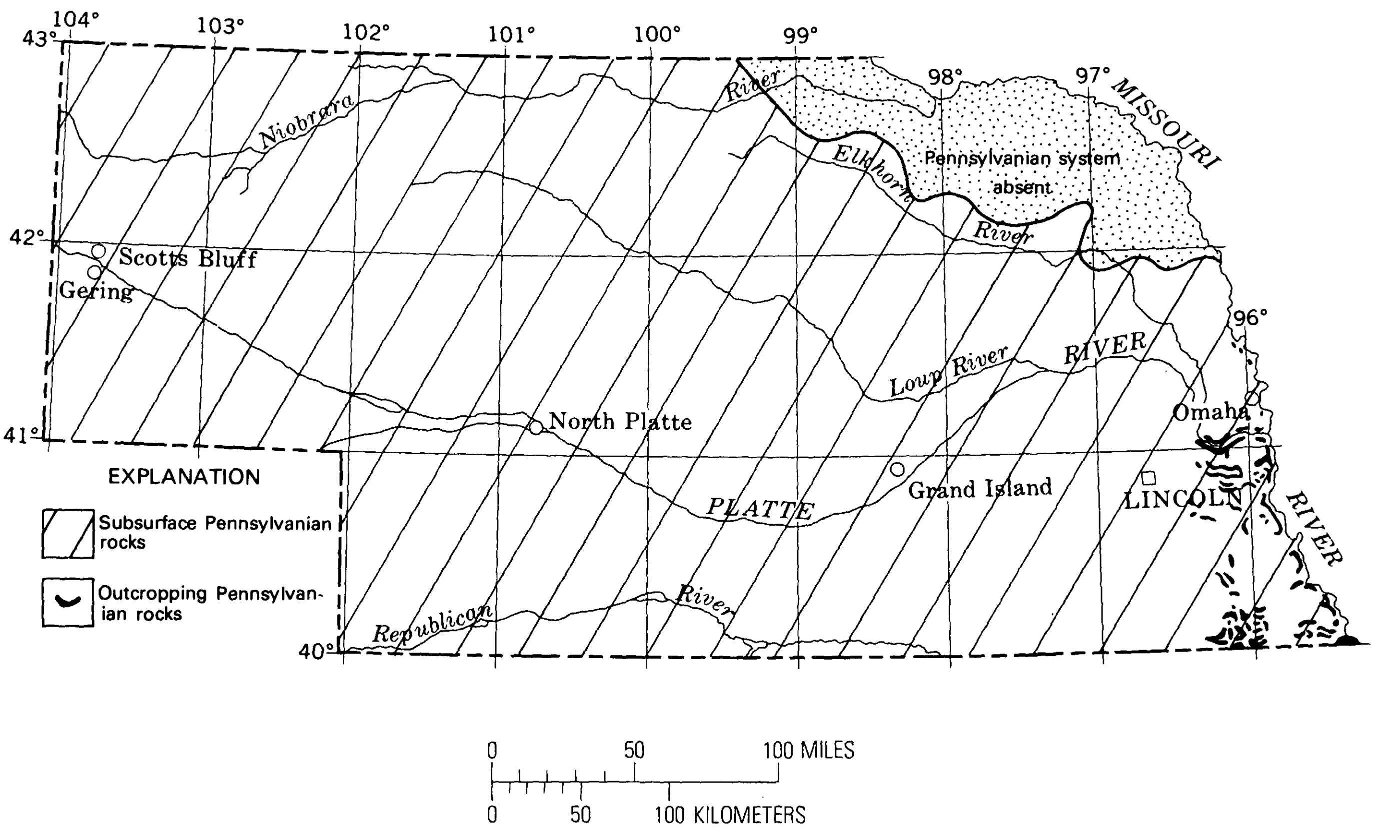


FIGURE 2.—Distribution of the Pennsylvanian System in Nebraska.

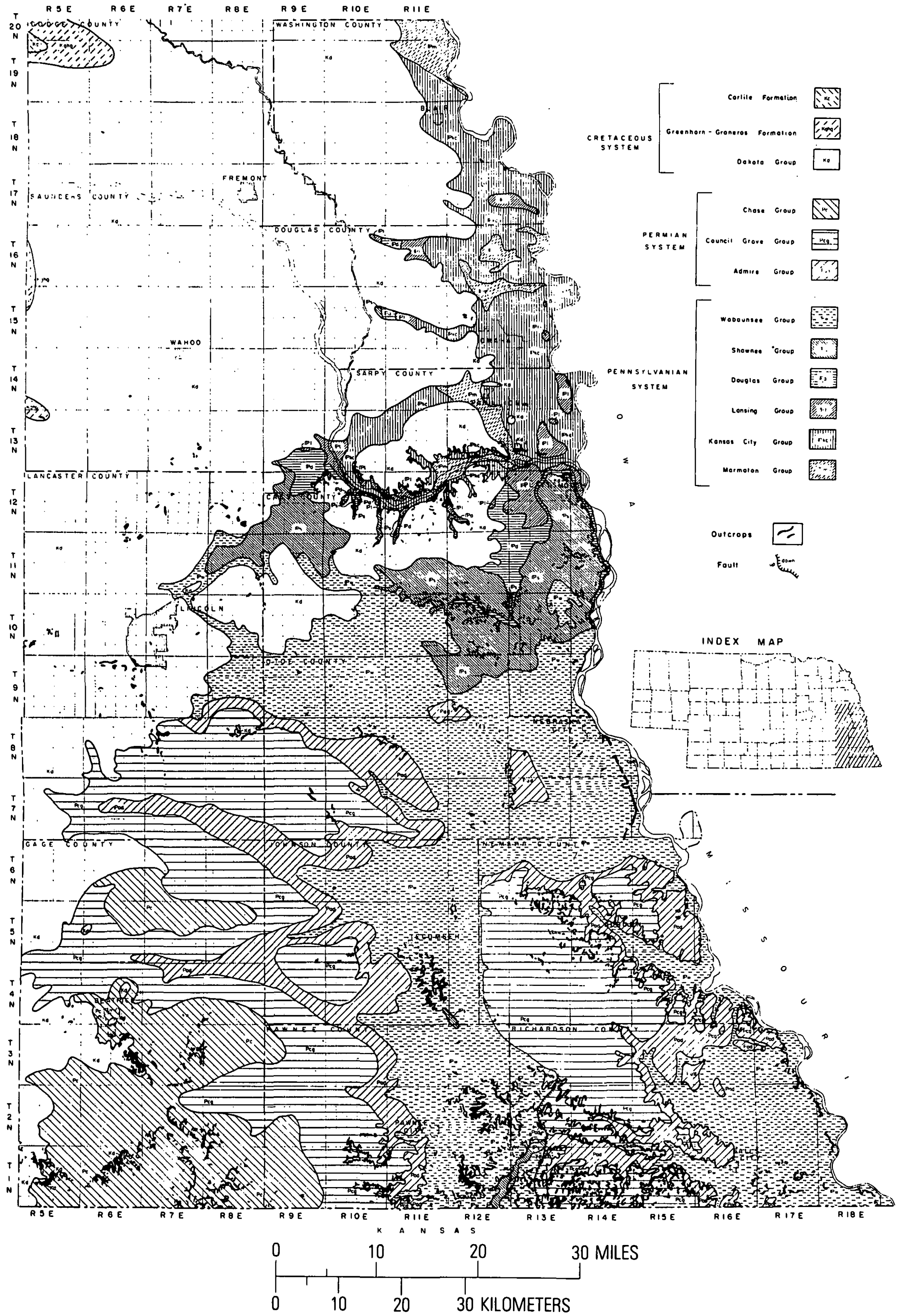


FIGURE 3.—Bedrock geology of southeastern Nebraska.

subsurface information at various locations, factual data were gathered for use in the correlation and classification of the units of the system. This kind of study by the Surveys, with cooperation from oil geologists, established a knowledge of the nature, thickness, and regional occurrence of the subdivisions in the region and gave a sound basis for the description, correlation, and naming of the members, formations, groups, and series. It revealed some errors in the early surveys and reports and, contrary to early interpretation, showed that the formations are comparatively uniform in thickness in the northern part of the region and that most of them were formed as cyclothemal units of marine and continental deposition.

During the past 65 years, but particularly after the appearance of the important paper on the Pennsylvanian of Nebraska by Condra and Bengston (1915), the modern Pennsylvanian classification has emerged. The classification for Nebraska was revised in 1927 (Condra, 1927) and then almost yearly for about 30 years (see especially Condra and Reed, 1943; Condra, 1949; Moore, 1932, 1936, 1944, 1948, 1949; Moore and Newell, 1937; Moore and others, 1944, 1951; Moore and Mudge, 1956; and Kansas Geological Society, 1957). More recent publications include Hershey and others, 1960; Howe and Koenig, 1961; Mudge and Yochelson, 1963; Reed and Burchett, 1964; Burchett and Carlson, 1966; Smith and Burchett 1967; Burchett and Reed, 1967; Jewett, O'Connor, and Zeller, 1968; Burchett, 1968, 1969, 1970, 1971; Fagerstrom and Burchett, 1972; Prichard, 1975; and Burchett, 1977.

GEOLOGIC SETTING

Pennsylvanian rocks in Nebraska rest unconformably on rocks ranging in age from Precambrian to Mississippian. Generally, the unconformity is marked by a basal sand or a detrital zone of angular quartz sand and weathered chert. The pre-Pennsylvanian rocks have been studied in detail by Carlson (1963, 1970).

Pennsylvanian strata in Nebraska are overlain by rocks ranging in age from Permian to Quaternary. Where the Pennsylvanian-Permian contact is visible in southeastern Nebraska, a period of erosion without significant diastrophism is indicated; at least one channel sand of Permian age cuts approximately 30 m (100 ft) into Pennsylvanian rocks. Placement of the Pennsylvanian-Permian boundary has been summarized by Moore (1940, p. 298–305; 1949, p. 19–22) and by Mudge and Yochelson (1962, p. 116–127).

The principal structural features of Nebraska are shown in figure 4. Major movement on many of these features took place in Early Pennsylvanian time.

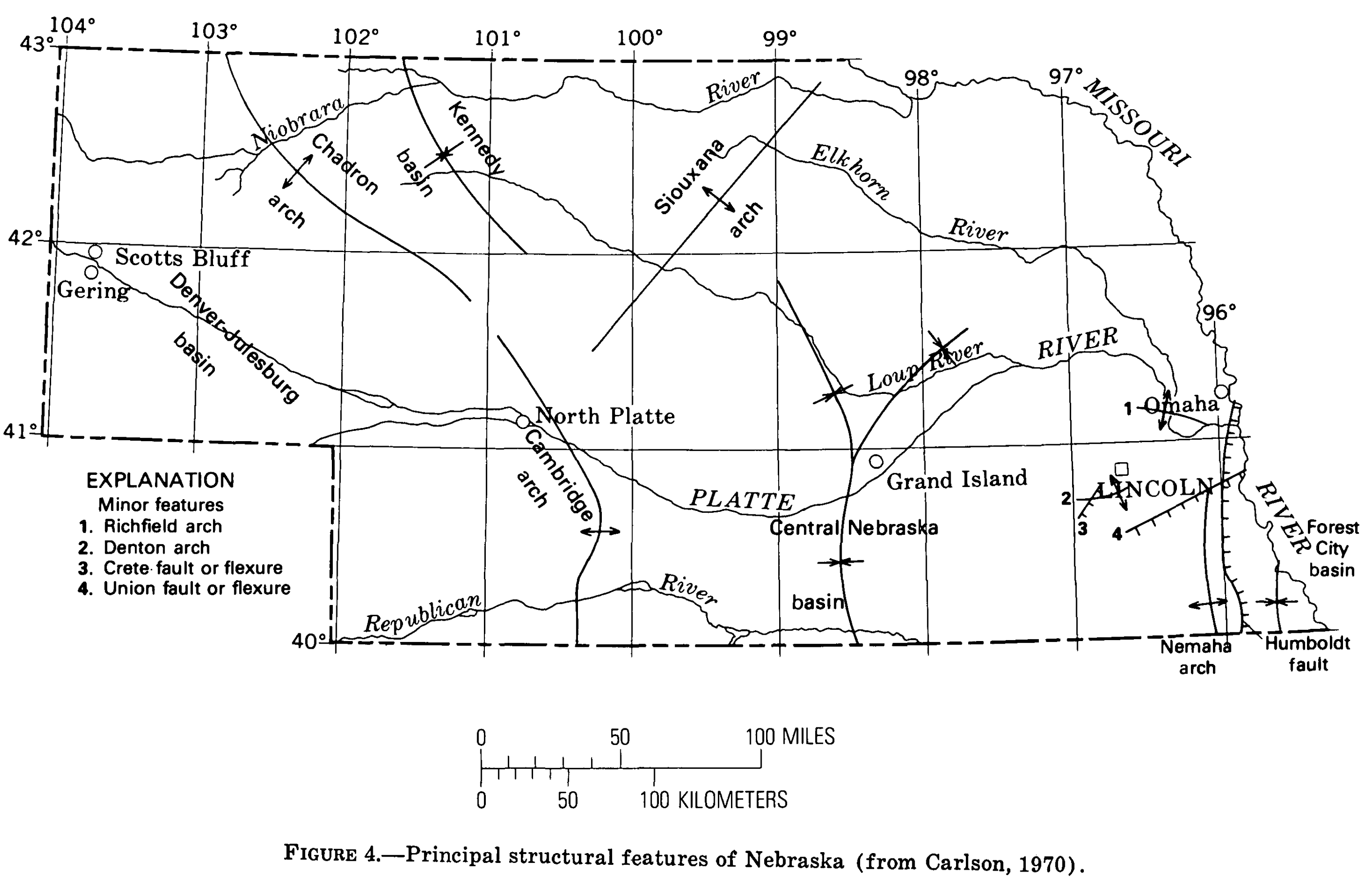


FIGURE 4.—Principal structural features of Nebraska (from Carlson, 1970).

Formation of the Nemaha arch in post-Mississippian pre-Des Moines (Middle Pennsylvanian) time accounts for the unconformable relation of Pennsylvanian strata to the underlying older rocks in eastern Nebraska. This arch extends from the vicinity of Omaha southward across Kansas and into Oklahoma. In much of Johnson and Pawnee Counties, Nebr., and in the adjoining part of Kansas, uplift was so great that all strata from Middle Ordovician to Mississippian age were removed by concurrent and subsequent erosion, resulting in exposure of Precambrian rocks. Although deposition of Pennsylvanian strata in eastern Nebraska began during Atoka time, complete burial of the arch did not take place until early Missouri time. Crustal unrest continued through the deposition of the Pennsylvanian rocks and into deposition of the Permian rocks. Contemporary subsidence of the areas on either side of the arch resulted in deposition of thicker sequences of Pennsylvanian strata in the downwarped basins. Because the area immediately east of the arch subsided at a faster rate, the thickness of Pennsylvanian strata accumulating there was significantly greater than that on the west side. Now known as the Forest City basin, that area of subsidence includes parts of Nebraska, Iowa, Missouri, and Kansas.

Not all the stratigraphic units in the Pennsylvanian sequence are useful as datum planes. For example, the initial clastic sediments were deposited on the irregular surface of older rocks and thus differ in thickness within short distances; furthermore, most individual beds deposited on or close to that surface probably deviated from the horizontal. Only those beds that originally were nearly horizontal, are of large areal extent, and are conformable with underlying and overlying beds are suitable as datums. One such bed is the Hertha Limestone, the base of which was used as the datum plane for the structural contour lines shown in figure 5. Altitudes of the base of the limestone are shown to range from slightly more than 305 m (1,000 ft) above mean sea level in Douglas and Sarpy Counties to slightly more than 91 m (300 ft) below mean sea level in Richardson County, where the basin is deepest. The Table Rock arch, a structural feature in the outcrop area, trends north over the buried Nemaha arch; the east margin of the Table Rock arch is sharply defined by the Humboldt fault and smaller faults associated with it. (See fig. 5.)

LITHOSTRATIGRAPHY OF THE PENNSYLVANIAN SYSTEM

The thickness of Pennsylvanian rocks in Nebraska ranges from a featheredge in the northeastern part of the State to slightly more than 625 m (2,050 ft) in the extreme southeastern part (fig. 6). Outcropping Pennsylvanian rocks have a combined thickness of about 244 m (800 ft), as shown in figure 7.

Composed primarily of thin to thick beds of shale, sandstone, and limestone, the Pennsylvanian sequence in Nebraska is characterized by a definite repetition of cycles of marine shale and limestone alternating with nonmarine deposits. However, lateral lithologic differences are found within the formations.

Pennsylvanian rocks in Nebraska have been divided into five series. In ascending order, these are the Morrow(?), Atoka, Des Moines, Missouri, and Virgil. Of these, only the rocks of the Missouri and Virgil Series are exposed.

MORROW(?) SERIES

The oldest Pennsylvanian rocks in Nebraska have been assigned to the Morrow(?) Series. They may be present in the southwestern part of the State, but definite age relationships have not been established. Rocks of the Morrow(?) Series are not present in the Forest City basin, where each of the four other series has been identified.

ATOKA SERIES

Strata assigned to the Atoka Series are the oldest Pennsylvanian rocks in southeastern Nebraska. In the Forest City basin, they are about 91 m (300 ft) thick and consist mostly of dark shale but include some sandstone layers. Because they do not crop out, they are known only from logs of drill holes.

DES MOINES SERIES

Rocks of the Des Moines Series overlie those of the Atoka Series. They are about 229 m (750 ft) thick in the deepest part of the Forest City basin, are thinner—less than 122 m (400 ft)—at the northern end of the basin, and are very thin or absent over the Nemaha arch. (See fig. 8.) The series has been divided into the older Cherokee Group and the younger Marmaton Group. Like the Atoka Series, the Des Moines Series does not crop out and is known only from logs of drilled holes.

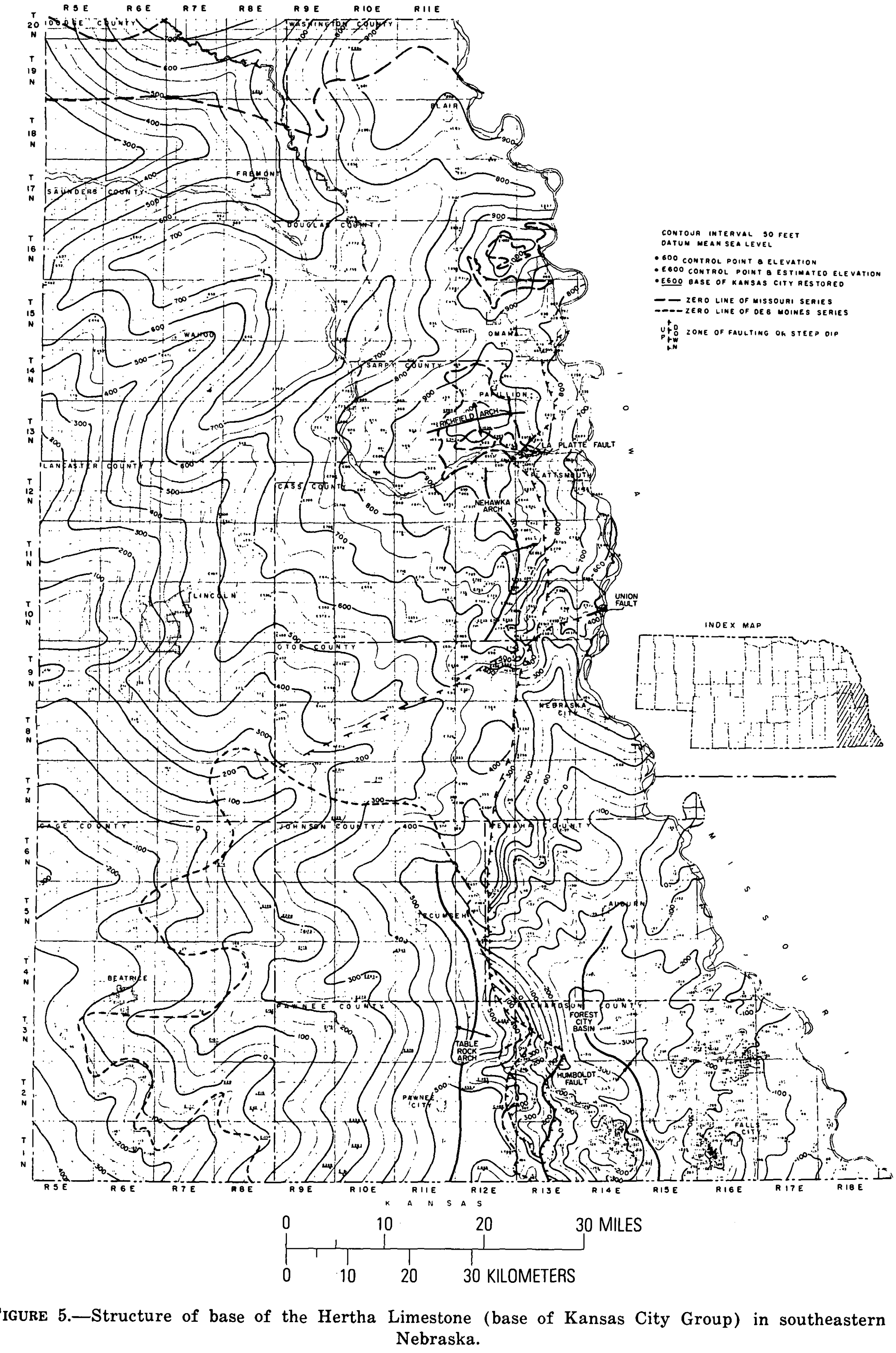


FIGURE 5.—Structure of base of the Hertha Limestone (base of Kansas City Group) in southeastern Nebraska.

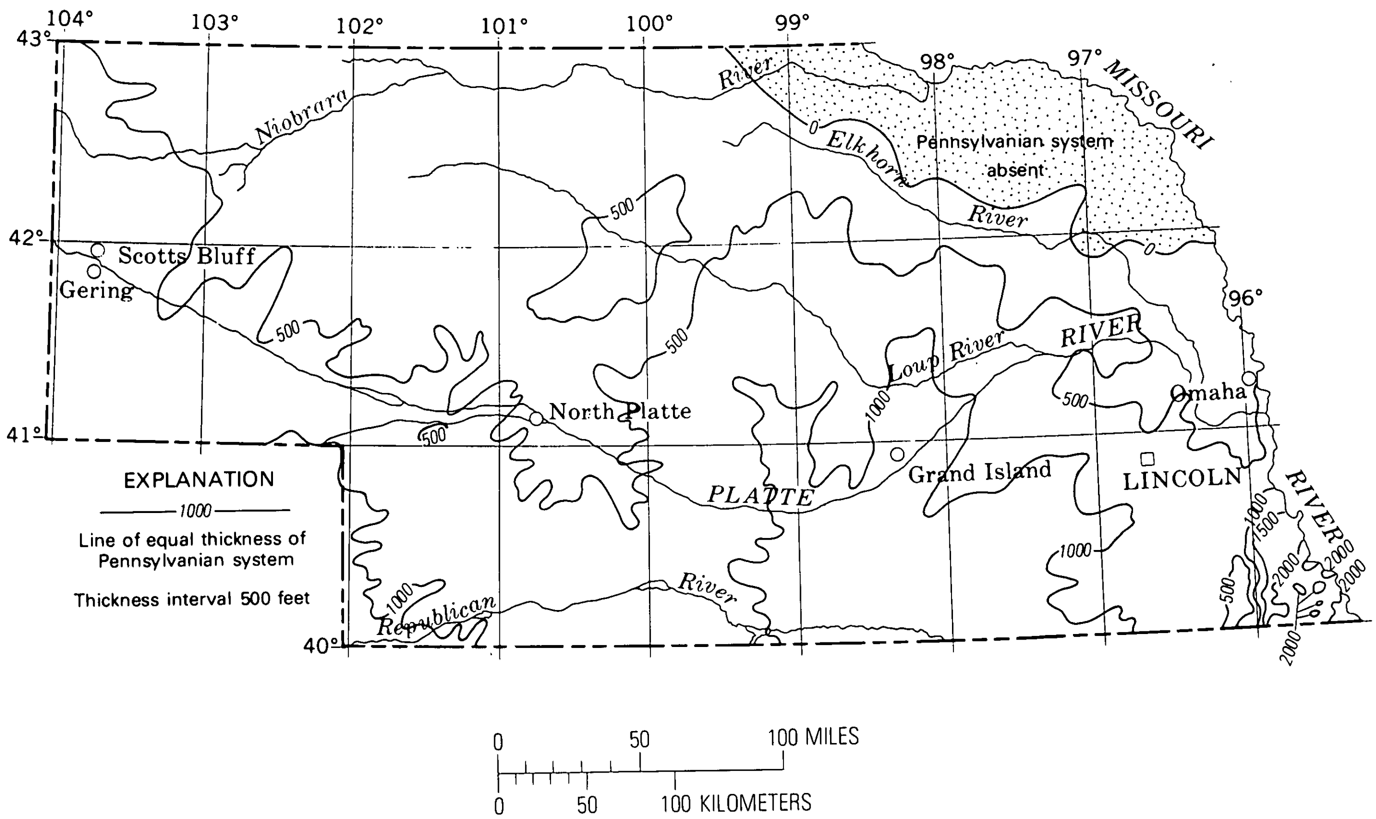


FIGURE 6.—Thickness of the Pennsylvanian System in Nebraska.

CHEROKEE GROUP

The Cherokee Group consists mostly of varicolored shale but includes several beds of sandstone and coal plus a few beds of thin limestone. In the deepest part of the Forest City basin, the group has a maximum thickness of about 175 m (575 ft), and near the northern end of the basin its thickness ranges from about 6.1 m (20 ft) to 76 m (250 ft). Entirely in the subsurface, the Cherokee Group has not been divided into formations in Nebraska.

MARMATON GROUP

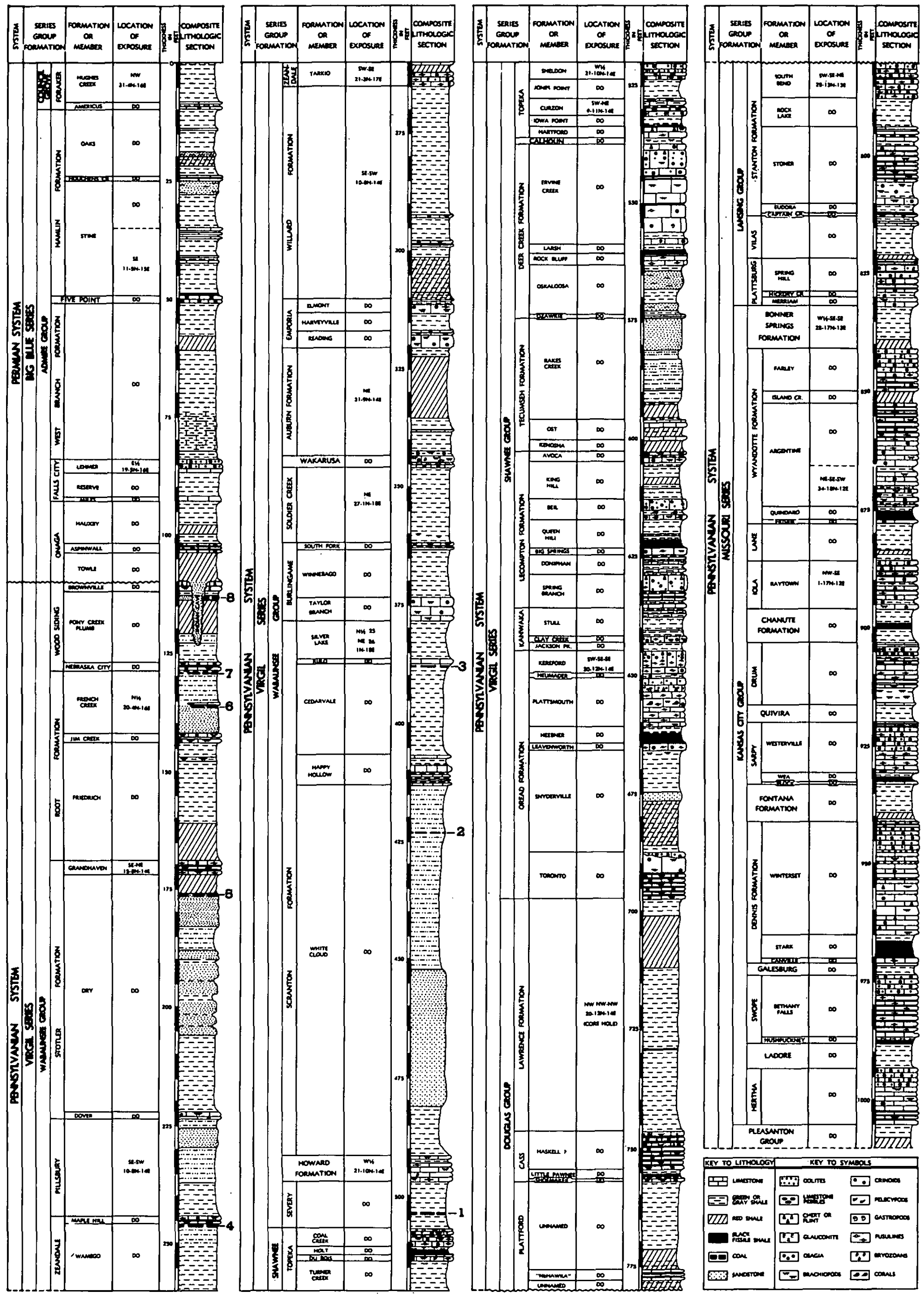
Marmaton rocks in southeastern Nebraska consist in large part of varicolored shale but include considerable limestone, some sandstone, and a few thin beds of coal. Thickness of the group ranges from about 53 m (175 ft) in the deeper part of the basin to 35 m (115 ft) in the northern part. Possibly some strata belonging to this group extend westward over the crest of the Nemaha arch. Although the Marmaton Group is somewhat difficult to divide in the subsurface of Nebraska, six formations have been tentatively identified. In ascending order, they are the Fort Scott Limestone, Labette Shale, Pawnee Limestone, Bandera Shale, Altamont Limestone, and Nowata Shale. A disconformity separating Marmaton rocks from the overlying Missouri Series indicates that the upper Marmaton strata were exposed to subaerial erosion before deposition began again in the area.

MISSOURI SERIES

The oldest outcropping rocks of Pennsylvanian age in Nebraska belong to the Missouri Series. They consist mainly of interbedded limestone and shale but include a few beds of sandstone. Individual beds of this series have a large areal extent and can be traced from one outcrop to another. The upper boundary of this series is unconformable with the overlying Virgil Series. The lower boundary is regionally disconformable with the underlying Des Moines Series. As shown in figure 9, the thickness of this series is slightly more than 91 m (300 ft) near the center of the Forest City basin and is about 61 m (200 ft) thick in the northern part of the basin. Three groups composing the series are, from oldest to youngest, the Pleasanton, Kansas City, and Lansing.

PLEASANTON GROUP

Most of the rocks of the Pleasanton Group are mottled red and green shales; some sandstone is also



8. Honey Creek coal
7. Lorton coal
6. French Creek coal
5. Dry coal
4. Wamego coal
3. Elmo coal
2. White Cloud coal
1. Nodaway coal

FIGURE 7.—Composite section of outcropping Upper Pennsylvanian and Lower Permian rocks in southeastern Nebraska.

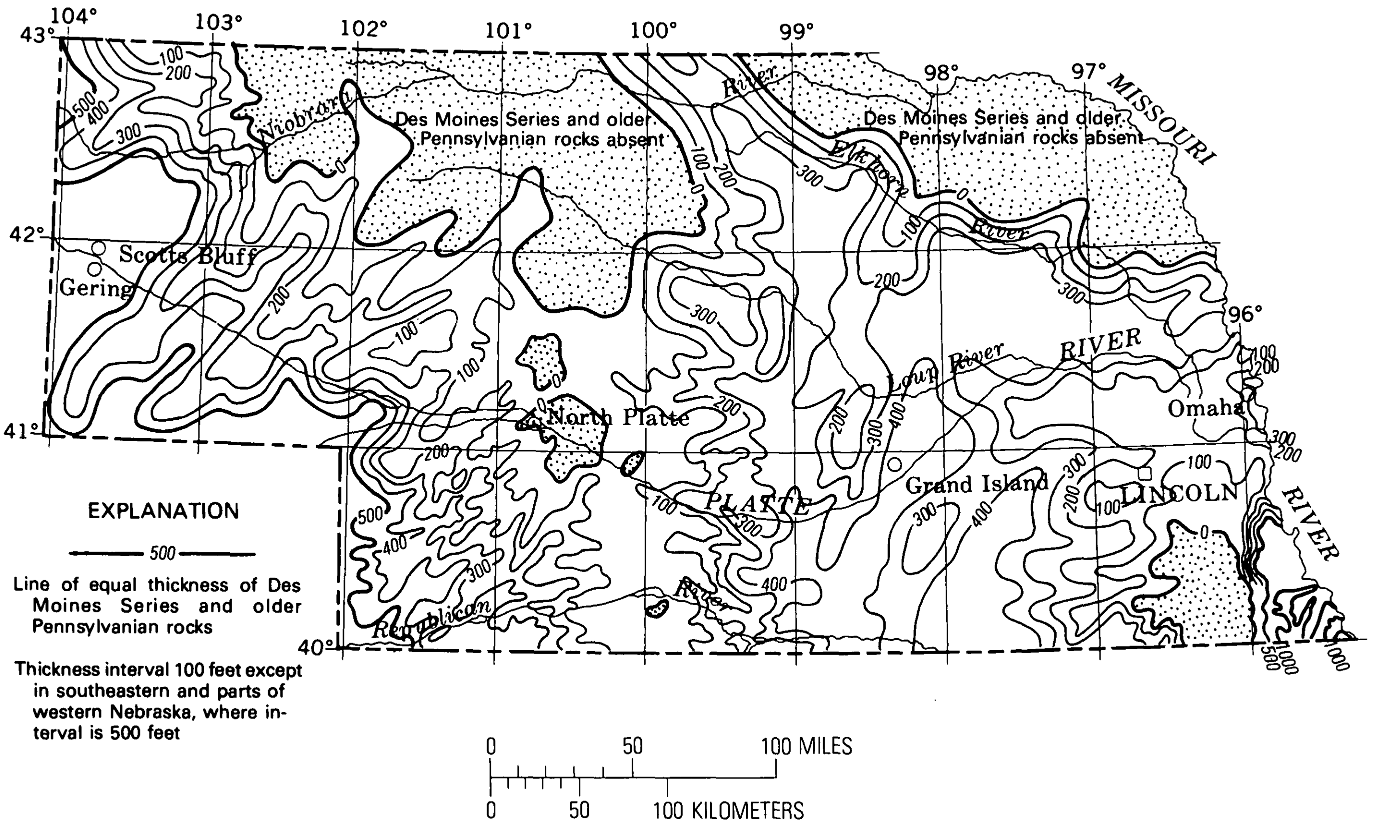


FIGURE 8.—Thickness of the Des Moines series and older Pennsylvanian rocks in Nebraska.

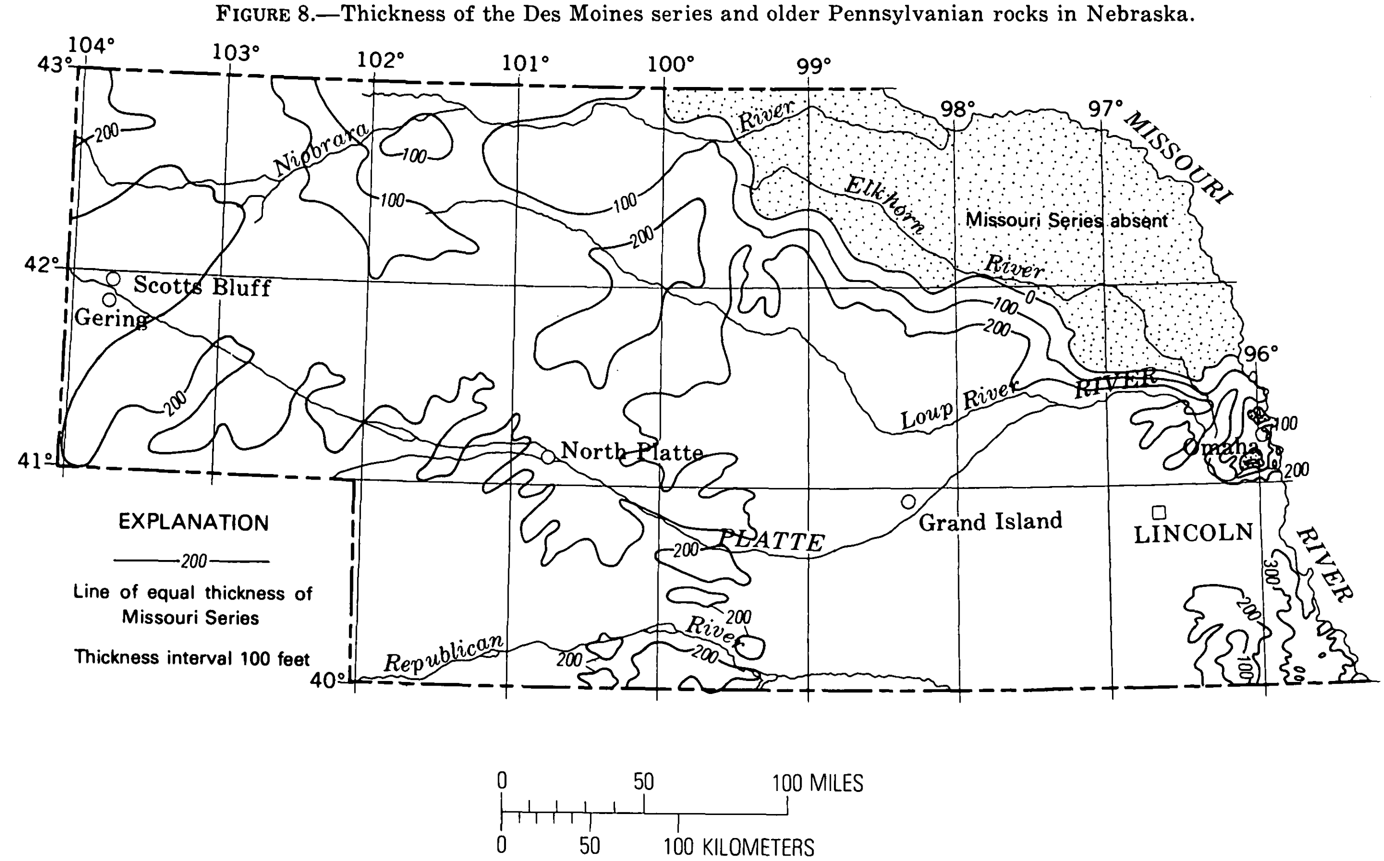


FIGURE 9.—Thickness of the Missouri Series in Nebraska.

present. Resting unconformably on rocks of the Marmaton Group, these clastic sediments range in thickness from a featheredge to about 8 m (25 ft) in southeastern Nebraska and do not crop out anywhere in the Nebraska part of the Forest City basin.

KANSAS CITY GROUP

Rocks of the Kansas City Group consist of cyclically alternating limestone and shale. In Cass, Sarpy, Saunders, and Washington Counties, where this group is exposed, its minimum thickness is about 49 m (160 ft), and, in the subsurface of southeastern Nebraska, its maximum thickness is about 79 m (260 ft). The lower half of the group is missing over the Nemaha arch. From oldest to youngest, the 14 formations composing the group are as follows: Hertha Limestone, Ladore Shale, Swope Limestone, Galesburg Shale, Dennis Limestone, Fontana Shale, Sarpy Limestone, Quivira Shale, Drum Limestone, Chanute Shale, Iola Limestone, Lane Shale, Wyandotte Limestone, and Bonner Springs Shale. (See fig. 7.)

LANSING GROUP

Limestone and shale are the principal constituents of the Lansing Group. Where this group crops out in Douglas, Sarpy, and Saunders Counties, it has a maximum thickness of about 15 m (50 ft), and, in the deeper part of the Forest City basin, its thickness is as much as 18 m (60 ft). Of the three formations composing this group, the Plattsburg Limestone is the lowest, the Vilas Shale is the middle, and the Stanton Limestone is the highest. The upper surface of the Stanton is an unconformity that separates the Missouri Series from the overlying Virgil Series.

VIRGIL SERIES

The youngest exposed Pennsylvanian rocks in Nebraska constitute the Virgil Series. These rocks are marked at the upper and lower boundaries by unconformities and are cyclical, as are those in the Missouri Series. The Virgil rocks consist mainly of shale, limestone, sandstone, and thin coal. Outcrops are widespread in southeastern Nebraska where some of the upper beds have been removed by erosion over the Nemaha arch and in the northern end of the Forest City basin. As shown by figure 10, the thickness of the Virgil ranges from a featheredge at the northern end of the basin to a little more than 244 m (800 ft) in the deeper part of the Forest City basin. The Virgil Series is divided into three groups in Ne-

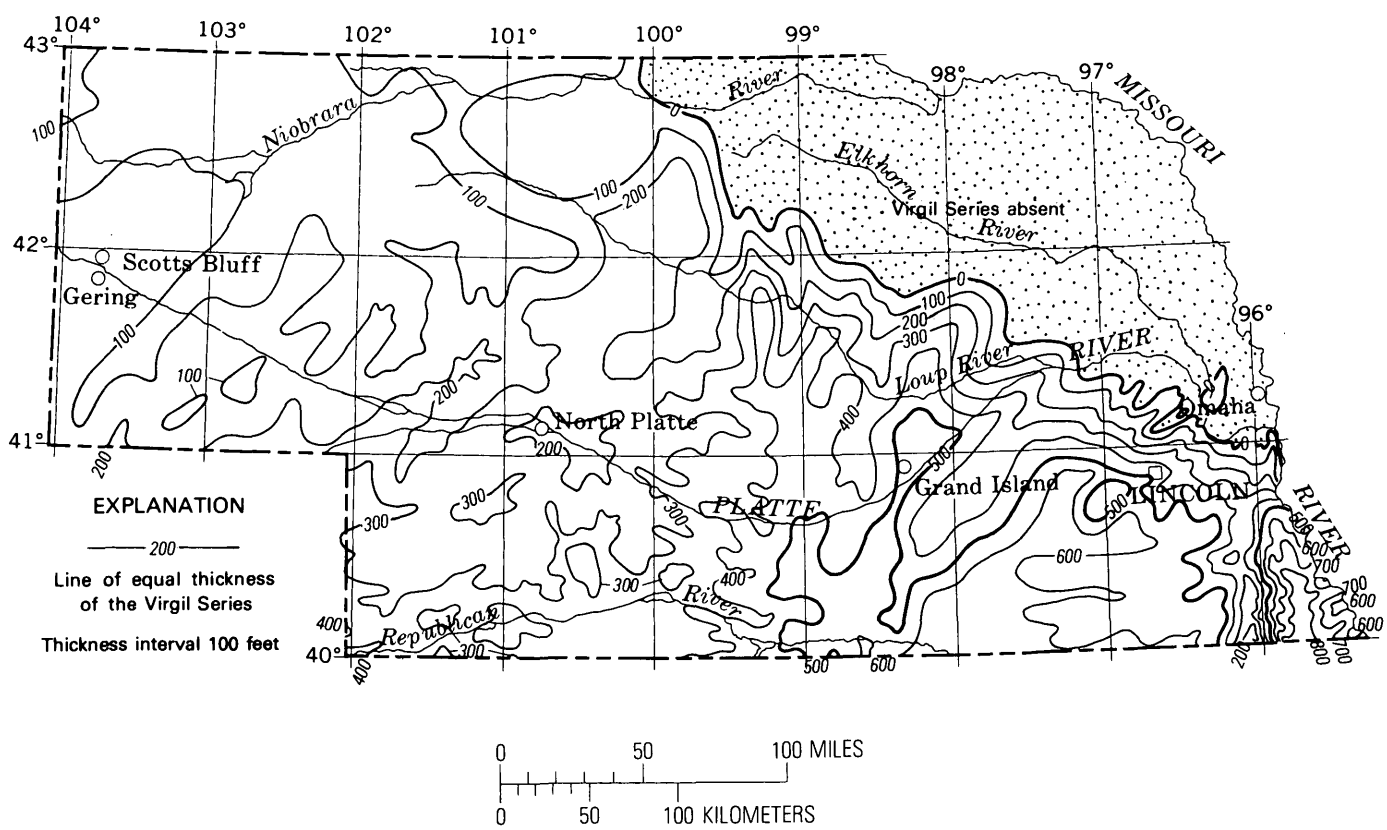


FIGURE 10.—Thickness of the Virgil Series in Nebraska.

braska, from oldest to youngest: Douglas Group, Shawnee Group, and Wabaunsee Group . (See fig. 7.)

DOUGLAS GROUP

The Douglas Group consists mostly of shale interbedded with some limestone but also includes some sandstone and a few beds of thin coal. Where the group crops out in Cass and Sarpy Counties, it is about 18 m (60 ft) thick, but in the subsurface it thickens southward to a maximum of 46 m (150 ft) in Richardson County. In ascending order, the Douglas Group is composed of the Plattford Shale, Cass Limestone, and Lawrence Shale.

SHAWNEE GROUP

Conformably overlying the Douglas Group is a distinctive cyclic sequence, the Shawnee Group, which consists mainly of limestone interbedded with several layers of shale and a few layers of sandstone. This group crops out in Cass, Otoe, Pawnee, and Saunders Counties. Its thickness ranges from 53 m (175 ft) on the Nemaha arch to 76 m (250 ft) in the Forest City basin. From oldest to youngest, the Shawnee Group comprises the following seven formations: Oread Limestone, Kanwaka Shale, Lecompton Limestone, Tecumseh Shale, Deer Creek Limestone, Calhoun Shale, and Topeka Limestone.

WABAUNSEE GROUP

Although the Wabaunsee Group includes several thin persistent layers of limestone and a few very thin coal beds, it consists mostly of shale, sandy shale, and sandstone. Outcrops occur in Cass, Johnson, Nemaha, Otoe, Pawnee, and Richardson Counties; wherever the basal beds are exposed, the Wabaunsee Group conformably overlies the Shawnee Group. Exposures of the group's upper boundary indicate that a period of erosion preceded deposition of the overlying rocks of Permian age. Locally, channels incised as much as 30 m (100 ft) into Wabaunsee rocks are filled with Permian sandstone. In the northern part of the Forest City basin and over the Nemaha arch, the upper part of the Wabaunsee Group either was not deposited or was removed by pre-Permian erosion. The maximum thickness of the group is about 122 m (400 ft). From oldest to youngest, the following 14 formations constitute the Wabaunsee Group: Severy Shale, Howard Limestone, Scranton Shale, Burlingame Limestone, Soldier Creek Shale, Wakarusa Limestone, Auburn Shale, Emporia Limestone, Willard Shale, Zeandale Limestone, Pillsbury Shale, Stotler Limestone, Root Shale, and Wood Siding Formation.

ENVIRONMENTS OF DEPOSITION

Pennsylvanian rocks in Nebraska, especially the Missouri Series and the lower part of the Virgil Series, consist of repeated sequences of beds. Such evidence for cyclic sedimentation characterizes the Pennsylvanian rocks throughout much of the midcontinent area.

Discovery of cyclic sedimentation is credited to Udden (1912), who noted that certain sequences of Pennsylvanian strata exposed in the Peoria quadrangle of western Illinois were clearly divisible into four parts, the same order of parts being repeated several times. Because one of the parts was a bed of coal and another a bed of limestone, each sequence, or cycle, indicated an alternation of nonmarine and marine deposition. A cycle was defined as all the strata between the base of one coal bed and the base of the next younger coal bed.

Weller (1930, 1931), who believed that diastrophism accounted for cyclical Pennsylvanian sedimentation, redefined a cycle. According to him (1931, p. 163), an ideal cycle consists of nine parts, listed from oldest (1.) to youngest (9.) as follows:

9. Shale containing "ironstone" bands in upper part and thin limestone layers in lower part.
8. Limestone.
7. Calcareous shale.
6. Black "fissile" shale.
5. Coal.
4. Underclay.
3. Fresh-water limestone.
2. Sandy and micaceous shale.
1. Sandstone with unconformity at the base.

In 1931, the Illinois State Geological Survey sponsored a symposium on cyclic sedimentation in the Pennsylvanian System. Those who attended included R. C. Moore, F. B. Plummer, D. B. Reger, G. H. Ashley, W. E. Stout, H. R. Wanless, and J. M. Weller, each of whom presented papers describing evidence for cyclic sedimentation in his part of the Central United States. The Illinois State Geological Survey published these papers in its Bulletin 60 (Weller, 1931).

Soon after the symposium, Wanless and Weller (1932) proposed that cyclic repetitions of strata, or "cyclothems," could be used to confirm interbasin and possibly regional correlations.

In 1936, Moore presented a slightly different, more detailed description of an ideal cyclothem. His sequence of beds, also listed oldest (.0) to youngest (.9), was as follows (1936, pp. 24–25) .

.9 Shale (and coal).
.8 Shale, typically with molluscan fauna.

.7 Limestone, algal, molluscan, or with mixed molluscan and molluscoid fauna.
.6 Shale, molluscoids dominant.
.5 Limestone, contains fusulinids, associated commonly with molluscoids.
.4 Shale, molluscoids dominant.
.3 Limestone, molluscan, or with mixed molluscan and molluscoid fauna.
.2 Shale, typically with molluscan fauna.
.1 *c*. Coal.
.1 *b*. Underclay.
.1 *a*. Shale, may contain land plant fossils.
.0 Sandstone.

In the same publication, Moore introduced the term "megacyclothem" for "a cycle of cyclothems." According to him, each cyclothem in a megacyclothem could be distinguished by the type of limestone it included. The limestone beds in cyclothems composing a megacyclothem were designated, in ascending order, as "lower," "middle," "upper," and "super" limestone sequences. Later regional studies showed that in a few localities the lower or upper parts of some megacyclothems are missing.

Moore's cyclothemic and megacyclothemic classification for Upper Pennsylvanian strata has aided the understanding of the changing environmental conditions during deposition of these beds.

Examination of sequential relationships in the Upper Pennsylvanian of the Nebraska part of the Forest City basin indicates that the megacyclothems differ somewhat from the ideal. If sedimentation took place during the early part of the megacycle, it usually was interrupted one or more times, resulting in the removal, by erosion, of most or all beds that had been deposited since the beginning of the megacycle. Thus, the early part of the megacyclothem generally is either thin or lacking. On the other hand, the middle and upper parts of the megacyclothems appear to be complete, or nearly so, and the sequence of beds is similar to that in the ideal megacyclothem already described.

Numbered from earliest (1.) to latest (7.), the succession of depositional events constituting most of the Nebraska megacyclothems is interpreted to have been about as follows:

7. Deposition of shale and sand under conditions of rapid accumulation and progressive return to nonmarine environment; swamp development in coastal plain.
6. Deposition of the "super" limestone in an increasingly brackish environment; only small to moderate amounts of argillaceous and siliceous impurities derived from adjacent land areas.
5. Deposition of dark shale in shallow seawater under conditions of slow accumulation of argillaceous and siliceous erosional products from adjacent land areas.
4. Deposition of the "upper" limestone when swamp drainage ceased and seawater cleared; shallowing of the sea resulted in formation of oolites in limestone and eventual cessation of carbonate deposition.
3. Deposition of black fissile shale when headward erosion of streams caused drainage of inland swamps and inflow of humic material to the sea.
2. Simultaneous deposition of thin "middle" limestone and shoreward expansion of inland swampy areas.
1. Simultaneous deposition of gray-green and red shale in a marine nearshore environment and formation of inland swamps.

The several rock groups composing the Upper Pennsylvanian in Nebraska differ somewhat in those parts of the complete megacyclothem of Moore (1936, pp. 24–25) that are represented, as indicated below:

Kansas City Group.—Part including "lower" limestone consistently absent; parts including "middle" and "upper" limestone sequences always present. Although the part that includes the "super" limestone sequence is generally present, it is absent from some megacycles.

Lansing Group.—All four limestone sequences generally represented.

Douglas Group.—Consists mostly of shale but includes the Cass Limestone, which is made up of a "middle" and an "upper" limestone separated by black fissile shale.

Shawnee Group.—All four limestone sequences generally represented.

Wabaunsee Group.—Consists primarily of shale, sandy shale, and sandstone interbedded with thin to moderately thick marine limestone that generally occurs in pairs. The lower limestone of each pair commonly is the thicker and more persistent; the upper commonly is less pure and is discontinuous. Absent from the Wabaunsee Group are limestone beds of the "lower" and "middle" limestone sequences, and also absent is the black fissile shale ordinarily associated with the "middle" limestone sequence. All the limestones in the group probably are thin representatives of the "upper" and "super" limestone sequences.

BIOSTRATIGRAPHY

Some of the marine beds in the Upper Pennsylvanian of Nebraska contain abundant fossils of invertebrates. Brachiopods, pelecypods, gastropods, crinoids, corals, bryozoans, and fusilinids are the more common types. A few fossils of trilobites and ostracodes also are found in these rocks. Publications by Condra (1903), Dunbar and Condra (1927; 1932), Miller, Dunbar, and Condra (1933), and Pabian (1970) describe the fossil fauna present in exposed Upper Pennsylvanian rocks.

Terrestrial beds, particularly those associated with coal, commonly contain fossil plant remains. Leaves of the ferns *Neuropteris Loshi* Brongt. and *Neuropteris hirsuta* Lesq., stems of rushes, and stems of *Calamites* were reported (Hayden, 1868, p. 327–329) from beds associated with coal seams at Brownville in Nemaha County and at Rulo in Richardson County. However, no detailed studies of the fossil flora have been made to date.

ECONOMIC PRODUCTS

COAL

Discovery of coal outcrops in southeastern Nebraska in the 1850's prompted wild claims about an inexhaustible supply of fuel reserves underlying that part of the State. Nor were the claimants quieted when F. H. Hayden (1868), a geologist from the University of Pennsylvania, conducted a survey of the area and concluded that no coal beds thick enough for large mining operations existed above drainage level. Even after borings failed to reveal coal beds more than a foot or two thick, many people were still convinced that workable supplies remained to be discovered.

Meanwhile, several small drift mines were being operated, but each supplied only enough coal for local use until, in 1906, the Honey Creek mine was opened in a seam 66 cm (26 in) thick. Located about 6.4 km (4 miles) southwest of Peru in Nemaha County, this was one of the few commercial mines to be operated in the State. Coal thickness was as much as 91 cm (36 in), but, in parts of the mine, the seam was too thin to be extracted profitably. Mining operations continued for only a year or two.

The oldest coal beds in Nebraska are in the Cherokee and Marmaton Groups, which make up the upper part of the Des Moines Series. Because neither of these groups crop out, the existence of the coal beds in them is known only from the results of test drilling. At least nine coal beds have been identified at depths ranging from 305 m (1,000 ft) to 701 m (2,300 ft) below land surface. Burchett (1977, p. 60) estimated that 7.7×10^9 metric tons (8.5 billion tons) of deeply buried coal are in southeastern Nebraska.

As many as eight coal beds are known in the Wabaunsee Group, the youngest of the three groups composing the Virgil Series. One or more of these coal beds crop out in each of the following counties: Cass, Otoe, Johnson, Nemaha, Pawnee, and Richardson. From oldest to youngest, these coals are known as the Nodaway, White Cloud, Elmo, Wamego, Dry, French Creek, Lorton, and Honey Creek (Burchett, 1977), as shown in figure 7. All but the last are very thin to thin, ranging in thickness from 3 cm (0.1 ft) to 0.37 m (1.2 ft). As stated earlier, the Honey Creek Coal is known to be as much as 91 cm (36 in) thick in the mine of the same name. The Nodaway, Elmo, Wamego, Lorton, and Honey Creek coals have the greatest lateral extent but differ in thickness from place to place. All have been mined, but only the Honey Creek supplied more coal than that needed for a few households. Burchett (1977, p. 38–55) estimated that 9.25×10^9 metric tons (10.2 million tons) of coal in southeastern Nebraska has an overburden that is less than 15.2 m (50 ft) thick.

All coal in the Pennsylvanian rocks in the Nebraska part of the Forest City basin is black and is either bituminous or subbituminous. It has a medium to high sulfur content.

OIL AND GAS

Oil production from Pennsylvanian age rocks beneath Dundy, Frontier, Furnas, Harlan, Hayes, Hitchcock, Lincoln, and Red Willow Counties in southwestern Nebraska is from zones ranging in depth from 914 m (3,000 ft) to 1,372 m (4,500 ft) below land surface. These zones occur in the Cherokee, Marmaton, Lansing-Kansas City, Douglas, and Shawnee Groups. Of the 9.825×10^5 kl (6,180,500 bbl) of oil produced in Nebraska in 1976, about 3.46×10^5 kl (2,176,600 barrels) were from rocks of Pennsylvanian age.

Volumes of gas produced were not measured.

SHALE

Pennsylvanian shale formerly was used in the manufacture of brick and tile at three Nebraska locations—Table Rock in Pawnee County, Humboldt in Richardson County, and Nebraska City in Otoe County. At the Nebraska City plant, which was the most recent to discontinue operations, Wabaunsee Group shale was used also for producing lightweight aggregate. Currently, shale of the Lansing Group is being used in the manufacture of cement at a plant near Louisville in Cass County.

LIMESTONE

About 200 quarries, 26 of which were active in 1976, have been opened in limestone beds of Pennsylvanian age. Three-fourths of these quarries are in Cass County; the others are in Johnson, Otoe, Pawnee, Richardson, Sarpy, and Washington Counties. Of the currently operated quarries, 20 are in Cass County, 3 are in Sarpy, 1 is in Saunders, and 2 are in Washington. Total production from these four counties, in 1975, was 3.45×10^6 metric tons (3.8 million tons).

Seven companies operate the 20 active quarries in Cass County. Production is from the Kansas City, Lansing, Douglas, and Shawnee Groups. Limestone from two quarries operated by the same company is used in the manufacture of cement. Most of the production from the other quarries is used as crushed stone for concrete aggregate, asphalt aggregate, road surfacing, riprap, and agricultural lime. The remainder is used for wallstone or is finely ground for use as filter or as a feed supplement.

The three limestone quarries in Sarpy County are in the Kansas City Group. Their production is used for road surfacing, riprap, and wallstone.

The only limestone quarry in Saunders County is in the Lansing and Douglas Groups. Its production is used for concrete aggregate, asphalt aggregate, road surfacing, agricultural lime, riprap, and wallstone.

A single company operates both of the limestone quarries in Washington County. Production is from the Kansas City Group and is used for concrete aggregate, asphalt aggregate, road surfacing, agricultural lime, and riprap.

WATER

In areas where Pennsylvanian strata are the uppermost bedrock, water generally can be obtained from the overlying glacial drift or from stream alluvium. However, in some places, wells have been extended into the Pennsylvanian limestone and sandstone where the mantling deposits either are not water bearing or yield less water than needed. Yields from bedrock generally are small.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— Kansas

By WILLIAM J. EBANKS, JR., LAWRENCE L. BRADY, PHILIP H. HECKEL, HOWARD G. O'CONNOR, GEORGE A. SANDERSON, RONALD R. WEST, *and* FRANK W. WILSON

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—KANSAS

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ABSTRACT

Carboniferous rocks crop out only in the eastern one-third of Kansas, but they are present throughout the rest of the State in the subsurface. Because of their economic importance as sources of building materials, industrial and agricultural minerals, metallic ores, and oil and gas, and because of their importance to understanding the geologic history of the Carboniferous of North America, Mississippian and Pennsylvanian rocks have been the subjects of study in Kansas since very early in the exploration and settlement of the area. The first State Geological Survey of Kansas was established in 1864. Concerted effort to refine nomenclature and correlation of these rocks with the Carboniferous of other areas began in the early 1900's and is continuing at present.

Tectonic events that began in early Mississippian time resulted in the pattern of regional structures that is evident in these rocks now. Unconformities separate Mississippian carbonate rocks from underlying Devonian-Mississippian shale and from overlying lower-to-upper Pennsylvanian limestone, shale, and sandstone formations. The contact of the Pennsylvanian rocks with overlying Permian beds is conformable in Kansas.

Although only rocks of Osagean age crop out in southeastern Kansas, rocks of all four stages of the Mississippian occur in the Kansas subsurface, where they are as much as 500 m (1,700 ft) thick. Kinderhookian beds thicken northward, but Osagean and younger beds thicken southward from Kansas. Most of the Mississippian limestone and dolomite were formed in a shallow, marine-shelf environment. Oolitic and bioclastic limestone and dolomite associated with evaporites are common in the section. Shaly, sandy formations are common in the uppermost Mississippian.

Morrowan and Atokan rocks are the oldest Pennsylvanian units present in the Kansas subsurface. They are as thick as 335 m (1,100 ft) in southwestern Kansas. Surface exposures of Desmoinesian, Missourian, and Virgilian are found throughout eastern Kansas, where they comprise 49 formations whose combined thickness averages 750 m (2,460 ft). These middle and upper Pennsylvanian stages are also present in the subsurface.

The repeated occurrence of similar types of rocks in vertical section has led to recognition of cyclothems in the Pennsylvanian of Kansas. The "Kansas cyclothem" consists of five depositional units that record a single transgressive-regressive sequence of events.

The biostratigraphy of Kansas Carboniferous rocks, which is well known in general, is subject to better definition in detail. Correlation of Mississippian rocks has been made on the basis of studies of bryozoans, brachiopods, conodonts, and calcareous foraminifers. Pennsylvanian beds have been correlated mainly on the basis of fusulinid foraminifers and brachiopods. The Pennsylvanian-Permian boundary is still a subject of dispute.

Middle and Upper Pennsylvanian coal is an important natural resource in Kansas. Seventeen of the 42 coal beds in eastern Kansas have economic reserves of medium- to high-sulfur coal. Surface strip mining is the most common method of recovering these thin coal beds.

Oil and natural gas have been produced from the Carboniferous rocks of Kansas since the 1860's, and these formations continue to be the targets of exploratory drilling and sites of experimentation with enhanced recovery methods. Approximately 40 percent of the oil and 10 percent of the gas produced in Kansas have come from these formations.

Lead and zinc ore was an important product of the Mississippian rocks of Kansas until 1970, and there has been renewed interest in mining deep subsurface deposits. Limestone and shale from Middle and Upper Pennsylvanian Series have been used extensively for crushed stone, building stone, cement, and for different clay products in Kansas.

INTRODUCTION

Carboniferous rocks crop out only in the eastern one-third of Kansas (fig. 1), but they are present throughout the State in the subsurface. Mississippian (lower Carboniferous) rocks are divided into the Lower Mississippian Series, comprising the Kinderhookian and Osagean Stages, and the Upper Mississippian Series, comprising the Meramecian

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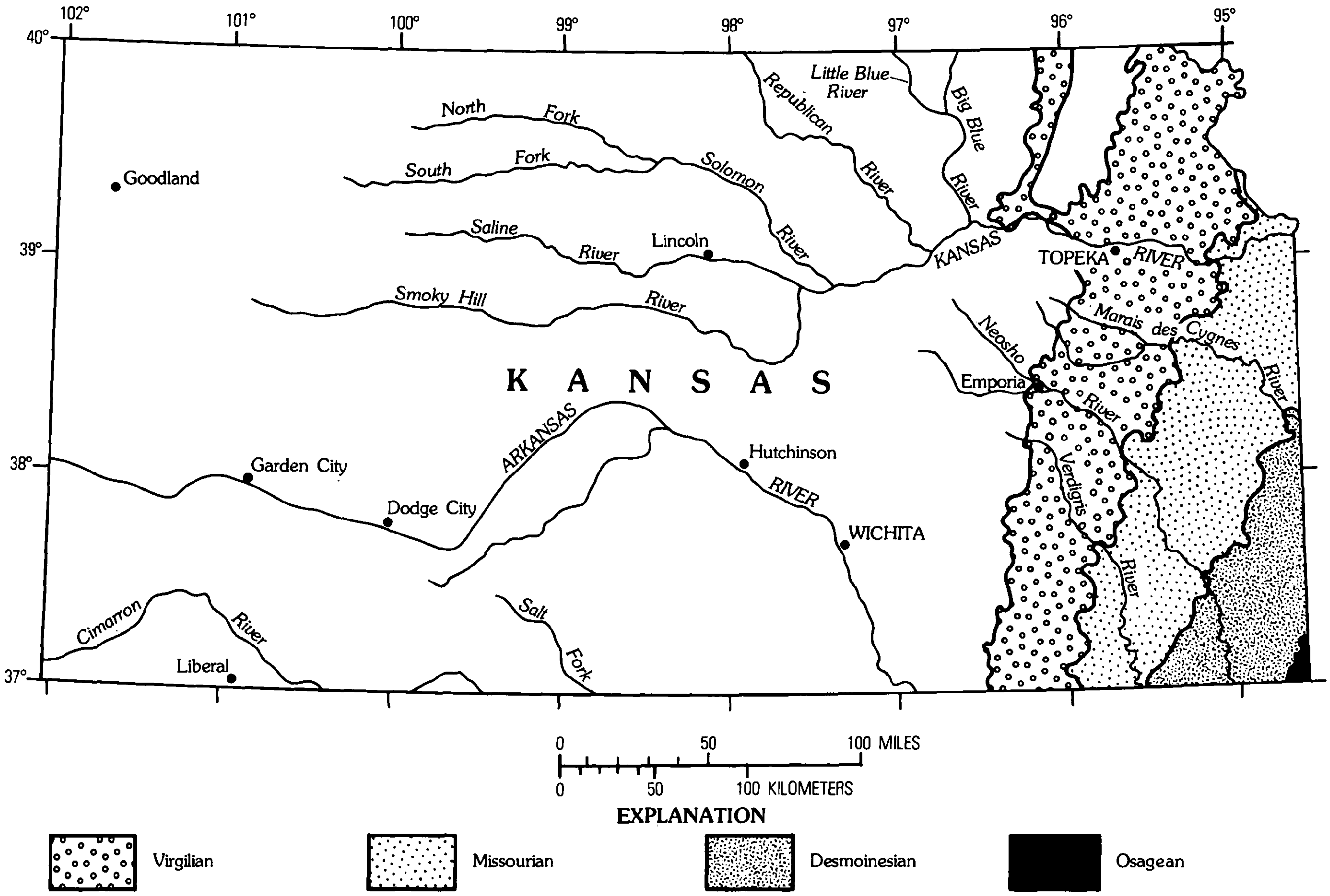

FIGURE 1.—Areas of outcrop of Carboniferous rocks in Kansas. North of the Kansas River and east of the Little Blue River, the surface is partly covered by Pleistocene glacial deposits.

and Chesterian Stages. Pennsylvanian rocks are divided into the Lower Pennsylvanian Series, comprising the Morrowan Stage, the Middle Pennsylvanian Series, comprising the Atokan and Desmoinesian Stages, and the Upper Pennsylvanian Series, comprising the Missourian and Virgilian Stages (fig. 2). Only the Upper and Middle Pennsylvanian and part of the Upper Mississippian are exposed at the surface.

In northeastern Kansas, in the area roughly bounded by the Kansas River on the south and the Big Blue and Little Blue Rivers on the west, the Carboniferous rocks are discontinuously overlain by as much as 100 m (330 ft) of Quaternary deposits, chiefly glacial till and eolian silt; nevertheless, Carboniferous rocks are exposed along most of the principal drainageways. This area coincides with the physiographic region known as the Dissected Till Plains of the Central Lowland.

In the remainder of the outcrop area, south of the Kansas River, Carboniferous rocks generally are overlain only by a thin soil cover, or, along valleys, by 15 m (50 ft) or less of fluvial deposits. The Carboniferous rocks have a slight regional west or northwest dip of about 2 to 6 m per km (10 to 30 ft per mi) in most of this area, which gives rise to gentle east-southeast-facing cuestas. This area of Carboniferous outcrops is part of the Osage Plains and the Cherokee Plain of the Central Lowland.

The effect of Cenozoic events on Carboniferous exposures has been, in eastern Kansas, a slight accentuation of relief and, in northeastern Kansas, the extensive covering of Carboniferous outcrops with glacial till, fluvial and lacustrine deposits, and discontinuous but extensive loess deposits. Loess deposits thin rapidly south of the Kansas River, and they are generally thin or absent in most of southeastern Kansas. Weathering of the Carboniferous rocks during the Cenozoic took place at about the same pace as erosion of the weathered materials, and throughout this area, soils are relatively thin.

The best natural exposures are along the streams

<table>
<tr><th></th><th>SYSTEM</th><th>SERIES</th><th>STAGE</th><th>GROUP</th></tr>
<tr><td rowspan="14">CARBONIFEROUS</td><td rowspan="10">PENNSYLVANIAN</td><td rowspan="6">UPPER</td><td rowspan="3">VIRGILIAN</td><td>WABAUNSEE</td></tr>
<tr><td>SHAWNEE</td></tr>
<tr><td>DOUGLAS</td></tr>
<tr><td rowspan="3">MISSOURIAN</td><td>LANSING</td></tr>
<tr><td>KANSAS CITY</td></tr>
<tr><td>PLEASANTON</td></tr>
<tr><td rowspan="3">MIDDLE</td><td rowspan="2">DESMOINESIAN</td><td>MARMATON</td></tr>
<tr><td>CHEROKEE</td></tr>
<tr><td>ATOKAN</td><td>UNNAMED</td></tr>
<tr><td>LOWER</td><td>MORROWAN</td><td>UNNAMED</td></tr>
<tr><td rowspan="4">MISSISSIPPIAN</td><td rowspan="2">UPPER</td><td>CHESTERIAN</td><td rowspan="4">UNNAMED</td></tr>
<tr><td>MERAMECIAN</td></tr>
<tr><td rowspan="2">LOWER</td><td>OSAGEAN</td></tr>
<tr><td>KINDER-HOOKIAN</td></tr>
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FIGURE 2.—Subdivisions of the Carboniferous in Kansas.

where the Carboniferous rocks are being actively eroded. Many excellent artificial exposures are found along the newer highways, particularly those trending east-west, and in pits and quarries where limestone, shale, sandstone, or coal have been extracted.

Both Mississippian and Pennsylvanian Systems are extremely important in the economy of Kansas as sources of fossil-fuel resources, metallic ores, agricultural minerals, building stone, and ceramic raw materials. Ground-water resources in rocks of the Upper Pennsylvanian Series are important only locally in eastern Kansas. Because of this economic importance, study of Carboniferous rocks began early in the exploration and settlement of eastern Kansas, and this study is continuing today.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Kansas Geological Survey.

HISTORY

Explorers, traders, missionaries, and others entering the eastern Kansas Territory in the early 1800's noticed coal-bearing strata there, especially along the route of the Santa Fe Trail and in the drainage tributaries through which they passed. The earliest purposeful stratigraphic studies of Carboniferous rocks in eastern Kansas were begun by G. C. Swallow (1855), the first State Geologist of Missouri, as he extended his study of the geology of Missouri along the west side of the Missouri River into what is now northeastern Kansas.

Swallow also became involved in a dispute over priority for the first recognition of Permian rocks, not only in Kansas, but also on the whole of the North American Continent. The Permian was, at that time, considered to be part of the Carboniferous. Major Frederick Hawn, a military surveyer and geologist stationed at Fort Leavenworth, submitted fossils that he had collected in the course of his work in Kansas to Swallow, and the two, on the basis of Swallow's recognition of the Permian age of the specimens, reported their findings in 1858, only a few days before an announcement of a similar finding by F. B. Meek (Meek and Hayden, 1858) which also was based on fossils submitted to him by Hawn. These early studies had economic implications because of the known occurrence of salt and gypsum in beds of Permian age and the relation of these younger beds to underlying Coal Measures, which were known to be widespread in the midcontinent area, even at that time.

Interest in economic development of the young State of Kansas led the Kansas legislature in 1864 to authorize the first State Geological Survey of Kansas; Benjamin F. Mudge was State Geologist. The enacting law provided for certain very specific assignments, most of which related to assessment of potentially valuable economic minerals and the suitability of Kansas soils for agriculture. Although this first State Survey was little more than a general reconnaissance, the geologists of that day had a good knowledge of the general stratigraphy of eastern Kansas, most of which involved rocks of late Carboniferous age. This first report included a review of the occurrence of coal, lignite, lime, marble, cement, gypsum, alum, salt, sandstone, lead, zinc, iron, and scattered surface indications of oil. The second State Geological Survey of Kansas began in 1865; G. C. Swallow was the second State Geologist. This work was largely an extension of the fieldwork begun by the Mudge Survey.

On the basis of Swallow's studies of the geology in the extreme southeastern corner of the State, the rocks there were for the first time correctly correlated with the lower Carboniferous "Mississippian" and the lead-bearing strata of southwestern Missouri. In the coal-bearing or upper Carboniferous "Pennsylvanian" rocks, Swallow identified 22 different seams of coal ranging in thickness from 0.3 to 2.1 m (1 to 7 ft). The "Report of the Geological Survey of Miami County, Kansas" (Swallow and Hawn, 1865), was the first geological report actually published by the initial two Geological Surveys in Kansas. The first map of the geology of Kansas was produced by Mudge (1875), in his capacity as a geologist on the State Board of Agriculture. During the period 1866–89, there was no official Geological Survey in Kansas, but work on the upper Carboniferous was proceeding (Broadhead, 1881; Hay, 1887).

After years of development in southwestern Missouri, important deposits of lead and zinc ore were finally discovered in rocks of Mississippian age in southeastern Kansas, near Galena, in 1876. Erasmus Haworth, in 1884, completed a study concerning the geology of Cherokee County in the area of this lead-zinc-mining activity. Continued interest in the economy of the developing State led the Kansas Legislature, in 1889, to pass an appropriation act for the University of Kansas which provided for the establishment of a Geological Survey. In 1895, the University formally established the University Geological Survey of Kansas.

Haworth, who was named to supervise the physical geology and mineralogy division of the new survey, had begun earlier to put students into the field to compile stratigraphic sections along the major streams in southeastern Kansas. The emphasis of this early work was in basic stratigraphy and paleontology to define the areal boundaries of the major geologic subdivisions. The preliminary results of these studies were published in a timely fashion in various issues of the "Kansas University Quarterly" (Haworth, 1894, 1895). A summary of this work, the first comprehensive stratigraphic description, correlation, and section of the Kansas Carboniferous, was published in Volume I, Kansas Geological Survey (Haworth, 1896). Continued refinement, including more detailed description and correction of earlier miscorrelations, was made by these early workers in the process of studying Carboniferous rocks in eastern Kansas and the economically important coal and deposits of oil and gas which they contained (Haworth, 1898). The lead- and zinc-mining activity in extreme southeastern Kansas (Haworth and others, 1904) continued to attract attention. The U.S. Geological Survey (Smith and Seibenthal, 1907) published a folio on the Joplin district of Missouri and Kansas, which included discussions of formation names and ages of important Mississippian limestone and chert units that were being mined.

In 1916, Raymond C. Moore became State Geologist of Kansas. The first report issued under his direction (Moore and Haynes, 1917) was a review of the oil and gas resources of Kansas. Moore's appointment marked the beginning of a period of concerted effort in the refinement of the correlation and nomenclature of Pennsylvanian rocks in Kansas, and of integration of these studies of Kansas formations with those of surrounding States (Moore, 1920, 1929, 1931, 1932a, b, 1933). In 1936, Moore published a stratigraphic classification of the Pennsylvanian rocks of Kansas. The work of many other geologists, whose work is cited by Moore, contributed to this report. This report included a review of early studies of the Pennsylvanian and a redefinition of the Pennsylvanian-Permian boundary, a considerable revision of the previous classification and nomenclature of the Pennsylvanian rocks of eastern Kansas, recognition of work in surrounding States, and an introduction of the concept of repetitive or cyclic sedimentation.

During the 1930's and early 1940's, Moore and other colleagues (Newell, 1935; Abernathy, 1937; Pierce and Courtier, 1938; and Jewett, 1933, 1941, 1945) extended their efforts to refine the classification of middle and upper Pennsylvanian rocks farther south and to integrate work on paleontology and stratigraphy of the Pennsylvanian in adjacent States with that in Kansas. The results of this work were summarized in reports by Moore (Moore, 1948, 1949).

Subsequent work has resulted in the reclassification of several groups and subgroups and in the presentation of a stratigraphic classification that is more acceptable for regional correlations (Moore and Mudge, 1956). Work also continued in the refinement of the Pennsylvanian-Permian boundary, which culminated in a publication by the U.S. Geological Survey (Mudge and Yochelson, 1962). Other stratigraphic work, designed to be more comprehensive and more detailed than earlier studies (Howe, 1956; O'Connor, 1963; Ball, 1964; Jewett and others, 1965), has resulted in other changes in the classification of Upper and Middle Pennsylvanian units. This work continues with recent studies of environments of deposition and of details of stratigraphy of the many limestone, shale, and sandstone units of the Pennsylvanian in surface exposures and in the subsurface.

Although oil had been discovered in southeastern Kansas Middle Pennsylvanian rocks in the 1860's, few wells were drilled below the top of the Mississippian limestone, because this was considered to be the lower limit of potential production. As additional drilling continued, however, more and more wells penetrated the Mississippian, and work was initiated to map the extent of recognizable Mississippian formations in the subsurface of eastern Kansas (Lee, 1939; Lee and Girty, 1940). Earlier work by Moore (1928) on the Mississippian of Missouri strongly influenced the assignment of formation names to Mississippian rocks in Kansas. Two parallel sets of terminology have evolved for the Mississippian rocks in the small area of outcrop in southeasternmost Kansas. One is based on the similarity of these rocks with those in northern Arkansas and Oklahoma (Smith and Siebenthal, 1907; McKnight and Fischer, 1970); the other is based on age equivalency and similarity of these rocks with those of northern Missouri and the Mississippi River valley (Moore, 1928; Kaiser, 1950). The latter has prevailed in later work on the Mississippian in the subsurface of Kansas. Studies of the Mississippian in the subsurface have resulted in the application of these formation names farther west (Clair, 1948; Goebel, 1966, 1968a, 1968b) and in the refinement of the ages of these rocks (Girty, 1940; Thompson and Goebel, 1963, 1968; Goebel, 1967). Studies of the Lower Pennsylvanian and Mississippian rocks in central and western Kansas are continuing today.

GEOLOGIC SETTING

The relationship of the Carboniferous rocks in Kansas to older underlying rocks is indicative of the eventful history of this sequence of rocks. Many tectonic events contributed to the fact that beds as old as Early Mississippian and as young as Late Pennsylvanian are in contact with underlying lower Paleozoic or Precambrian rocks.

Over much of the eastern two-thirds of Kansas, the dark Mississippian-Devonian Chattanooga Shale is in unconformable contact with older rocks. This unit is overlain unconformably by Mississippian carbonate rocks and by Pennsylvanian rocks. The importance of a northern Kansas basinal area and a southeastern Kansas archlike positive area, which existed until Early Mississippian time, is apparent from the distribution and thickness of rocks deposited before that time (Jewett, 1951).

Tectonic events that began early in the Mississippian resulted in the pattern of structures that are apparent in these rocks at present (fig. 3). Most notably, the central Kansas uplift evolved from an older, more subtle, archlike feature in central and northwestern Kansas, and the Nemaha uplift began in eastern Kansas in Late Mississippian time as a strongly positive elongate feature, which today extends from eastern Nebraska to southern Oklahoma (Jewett, 1951).

The continued activity on these positive structural elements and the increased importance of the Anadarko basin in Oklahoma strongly affected patterns of sediment distribution and areas of erosion throughout the Carboniferous. Not until Late Pennsylvanian time did sediments finally cover all these underlying structures (Lee and Merriam, 1954).

Lower Mississippian carbonate rocks, where the basal Chattanooga Shale is absent, rest unconformably on all older deposits. An important period of erosion removed these beds from the Nemaha and central Kansas uplifts near the close of Mississippian time. Upper Mississippian (Chesterian) beds, which are present only in the subsurface of southwestern and southeastern Kansas, are probably unconformable with older rocks, but the nature of their contact with overlying Lower Pennsylvanian, Morrowan, beds is unclear. In some areas, this contact surely is unconformable, but in others, it may represent continuous, but varied, sedimentation or a long period of almost no deposition.

Sedimentation during the Pennsylvanian Period covered progressively larger areas, resulting in overlap of older stratigraphic units by younger ones until areas of uplift were covered. The Missourian

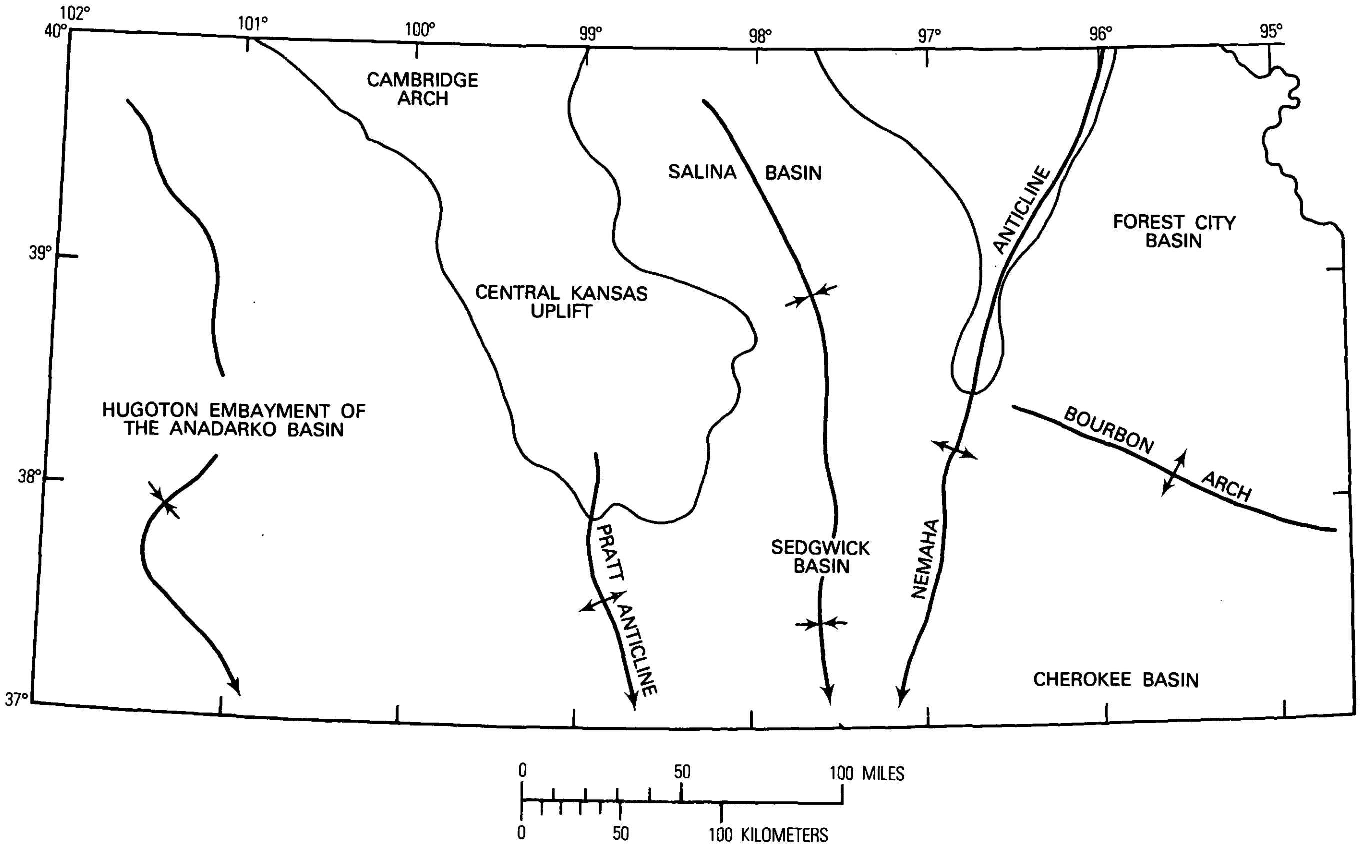

FIGURE 3.—Paleotectonic features of Kansas that were formed in Late Mississippian–Early Pennsylvanian time (modified from Stewart, 1975, fig. 19).

Kansas City Group is in direct contact with Precambrian crystalline rocks over parts of these old positive areas. Considerable local relief was present on the Precambrian terrane, which has been discovered by drilling into buried high areas on the ancient rocks beneath the Pennsylvanian cover (Walters, 1946).

The close of Carboniferous sedimentation in Kansas was not marked by any important tectonic event. The contact of Pennsylvanian with younger Permian beds is conformable, except for local areas on the Nemaha uplift, which may have been the sites of intermittent structural movement and of channel-cutting by terrestrial streams at about this time.

The Pennsylvanian-Permian contact has been chosen by convention to be the top of the Brownville Limestone Member of the Wood Siding Formation of the Virgilian Stage (Moore, 1936). No important discontinuity in the fossil record and no important tectonic event mark this transitional boundary in Kansas. Some States, notably Oklahoma, disagree on the placement of this systemic boundary; in fact, it has been a subject of dispute for a long time (Moore, 1936; Branson, 1962; Mudge and Yochelson, 1962).

After the Carboniferous, no large-scale structural movements have taken place in Kansas, other than regional tilting to the west and overall uplift, which have affected mainly the distribution and character of post-Permian rocks (Merriam, 1963).

LITHOSTRATIGRAPHY

MISSISSIPPIAN

In Kansas, rocks of Mississippian age are exposed at the surface only in the southeastern corner of the State (fig. 1). Consequently, very little detailed mapping of Mississippian formations has been accomplished except at a local level, such as in the Kansas part of the Tri-State mining district. Beds representing all four stages of the Mississippian have been recognized in the Kansas subsurface, and they are as much as approximately 500 m (1,700 ft) thick in southwestern Kansas. Thinner sections of Mississippian rocks are present in most other areas of the State except over the crests of the major uplifts, where Mississippian rocks are absent (Goebel and Stewart, 1978).

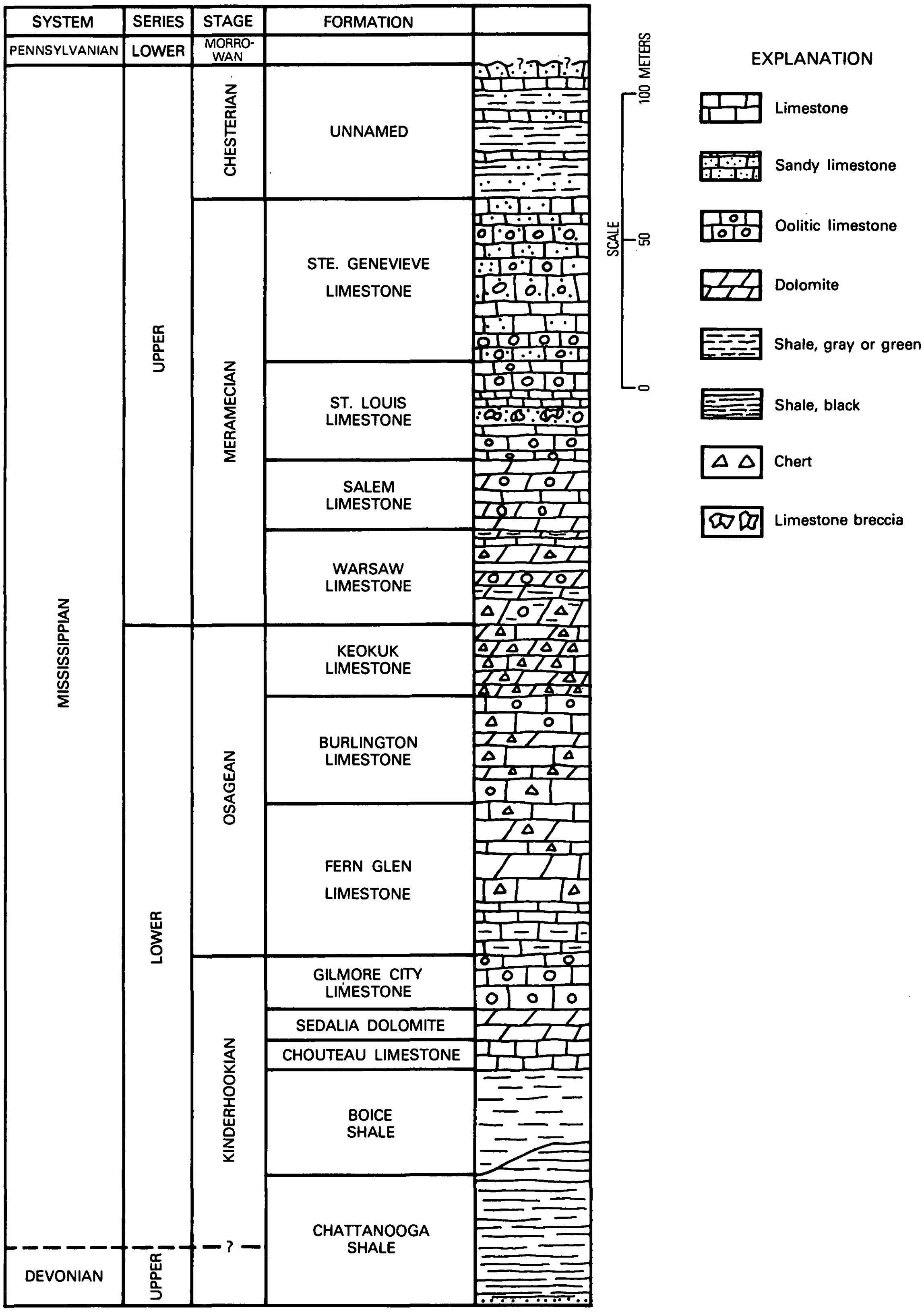

FIGURE 4.—Generalized lithostratigraphic column of Mississippian formations in Kansas (modified from Goebel, 1968b). Thicknesses represented are averages.

The earliest attempts to map the lithologic units within the Mississippian over large areas in Kansas were based on studies of well cuttings and cores (Lee and Girty, 1940). As more data became available, formation names from type areas in Missouri and Iowa were applied to Kansas rocks; however, because these formations in some areas of Kansas are different from the type sections, a rather loose system of names using partly rock-stratigraphic and partly time-stratigraphic terms or letter-designated subdivisions is in use by economic geologists.

The following brief discussion is based on the formal classification recognized by the Kansas Geological Survey, as derived from the works of Lee and Girty (1940) and Lee (1943, 1956), Moore (1957), and Goebel (1966, 1968a, 1968b) (fig. 4).

The contact between the Devonian and the Mississippian Systems is not exposed at the surface in Kansas. It is placed within the Chattanooga Shale, which is present beneath most of eastern and central Kansas. The black Chattanooga Shale and the overlying greenish-gray Boice Shale are more than 75 m (250 ft) thick in northeastern Kansas, but they are not present on the northern end of the Nemaha uplift or in the southeastern corner of Kansas. The Chattanooga Shale is exposed at the surface beneath Mississippian limestone in southwestern Missouri and nearby in Oklahoma.

Kinderhookian limestone and shale are present in the subsurface of eastern Kansas, and successively younger formations overlap older ones northwestward, the formations thickening northward into Nebraska and Iowa. These rocks are absent in the area of outcrop in southeastern Kansas.

In contrast, Mississippian rocks of Osagean and younger age thicken southward, and they overlie Kinderhookian beds with angular unconformity on a regional scale. Osagean rocks consist mainly of coarse-grained crinoidal and finer grained, mixed-fossil, bioclastic limestones, some fine-grained dolomite, and dolomitic limestone, all of which are usually cherty to some degree. In south-central Kansas, in the subsurface, a facies change takes place, and the rocks of equivalent and younger age are reddish and greenish, crinoidal, thin-bedded limestone interbedded with green or gray shale.

Almost all the outcropping Mississippian rocks in southeastern Kansas belong to the Keokuk Formation and have been assigned an Osagean age on the basis of studies of conodonts (Thompson and Goebel, 1968). These beds are cherty limestone and dolomite, fossiliferous limestone, and chert in beds and nodules. One thin oolitic limestone, the "Short Creek Oolite," near the top of the Keokuk, has been mapped in local areas to define structure of the ore-bearing Mississippian in the Tri-State District.

In southeastern Kansas and contiguous areas of Missouri and Oklahoma, the outcropping Osagean and Meramecian rocks, which are estimated to be 90–135 m (300–450 ft) thick, have been designated the "Boone Formation," from studies of rocks in northern Arkansas. Different authors have informally subdivided the "Boone" into 16 beds designated by letters of the alphabet (Fowler and Lyden, 1932), or into seven named members (McKnight and Fischer, 1970), but there is disagreement on where in the section to place the Osagean-Meramecian contact. The Kansas Geological Survey recognizes the "Boone Formation" only as an informal term.

As noted above, rocks of the Upper Mississippian Series may be in disconformable contact with older Mississippian rocks in southeastern Kansas, but elsewhere in the State, this unconformity is not present or is obscure. Beds of the Meramecian and Chesterian Stages, which are only in the subsurface, are themselves separated by an unconformity of regional importance.

Meramecian rocks are as thick as 260 m (850 ft) in southwestern Kansas, but these units are much thinner or are absent in northern Kansas and on the crests of pre-Pennsylvanian structures (Goebel, 1966). Lower Meramecian beds consist of shaly, cherty dolomite and interbedded thin limestone beneath oolitic and bioclastic, slightly cherty limestone. Upper Meramecian beds are silty and sandy fossiliferous or oolitic limestone.

Rocks of the Chesterian Stage in Kansas are confined mainly to southwestern and southeastern parts of the State (Lee and Girty, 1940; Goebel, 1966; Nodine-Zeller and Thompson, 1977). Bioclastic limestone near the outcrop in the southeast is much better known at localities in adjoining States. In the southwest, beds of fine-grained sandy limestone and shaly crinoidal limestone are interbedded with green or gray shale. These beds are more than 100 m (300 ft) thick in southwestern Kansas near the Oklahoma line.

These descriptions of stratigraphic units of Mississippian age in Kansas are based on presently available data. Additional studies, in progress, suggest that some of the lithofacies in western Kansas may be somewhat younger, by as much as a stage, than they were thought to be previously. Environments of deposition of the Mississippian have been described earlier only in very general terms. Later work (Ball, 1966; Goebel, 1968b; Ebanks and others, 1977) has demonstrated that there is much room for expansion and clarification of knowledge about these subsurface rocks, especially because of their economic importance.

PENNSYLVANIAN

Outcrops of Pennsylvanian rocks are widespread in eastern Kansas (fig. 1), and these serve as a standard of reference in the study of deposits of equivalent age in other parts of the Continent. The system in Kansas is divided into Lower, Middle, and Upper Pennsylvanian Series, which comprise five stages, in ascending order: Morrowan, Atokan, Desmoinesian, Missourian, and Virgilian.

LOWER AND MIDDLE PENNSYLVANIAN

Morrowan and Atokan Stages.—Rocks of Morrowan age in Kansas are thought to be restricted to the Hugoton Embayment. Atokan rocks, likewise, are restricted mainly to the subsurface of southwestern Kansas, but some Atokan shale may be present beneath Desmoinesian rocks in southeastern and northeastern Kansas subsurface areas (Stewart, 1975; Nodine-Zeller and Thompson, 1977). Rocks of the Morrowan stage in the subsurface of southwestern Kansas comprise approximately 185 m (600 ft) of shale, limestone, and sandstone. These rocks have been described as the Kearny Formation (Thompson, 1944), and trends in lithostratigraphy have been mapped (McManus, 1959). They are overlapped in western Kansas by rocks of Atokan age, but there is disagreement about the extent of the Atokan and Morrowan rocks in southwestern Kansas (Rascoe, 1962; Stewart, 1975). Rocks of Atokan age in Kansas consist of interbedded dark-gray, black, and dark-brown cherty limestone and dark-gray to black shale, which form a sequence as thick as 150 m (500 ft). The Atokan rocks, where present, are in gradational contact with overlying Desmoinesian rocks.

MIDDLE AND UPPER PENNSYLVANIAN

The Pennsylvanian System on outcrop in eastern Kansas comprises 49 formations that have been divided into 129 formally named members and aggregated into 8 groups and 3 stages of the Middle and Upper Pennsylvanian Series (fig. 5). Middle Pennsylvanian rocks are exposed in southeastern Kansas and belong entirely to the Desmoinesian Stage. A complete sequence of Upper Pennsylvanian rocks is well exposed in eastern Kansas, particularly in the Missouri River valley along the northeastern border and south of the glacial limit at the Kansas River valley, where it serves as the type region for the Missourian and Virgilian Stages of this series. The following summary is derived from Moore (1949), Moore and others (1951), O'Connor (1963), Jewett and others (1968), and from observations by P. H. Heckel.

Middle Pennsylvanian, Desmoinesian Stage.—Rocks of the Desmoinesian Stage are divided, in ascending order, into the Cherokee and Marmaton Groups, which together have an aggregate thickness of 180 to 190 m (600 to 625 ft).

The Cherokee Group rests unconformably on Mississippian carbonate rocks and consists largely of shale and subordinate sandstone, coal beds, and thin limestone beds. Although some cyclothems based on coals have been recognized in the Cherokee, difficulty of mapping has led more recent workers to subdivide the Cherokee into just two formations, Krebs and Cabaniss; only a few well-exposed or persistent sandstones and limestones receive formal member names. The Cherokee Group ranges from 100 to 150 m (325 to 500 ft) in thickness.

The Marmaton Group lies conformably upon the Cherokee. In contrast to the Cherokee, it contains mappable, laterally persistent limestones alternating with laterally persistent sandy shales, which provide the basis for subdivision into eight formations (fig. 5). Each limestone formation is subdivided into members on the basis of persistent thin shales, and the lower three shale formations each contain a formally named local sandstone member. Each limestone and shale formation is usually 6–10 m (20–35 ft) thick, and the Marmaton Group has a total thickness of about 75 m (250 ft).

Upper Pennsylvanian, Missourian Stage.—Rocks of the Missourian Stage, the lower subdivision of the Upper Pennsylvanian Series, are divided into three groups, in ascending order, Pleasanton, Kansas City, and Lansing. Total thickness averages about 200 m (650 ft).

The basal Pleasanton Group rests unconformably on Desmoinesian rocks of the upper Marmaton Group. It is 9–40 m (30–130 ft) thick and consists mainly of shale, but locally thick sandstone is present at the base, and minor sandstone, limestone, and coal, above. Three constituent formations are recognized only in southernmost Kansas, where the Checkerboard Limestone, an important marker unit in Oklahoma, is present between the two terrigenous detrital formations of the Pleasanton.

The Kansas City Group conformably overlies the Pleasanton Group. Like the Marmaton Group, it con-

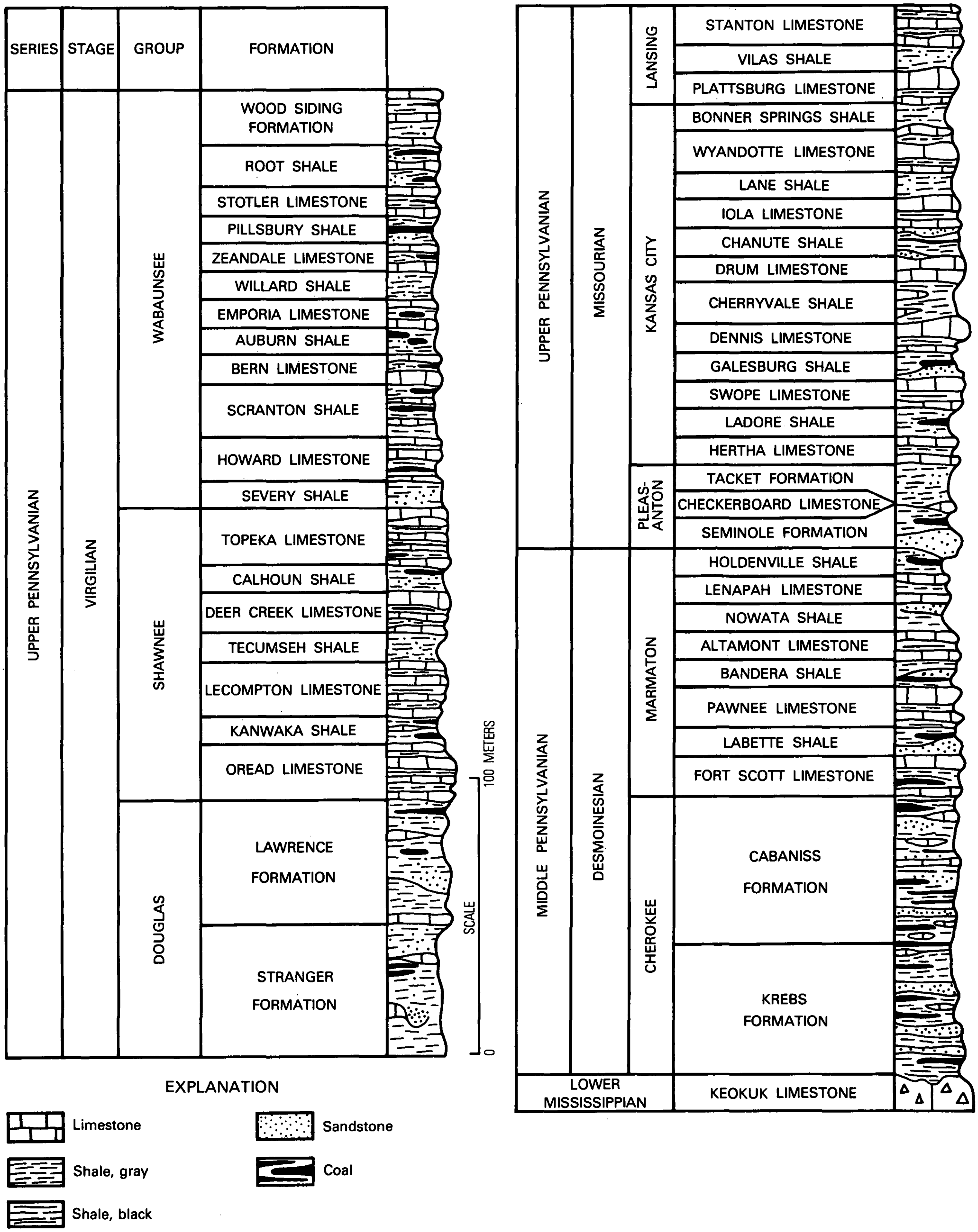

FIGURE 5.—Generalized lithostratigraphic column of Pennsylvanian formations at the surface in eastern Kansas (adapted from Zeller, 1968, pl. 1). Names of members are omitted, but nature of members is indicated by lithic symbols within formations. Thicknesses are only approximate; limestone, black shale and coal are generally expanded at the expense of gray shale and sandstone.

sists of alternating mappable, laterally persistent limestones and sandy shales, which provide the basis for division into 12 formations (fig. 5). All limestone formations are subdivided into members, mainly on the basis of thin but laterally persistent shales. Certain of the shale formations contain formally named sandstone and limestone members, and thin coal beds. Maximum thickness of the six limestone formations ranges from 10 to 30 m (35–100 ft). Three of these limestone formations disappear southward toward the Oklahoma border (Hertha, Swope, Wyandotte); the other three thin to about 2 m (6 ft) southward at the border. The intervening shale formations range from minimum thicknesses of 1–4 m (3–13 ft), in northeastern Kansas and over thickened facies of the underlying limestones, to 15–60 m (50–200 ft) farther south over thinner facies of the underlying limestones, particularly southward toward the Oklahoma border. Total thickness of the Kansas City Group is about 107 m (350 ft) in east-central Kansas.

The Lansing Group conformably overlies the Kansas City Group. It consists of three formations (fig. 5), which essentially continue the alternation of limestone and shale formations from the Kansas City Group. Both limestone formations are divided into members on the basis of persistent thin shales, and both formations attain maximum thicknesses of 25–36 m (80–120 ft) in southeastern Kansas before thinning southward as they grade into shale and sandstone near the Oklahoma border. The intervening shale formation ranges in thickness from 1 m (3 ft) or less over thick limestone (Plattsburg) facies up to 30 m (100 ft) over thin Plattsburg facies. Total thickness of the Lansing Group ranges from about 24 m (80 ft) in northeastern Kansas to about 60 m (200 ft) in southeastern Kansas.

Two thick Missourian shale and sandstone formations in the Kansas-Oklahoma border region (Coffeyville; Wann) are correlated with several Kansas formations (Tacket through Galesburg; Lane through Stanton, respectively), where intervening limestones pinch out near the Kansas-Oklahoma border.

Upper Pennsylvanian, Virgilian Stage.—Rocks of the Virgilian Stage, the upper subdivision of the Upper Pennsylvanian Series, have an aggregate thickness of about 360 m (1,200 ft) and are divided into three groups, in ascending order, Douglas, Shawnee, and Wabaunsee.

The Douglas Group (fig. 5) consists predominantly of shale and sandstone and subordinate thin limestone beds and coal beds. The group ranges in thickness from 72 m (240 ft) in northeastern Kansas, where sandstone is only locally prominent, to 120 m (400 ft) southward near the Oklahoma border, where sandstone occupies a greater part of the thickness. The Douglas is divided at the most laterally persistent limestone unit into two subequal formations; each formation contains two shale members, two thin limestone members, and a thick local sandstone member. As presently recognized, the Douglas includes the former Pedee Group, rocks that were thought to lie unconformably beneath the overlying part of the Douglas. Stratigraphic relations within the Douglas, as it is presently defined, and between the Douglas and the underlying Lansing Group of the Missourian Stage are now considered to be essentially conformable on a regional scale.

The Shawnee Group is about 100 m (330 ft) thick and conformably overlies the Douglas Group. Like the Marmaton, Kansas City, and Lansing Groups below, it consists of alternating dominantly limestone and dominantly sandy shale strata, which provide the basis for division of the Shawnee into seven formations (fig. 5). Each of the limestone formations is subdivided into at least five members on the basis of thin, laterally persistent shales. These limestone formations range from 6 to 30 m (20–100 ft) in thickness, with an average thickness of 10–20 m (33–66 ft). The shale formations are as thick as 15–45 m (50–150 ft), but only one of them is divided into members. Only the lowermost shale formation shows increased thickening toward Oklahoma like the shale formations below; the upper two shale formations attain maximum thickness in northeastern Kansas and thin southward as well as northward.

The Wabaunsee Group has a thickness of about 150 m (500 ft). It conformably overlies the Shawnee Group and caps the Virgilian Stage and Pennsylvanian System in Kansas. It consists of a greater proportion of sandy shale, containing several thin coal beds, than does the underlying Shawnee Group, but it contains enough persistent thin limestones to allow subdivision into 12 formations (fig. 5). As in several groups below, each limestone formation is subdivided into members on the basis of laterally persistent shales, but these shale members tend to be thicker and more sandy than most of those in limestone formations lower in the section. Some of the shale formations are similarly subdivided on the basis of thin persistent limestones. Limestone formations generally range from 2 to 12 m (7 to 40 ft) in thickness. The shale formations typically are thicker than 10 m (35 ft), and one attains a thickness of nearly 40 m (130 ft). Lower Permian rocks overlie the top of the Wabaunsee Group (Brownville Limestone Member of Wood Siding Formation) with apparent conformity in most places.

FACIES AND DEPOSITIONAL ENVIRONMENTS OF PENNSYLVANIAN ROCKS

Cyclothems have long been recognized in the Middle and Upper Pennsylvanian sequences of eastern Kansas (Moore, 1936, 1949, 1950; Weller, 1958). Moore recognized the basically transgressive-regressive sedimentation responsible for formation of simple cyclothems in the Cherokee and Wabaunsee Groups, in which largely nonmarine sandy shale containing coal alternates with marine shale and limestone, allowing relatively straightforward interpretation of depositional environments. Moore also devised a hierarchical classification of cyclothems of the Wabaunsee type (which became viewed essentially as shale-limestone couplets) grouped to form megacyclothems, or complex but distinctive successions of different shale-limestone couplets, in the Marmaton, Kansas City, Lansing, and Shawnee Groups. Megacyclothems in these groups are nucleated around the limestone formations containing the thin shale members, and the megacyclothem boundaries lie within the intervening sandy shale formations.

More recent work, summarized by Heckel and Baesemann (1975) and Heckel (1977), has shown that only the middle part of the megacyclothem—specifically the "middle" and "upper" limestone members and intervening thin shale member that typically contains a phosphatic black shale facies—occurs commonly enough throughout these groups to have basic genetic significance. Each depositional unit, or "Kansas cyclothem," presently recognized in these groups is nucleated around the limestone formation; it records a single transgressive-regressive marine sequence, consisting, in ascending order, of (1) thick nearshore sandy ("outside") shale (top of underlying shale formation); (2) thin transgressive ("middle") limestone; (3) thin offshore ("core") shale (often containing phosphatic black facies); (4) thick regressive ("upper") limestone; and (5) thick sandy ("outside") shale again (base of overlying shale formation) (fig. 6). The older usage of "cyclothem" in the Marmaton through Shawnee Groups for a shale-limestone couplet within a megacyclothem is abandoned, and the term, "megacyclothem," is applied only to the concept of more complex sequences.

Facies and depositional environments are considered first for the Marmaton through Shawnee Groups within the framework of the basic Kansas cyclothem (Heckel, 1977), with comments on major facies changes (fig. 7) observed along the Midcontinent outcrop belt extending southward into Oklahoma and northward into Missouri, Iowa, and Nebraska.

Nearshore shales.—Nearshore shales comprise mainly those formations that alternate with limestone formations in the cyclic sequence. The fact that these shale formations lie "outside" the "bundle" of limestones and thin shales that constitute the limestone formations caused these shale parts (or "members") of the cyclothem to be termed "outside shales," in a positional sense, before their depositional significance was thoroughly established.

Nearshore shale formations are typically sandy; though variable in thickness, they are usually thick, often attaining 15 m (50 ft) and locally 40 m (130 ft) in thickness. Commonly, they contain thin layers of siltstone and sandstone, which carry macerated plant fragments and only a sparse marine faunal assemblage of low diversity. They locally contain deposits that are demonstrably nonmarine, such as coal and underclay, shale containing well-preserved land-plant fossils, and channel sandstones, all of which lack marine fossils. Outside shales are the units within which Wanless and his coworkers (1970) have mapped many deltaic sequences.

A depositional model in which abundant terrigenous detritus was deposited in a shallow sea by prograding, laterally migrating lobes of a delta readily accounts for the characteristics of these shale formations. Variability in thickness reflects the local extent of each delta lobe. The nonmarine deposits record the subaerial deltaic plain. Rocks containing sparse marine fossils of low diversity record prodelta to delta-front environments where rapid deposition, increased turbidity, and fluctuating salinity reduced the abundance and diversity of marine organisms.

In southeastern Kansas, nearshore shale formations constitute a proportionately greater amount of the total section, and they thicken substantially in the Kansas-Oklahoma border region, in the direction of a major deltaic detrital source farther south in Oklahoma. The only exceptions to this are places in which these shales thin over thickened facies of underlying limestone.

Most nearshore shale formations tend to thin northward into Iowa and Nebraska, away from the major directions of detrital influx farther south in

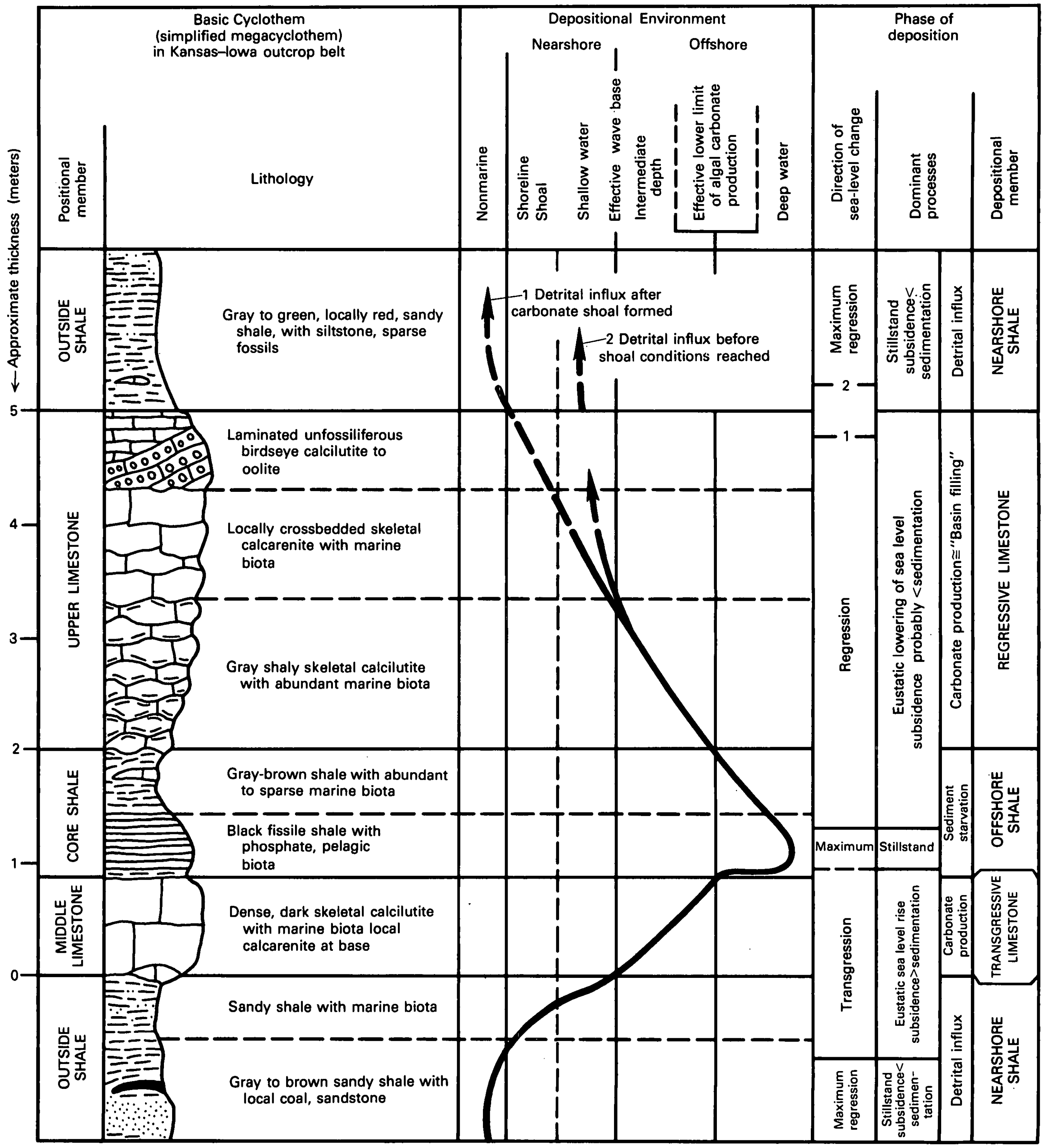

FIGURE 6.—Basic vertical sequence of an individual Kansas Pennsylvanian cyclothem, which is generally characteristic of Marmaton, Kansas City, Lansing, and Shawnee Groups, showing lithology and interpreted environments and phases of deposition. Terms on left for cyclothem members describe position in cyclothem; "middle" and "upper" for limestones derive from megacyclothem classification of Moore (1936, 1949); "outside" and "core" for shales derive from analysis by Heckel and Baesemann (1975). Terms on right for cyclothem members describe phase of deposition and are preferable when environments are reasonably well established (modified from Heckel, 1977, fig. 2).

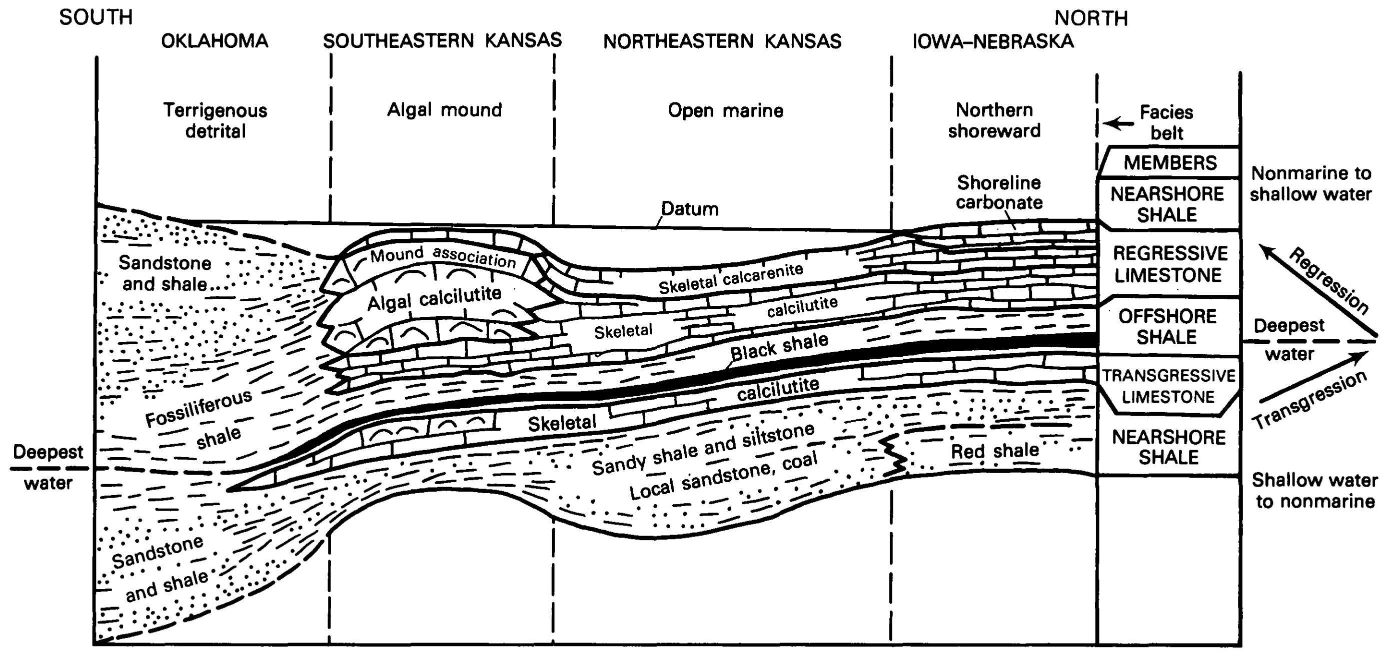

FIGURE 7.—Basic pattern of lateral facies relations in generalized Kansas cyclothem across facies belts exposed along Midcontinent outcrop. Datum is interpreted approximate sea level at time that increased detrital influx terminated deposition of regressive limestone member (modified from Heckel, 1977, fig. 4).

Oklahoma. Some of these shales become more abundantly and diversely fossiliferous, reflecting less detrital influx with its attendant unstable conditions, whereas some become red, which may indicate enough subaerial exposure for oxidation and dehydration of iron minerals.

Transgressive limestones.—Transgressive limestones are the thin (0.3–1.5 m) (1–5 ft) dense dark skeletal calcilutites denoted as the "middle" limestone of Moore's megacyclothem. They carry a diverse and relatively abundant marine biota comprising all the major phyla, although fossils do not seem abundant on outcrop because of the density of the rock. Fine grain size and diverse biota indicate deposition in the open marine environment far enough offshore to be below effective wave base, though above the effective base of the photic zone. If shoal-water facies such as oolite and stromatolite are present with the skeletal calcilutite, they are present only at its base. Aside from a few transgressive limestones that thicken as content of phylloid algae increases in southeastern Kansas, most limestones undergo little lateral facies change either northward or southward. Transgressive limestones generally record widespread marine inundation of the midcontinent and carbonate sediment production mainly at depths below effective wave base, where minor variations in topography on the underlying delta lobes would have caused little lateral variation in texture and composition of the limestone.

Offshore shales.—Offshore shales are thin (0.3–2.0 m) (1–7 ft), laterally persistent, only slightly sandy, gray, and of marine origin. They are included within limestone formations and were termed "core shales" by Heckel and Baesemann (1975) because of their central position within the megacyclothem. They typically contain a black fissile shale facies, which is rich in organic matter and generally contains nodules and laminae of nonskeletal phosphorite and relatively high concentrations of certain heavy metals. The black facies has no definitely benthic fossils, containing instead mainly conodonts in great abundance, fish remains, orbiculoid brachiopods, and other fossils reasonably inferred to have been pelagic or epipelagic. The gray facies typically includes only a sparse assemblage of several benthic invertebrate groups, including echinoderms, in addition to those found in the black facies; only away from the black facies do any of the offshore shales include abundant and diverse fossils. Like transgressive limestones, offshore shales also change laterally very little along the entire length of the outcrop. Those offshore shales in which the black facies disappears laterally have a sparsely fossiliferous, gray facies, which contains phosphate nodules locally.

Thinness in conjunction with great lateral persistence, fineness of detrital grain size, presence of marine fossils, great abundance of conodonts, and conspicuous nonskeletal phosphorite, all point to very slow sedimentation away from detrital influx, far offshore in deeper water. Preservation of abundant organic matter and an absence of benthic fossils in these shales indicate anoxic bottom conditions during their deposition. Lack of bottom oxygen in conjunction with very high concentrations of phosphate and heavy metals is explained by Heckel (1977) as the result of a two-layered quasi-estuarine circulation cell in the Midcontinent Sea, established when water became deep enough to form a thermocline above the bottom and to prevent vertical circulation and replenishment of bottom oxygen. Surface water, which was driven out of this sea by prevailing Pennsylvanian trade winds, was replaced by upwelling currents of phosphate-rich water from the deeper oxygen-minimum zone. This phosphate-rich water had been drawn in below the thermocline from intermediate depths of the open ocean. Nutrients brought into the photic zone by this upwelling promoted immense blooms of plankton, which settled back into the incoming lower water layer. There, massive organic decay further depleted the bottom water of remaining oxygen, continually enriched it in phosphate (as well as in heavy metals that are concentrated by organisms), and ultimately caused deposition of unoxidized organic matter and phosphorite on the sea bottom along with only very small amounts of other sediment to produce the phosphatic black mud facies.

Similar circulation, which concentrated phosphate, but which did not deplete all bottom oxygen, accounts for the sparsely fossiliferous gray facies containing phosphorite nodules. Deposition of terrigenous mud far from shore, probably below the effective limit of algal production of carbonate mud, but without establishment of the oxygen-depleting, quasi-estuarine circulation cell, accounts for the less common, more abundantly and diversely fossiliferous facies of some offshore shale.

Regressive limestones.—Regressive limestones constitute the "upper," sometimes with the "super," limestone members of Moore's megacyclothem. They are generally thicker (as much as 9 m (30 ft)) than transgressive limestones and contain a greater variety of facies.

The lower part of this limestone sequence consists largely of wavy-bedded skeletal calcilutite with many shale partings and containing an abundant and diverse marine biota consisting of elements of all major phyla. Lateral persistence of this facies, in conjunction with fine-grained lithology and diverse biota, indicates that this part of the regressive limestones, like most of the transgressive limestones, was deposited offshore below effective wave base but above the lower limit of algal carbonate production. In places, a type of skeletal calcarenite is present at the base of the regressive limestone. These calcarenites consist entirely of invertebrate grains that show no evidence of grain-abrasion, cross-bedding, or definite algal activity; thus, they probably formed in quiet water below effective wave base and probably below the limit of algal photosynthesis.

The upper parts of regressive limestone members have the most conspicuous lateral variation of facies along the midcontinent outcrop belt (fig. 7), as would be expected in deposits formed in shallow water where minor differences in bottom topography produce conspicuous lateral changes in facies. In northeastern Kansas, most regressive limestone grades upward to skeletal calcarenite that contains various proportions of abraded algal and invertebrate grains, osagia-coated grains, and grains with "micrite envelopes," which resulted from boring by algae; crossbedding is apparent in places. Although this vertical succession records increasing agitation of the water with time as water shallowed above effective wave base, this facies still records a relatively open marine environment. This part of the outcrop belt is thus defined as the "open marine facies belt," where good water circulation and normal marine salinity persisted longest during deposition of the regressive limestone. In places, the tops of some regressive limestones include deposits formed in very shallow water, deposits such as oolite and sparsely fossiliferous laminated calcilutite, which probably record deposition in local lagoons. Paleocaliche has been identified at the top of one regressive limestone.

Northward from Kansas, nearly all regressive limestone grades upward into unfossiliferous, laminated dolomitic calcilutite containing mudcracks and birdseye structures, which indicate shoreline and tidal-flat environments of deposition. This facies defines the "northern shoreward facies belt," which is thickest in Iowa and Nebraska.

In southeastern Kansas, most regressive limestones thicken as they grade upward into phylloid-algal mound facies, which defines the facies belt of that name. Mounds consist primarily of algae-dominated skeletal calcilutite, in which large blades of phylloid red and green algae characteristically shelter spar-filled voids. Mound-associated facies, particularly crossbedded, abraded-grain skeletal calcarenite and oolite, overlie and flank some of the mound facies, reflecting very shallow water over the buildups during later stages of regression.

Southward, in southernmost Kansas and northern Oklahoma, most regressive limestones grade into shale and sandstone, which define the "terrigenous detrital facies belt" and which represent a wide range of offshore to deltaic environments dominated by terrigenous clastic deposits from the Oklahoma detrital source.

Possible controls.—Typical Kansas cyclothems were initiated by influx of abundant terrigenous detrital sediment from encroaching deltaic shorelines, which formed the nearshore shale during a time of relatively shallow water in the Midcontinent Sea. This shallow-water condition could have resulted from either a eustatic fall in sea level or from basin-filling by rapid accumulation of carbonate sediment that formed the regressive limestone, and the subsequent influx of terrigenous detritus that produced the overlying nearshore shale.

Transgression, which resulted from either a eustatic rise in sea level or from an increase in subsidence of the basin, caused retreat of detritus farther away from the deepening basin in Kansas. This transgression allowed a thin layer of relatively pure carbonate to accumulate fairly uniformly over most parts of the inundated delta lobes, probably below effective wave base in deeper water. This process formed the calcilutitic transgressive limestone.

When water became deep enough to inhibit activity of benthic algae and to form a thermocline (perhaps as little as 100 m near the mouth of the sea; see Heckel, 1977), a quasi-estuarine water-circulation cell was set up. This circulation pattern drew phosphate-rich, oxygen-poor water in from intermediate depths of the open ocean, and, through upwelling and the concomitant nutrient-concentrating effect of decay of abundant trapped organic matter, oxygen in bottom water was depleted to various degrees to form both the gray and black phosphatic facies of the offshore shale.

Shallowing of the sea then destroyed the thermocline and broke up the quasi-estuarine cell to allow significant reoxygenation of the bottom and reestablishment of benthic invertebrate and algal carbonate production to initiate formation of the regressive limestone. Eustatic fall of sea level seems most reasonable as a cause of this shallowing because too little sediment to cause significant basin-filling is evident in the offshore shale facies, and tectonic reversal of bottom subsidence, particularly in a cyclic fashion, seems less likely. Continued shallowing of the sea allowed the formation of upper regressive limestone facies in shoal water, lagoons, and tidal flats. Fine terrigenous material from distal ends of progressively encroaching deltas accounts for the abundance of shaly partings in most regressive limestone. The thickness of sediment in the regressive limestone is sufficient to account for a slight relative lowering of sea level by partial filling of the basin with carbonate sediment; it seems insufficient, however, to account, in itself, for the small vertical distance (as little as 6 m, or 12 m, allowing for 50 percent compaction) between the top of the phosphatic black-shale facies and the supratidal facies at the top of the regressive limestone, in light of our present understanding that deposition of this type of phosphorite requires water of substantially greater depth.

Formation of the regressive carbonate facies ceased when the sediment-producing organisms were overwhelmed, in various stages of facies development (depending on topographic position, which was related generally to facies belt), by influx of terrigenous detritus from prograding delta lobes that initiated the succeeding nearshore shale. The cycle then repeated when another relative rise in sea level took place.

Although repeated eustatic rise and fall of sea level from some external control seems to be the simplest explanation for most aspects (and the best for some aspects) of the overall sequence of Kansas cyclothems, this explanation by no means negates the probability that other cyclic mechanisms played a role. In fact, the model of delta progradation, abandonment, and subsequent progradation elsewhere, which has been shown to be applicable to other cyclic sequences by several authors (see Ferm, 1970, for the Appalachian Pennsylvanian), not only explains the complex changes in thickness and facies development in nearshore shales, but this model also probably accounts for the occurrence in some areas of less common "lower" and "fifth" limestone members in Moore's megacyclothem as resulting from production of abundant carbonate sediment during a relatively long-term shift of active delta building away from the Kansas outcrop during the general phase of greatest regression.

Cherokee cyclothems.—The interplay of both depositional models needs only slight modification to account for the simpler cyclothems that Moore (1949, 1950) detected in the Cherokee Group (fig. 5). In the ascending repeated sequence of lithologies—sandstone, sandy shale, underclay, coal, black shale, gray shale, limestone, calcareous shale, sideritic shale—the sandstone through coal part is nonmarine, whereas the black shale through at least the lower calcareous shale part is marine. Those sequences in which phosphatic black shale is lacking may reflect merely local delta abandonment during regression without initiation of a new transgressive cycle. Sequences that contain phosphatic black shale, however, are examples of Kansas cyclothems that are more dominated by nearshore terrigenous sediment than are the limestone-rich cyclothems characteristic of the Marmaton-Missourian-Shawnee section.

Nonmarine delta-plain deposits are very conspicuous in the initial, "nearshore shale" part of Cherokee cyclothems. Transgressive limestone is rarely present above the coals, because few algae or calcareous invertebrates could colonize the generally inimical low-oxygen environment of the deepening sea bottom over partially decayed vegetation of the coal swamp. During transgression, the little detrital sediment that was carried from the increasingly distant shoreline to the drowned swamps in Kansas was incorporated with the shells of the few more tolerant organisms to form the base of the black shale overlying the coal. The remainder of the black shale was deposited during maximum transgression, in depths great enough for establishment of the quasi-estuarine circulation cell that led to formation of non-skeletal phosphate nodules.

Then shallowing of the sea brought about deposition of the gray shale as the bottom became reoxygenated. Further shallowing allowed limestone (equivalent to the regressive limestone of later cyclothems) to form, as algae and more invertebrates became established. The final units (calcareous shale and sideritic shale) are the initial prodeltaic and delta-front deposits of the succeeding nearshore shale, which prograded seaward rapidly enough to prevent the underlying regressive limestone (where present at all) from becoming very thick or developing shoal-water facies.

Wabaunsee cyclothems.—The cyclothems described by Moore (1936, 1949, 1950) in the Wabaunsee Group are basically alternations of nearshore sandy shale containing nonmarine sandstone and coal, with marine limestone and thin marine shale beds. The lowest limestone formation (Howard) (fig. 5) contains a black shale (and gray shale) between a thin dense limestone, below, and a thicker limestone, above, which is similar to the typical Kansas cyclothems in older groups (Moore, 1936, p. 206).

Unlike the relationship between cyclothems and lithic subdivisions in lower groups, Wabaunsee nearshore shales above the Howard comprise both the shale formations and the shale members of the more recently named limestone formations, whereas the limestone part of the cyclothem is composed of a single limestone member, whether or not it is grouped with another limestone member in a limestone formation (compare Moore, 1949, p. 180–181, with Zeller, 1968, pl. 1). Black phosphatic shale has not been reported from any Wabaunsee unit above the Howard, and the possible presence of nonblack offshore shale is not established at this time. Thus, which of the higher Wabaunsee cyclothems resulted from major transgressive-regressive events and which are merely the result of local delta abandonment during general regression is not known. One would suspect, however, that laterally persistent limestone that contains a persistent medial shale bed might represent the former, whereas limestone of limited extent might represent the latter.

BIOSTRATIGRAPHY

MISSISSIPPIAN

No comprehensive paleontologic study of the outcropping Mississippian limestones in southeastern Kansas has been accomplished. Girty (cited in Smith and Siebenthal, 1907) recognized rocks in the Missouri-Kansas border area that are equivalent in age to the Keokuk Formation (Upper Osagean) of the upper Mississippi River valley on the basis of fossil brachiopods, bryozoans, and corals. Thompson and Goebel (1968) have recovered conodonts at two localities from limestone in surface exposures of the Mississippian sequence in Cherokee County, which indicate that these rocks are equivalent in age to the Keokuk Formation. Nodine-Zeller and Thompson (1977) have discussed endothyrid foraminifers and conodonts in a core from a depth of 15–23 m (50–76 ft) near outcrops in Cherokee County, which suggest that the uppermost Mississippian in the core is Chesterian in age and that it overlies rocks of Meramecian age.

Girty (1940) recognized rocks in the subsurface of Kansas that are equivalent in age to those of all four stages of the Mississippian in the Mississippi River valley area; his studies were based on fossil invertebrates, mainly bryozoans and brachiopods, in cores from wells drilled in Kansas.

Thompson and Goebel (1963, 1968) and Goebel (1966) studied the conodont biostratigraphy of Mississippian rocks in well cores from western Kansas.

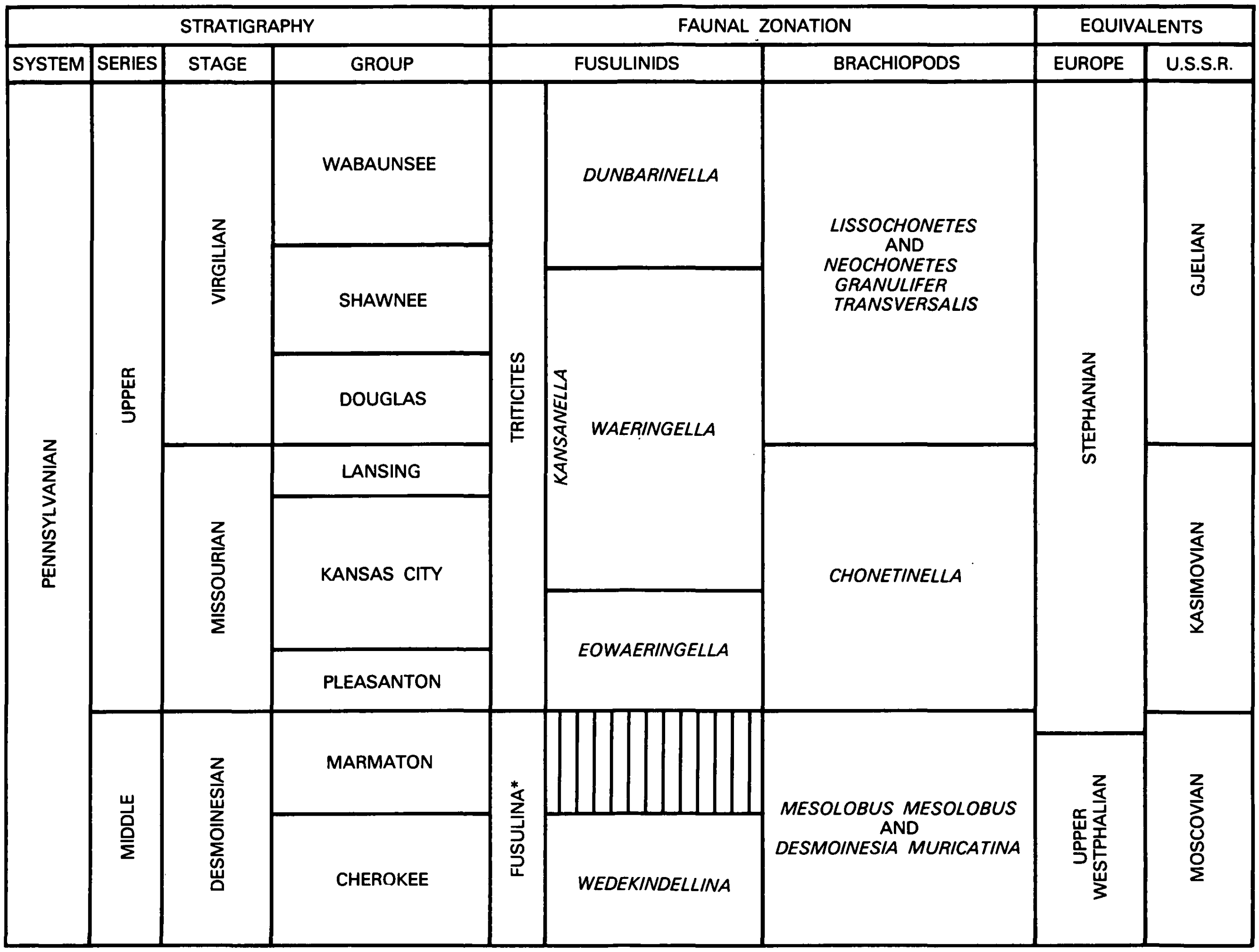

FIGURE 8.—Chronostratigraphic chart of outcropping Pennsylvanian rocks in Kansas.

On the basis of assemblages of conodonts that are only partly comparable with those of similar age in the type areas of the Mississippi River valley, they revised somewhat the earlier work of Lee and Girty (1940) and extended the knowledge of lithostratigraphy of these subsurface beds. Their results are complementary to unpublished work by Selk and Ciriacks (1968). Zeller (*in* Ebanks, Euwer, and Nodine-Zeller, 1977) has identified rocks of Meramecian age, on the basis of endothyrid foraminifers, from well cores in Hodgeman County, western Kansas. There is much room for future work in rectifying the lithostratigraphic correlations of petroleum geologists by using the scanty biostratigraphic determinations now available in western Kansas; also, some disagreement exists between the ages assigned to certain Mississippian beds on the basis of assemblages of conodonts and those assigned on the basis of calcareous microfossils (D. E. Nodine-Zeller, unpub. data, 1978).

PENNSYLVANIAN

Beds of Morrowan age have been recognized in the subsurface of southwestern Kansas by Thompson (1944) through identification of *Millerella* and forms that would now be assigned to *Eostaffella* in a well core. Dark limestone and shale above this Morrowan shale and sandstone sequence have received very little paleontologic study, but McManus (1959, p. 49) mentioned the presence of *Fusulinella* sp. in these rocks. If *Fusulinella* is the only fusulinid found in these rocks they could be considered as tentatively Atokan in age. However, the genus *Fusulinella* ranges from upper Atokan to upper Desmoinesian (Douglas, 1977, p. 476, fig. 5). Basal Pennsylvanian beds in the subsurface of northeastern Kansas have also been assigned to the Atokan, but there are no paleontologic data to confirm this.

The oldest Pennsylvanian rocks that crop out in Kansas are thin beds of dark shale of possible Atokan age, which occur only in isolated areas on the eroded surface of Mississippian limestone in southeastern Cherokee County (Stewart, 1975). Paleontologic data for this age assignment do not exist; in fact, these beds, where present, are not formally distinguished from the overlying Cherokee Group (Desmoinesian) because of similarity of lithology of the two and the lack of paleontologic or palynologic studies to define the extent of their contact. Stratigraphy and invertebrate paleontology of the Desmoinesian formations in southeastern Kansas are described in reports by Jewett (1945) for the Marmaton Group, and by Williams (1938) and Howe (1956) for the Cherokee Group.

Subdivisions of the outcropping Pennsylvanian strata in Kansas are based on the occurrence of widely recognized fusulinid genera and on evidence of physical breaks in the sedimentologic record (Moore, 1936, 1949; Moore and others, 1944) (fig. 8). The contact between Middle Pennsylvanian (Desmoinesian) and Upper Pennsylvanian (Missourian) rocks is placed at the base of the Hepler Sandstone Member of the Seminole Formation, basal unit of the Pleasanton Group, because this boundary has been thought to separate the zones of *Triticites,* above, and *Fusulina,* below; this contact may correspond to a regional disconformity (Moore and others, 1944). One species, *Fusulina fallsensis,* however, has been found only in the lower Missourian Bethany Falls Limestone Member of the Swope Limestone (fig. 5) (Thompson and others, 1956; Thompson, 1957) associated with *Eowaeringella ultimata* (fig. 8). This occurrence is below the first appearance of *Triticites,* in the Winterset Limestone Member of the Dennis Limestone (fig. 5), so the generic ranges of *Fusulina* and *Triticites* do not overlap here. This relationship raises a question regarding both the temporal magnitude of the physical break at the base of the Pleasanton and the placement of the Desmoinesian-Missourian boundary. Missourian and Virgilian beds are distinguished on the basis of paleontology (see fig. 8) and differences in the lithologic character of cyclic rock sequences (Jewett and others, 1968).

One of the most interesting areas of biostratigraphic work in Kansas has been in recognition of the Pennsylvanian-Permian boundary. History of the changing placement of this boundary was reviewed by Moore (1949). Mudge and Yochelson (1962) studied the paleontology of beds above and below this important datum, which presently is placed at the top of the Brownville Limestone Member of the Wood Siding Formation, and concluded that this assignment is arbitrary but that it is justified on the basis of practicality, long-time usage, and some paleontologic evidence.

The Pennsylvanian-Permian contact continues to be a subject of controversy, mainly because of differences in opinion between paleontologists and palynologists. Studies of palynomorphs have led some workers to suggest that all of the Gearyan stage (Lower Permian) in Kansas should be included in the Pennsylvanian, the systemic boundary being placed much higher in the section than at present (Clendening, 1971, 1975; Wilson and Rashid, 1971). Several papers referring to the paleontologic and palynologic aspects of this boundary in Kansas are included in a recent symposium (Barlow, 1975), which is important here, because the Pennsylvanian-Permian boundary in Kansas has long been the popular one in America. Unfortunately, many biostratigraphers persist in arguing about the relative merits of diverse fossil groups, which may be stratigraphically or ecologically incompatible and therefore do not justify arbitrary juggling of chronostratigraphic boundaries.

The status of biostratigraphically useful fossils (land plants, fusulinids, sponges, corals, bryozoans, brachiopods, bivalves, gastropods, cephalopods, trilobites, ostracodes, and crinoids) that are applicable to the Pennsylvanian in Kansas was given by Moore and others (1944, p. 668–678). Additional notations and references to the paleontology of the outcropping Pennsylvanian rocks in Kansas may be found in Moore and others (1951), Thompson (1957), and Jewett and others (1968). The taxonomic groups listed above are abundant and diverse in Pennsylvanian rocks of Kansas, but there have been few detailed biostratigraphic studies of occurrences of these fossils in Kansas since the 1944 summary by Moore.

Studies by Cridland and others (1963) on land plants, by Jeffords (1947) and Cocke (1970) on corals, and by Strimple (1951), Strimple and Moore (1971), and Moore and Jeffords (1967) on crinoids suggest that these groups could be of some biostratigraphic value, but much remains to be done. It would be especially valuable to have other summaries, such as that given by Moore and Strimple (1973, table 1) for Early Pennsylvanian crinoids. Cephalopods, a particularly useful biostratigraphic fossil for the Carboniferous, should be studied more thoroughly in Kansas, as only one of the eight genera of ammonites (*Prouddenites*), listed as indicative of Pennsylvanian faunal zones by Moore and others

(1944), was reported by Miller and others (1957) to occur in Kansas.

The importance of studies of palynomorphs in recognition of the Pennsylvanian-Permian boundary has already been mentioned, and they will be valuable in future biostratigraphic subdivisions of these systems. Numerous studies of conodonts in Pennsylvanian rocks of adjacent States have biostratigraphic application to the Kansas section (Lane, 1967; Henry, 1970; Lane and Straka, 1974). Within Kansas, conodonts have been used more commonly to interpret environments of deposition represented by the cyclic sequences of rocks in the Upper Pennsylvanian (Von Bitter, 1972; Baesemann, 1973; Heckel and Baesemann, 1975; Perlmutter, 1975; and Wood, 1977). From these studies, we should be able ultimately to develop a better understanding of the biostratigraphic significance of these enigmatic fossils.

Concerning fusulinids, Wilde (1975, p. 123) suggested that the basic pattern of stratigraphic zonation based on fusulinids has been established for decades but that refinement of zones continues. This is particularly true of the Virgilian of Kansas, where very little careful biostratigraphic work has been published.

Calcareous phylloid algae are important constituents of middle and upper Pennsylvanian limestone in Kansas (Johnson, 1946; Harbaugh, 1960; Wray, 1964). Although these fossils have not been particularly useful in studies of biostratigraphy, other fossil algae, such as *Komia,* may be useful for recognizing certain time-stratigraphic subdivisions of the Pennsylvanian (D. E. Nodine-Zeller, unpub. data, 1979).

Much research utilizing fossils of the Kansas Pennsylvanian in recent years has been to aid in the interpretation of facies and environments of deposition (Toomey, 1969; Heckel, 1975; Senich, 1975, 1978). Other studies have dealt with ecological aspects of fossil populations (Koepnick and Kaesler, 1971; Brondos and Kaesler, 1976; Songsirikul, 1977) and with "community" ecology (Scott, 1973; Pearce, 1973). Although these studies are needed and should be encouraged, the need is also great for monographic studies of different phyla, classes, orders, etc., and for studies of the complete preserved fossil assemblage by a team of specialists (that is, faunal studies). Examples of faunal studies are those by Williams (1938) on the Desmoinesian invertebrates of southeastern Kansas and by Mudge and Yochelson (1962) on Upper Pennsylvanian and Lower Permian rocks of Kansas. Investigations like those by Cooper and Grant (1972–1977) on the Permian brachiopods of west Texas and by Sutherland and Harlow (1973) on the Pennsylvanian brachiopods of New Mexico are examples of the type of monographic studies needed.

Future studies should use large collections to evaluate existing taxa (Koch, 1977) and to create new ones. This use will provide a firmer biological basis for taxa, and biology is needed in biostratigraphy as well as in the other areas of paleobiologic research. Three studies (two on brachiopods and one on sponges) illustrate the advantages of studying large collections. An ecological study of some "in situ" chonetellids from the Tacket Formation (Missourian), using hundreds of specimens from a single stratigraphic and geographic locality in southeastern Kansas, suggests that *Chonetinella flemingi* and *Chonetinella alata* are end members of a single species (R. R. West, unpub. data, 1978). The second study (Gundrum, 1977) showed that in a large collection (hundreds of specimens) from a single stratigraphic and geographic locality, specimens of *Mesolobus mesolobus* cannot be separated from those of *Eolissochonetes bilobatus.* Lastly, a careful examination of silicified specimens of some Missourian heliosponges suggests that certain established taxa possess many overlapping characteristics and that revision is needed (Gundrum, in press). If this is true for brachiopods and sponges, it is very probably true also for other biologic groups. The ease of collecting abundant fossils from most localities in the Kansas Pennsylvanian should facilitate studies of large collections.

Further advances in studies, in Kansas, of the biostratigraphy of Pennsylvanian rocks, which comprise interbedded marine, transitional, and nonmarine rocks, awaits integration of knowledge of different groups of fossils and better definition of species and higher taxa on the basis of sound biological concepts. In the future, the composite standard technique (Shaw, 1964) may well provide the most practical basis for biostratigraphic subdivision of the Kansas Pennsylvanian.

FOSSIL COLLECTING

Excellent fossil collecting is available from Pennsylvanian rocks in Kansas. Some references given here (for example, Williams, 1938; Mudge and Yochelson, 1962; Ball, 1964) have good descriptions of localities, most of which are still accessible. Merriam (1963) has provided road logs for outcrops in different parts of the State, especially along major highways. Guidebooks of the Kansas Geological Society that deal with outcrops in eastern Kansas, especially in the Kansas River valley and from Kansas City southward to Oklahoma, are very useful. Topographic and geologic maps and other information are available at the Kansas Geological Survey in Lawrence.

ECONOMIC PRODUCTS

COAL

Rocks of the Middle and Upper Pennsylvanian Series in Kansas include the economically important coal beds that were mined in the past and the economically important reserves of coal yet to be mined. At least 42 coal beds are present in the Pennsylvanian strata of Kansas, and 17 of these beds have economic coal reserves.

One coal bed, the Weir-Pittsburg coal, has accounted for approximately 180 million metric tons (200 million short tons) of total Kansas production. Production from this coal bed was in Cherokee and Crawford Counties, primarily by room-and-pillar mining (Abernathy, 1944; Young, 1925, p. 60–96). Besides the Weir-Pittsburg coal, other important Cherokee Group coals that have been extensively mined include the Mineral, Fleming, Croweburg, Bevier, and Mulky coals. These coals, described by Pierce and Courtier (1938) and Howe (1956), have accounted for an additional 64 million metric tons (70 million short tons) of production, mainly by surface mining. A petrographic study by Hambleton (1953) related the characteristics of these Cherokee coals to their potential for utilization.

Two other important commercial coals in Kansas are the Mulberry coal of the Marmaton Group (Schoewe, 1955) and the Nodaway coal of the Wabaunsee Group (Schoewe, 1946). Approximately 8 million metric tons (9 million short tons) of Mulberry coal was mined, mainly in Linn County, and nearly 11 million metric tons (12 million short tons) of Nodaway coal. Most of the Nodaway coal was mined in Osage County by old longwall methods (Young, 1925, p. 118–119), whereas the Mulberry coal was recovered chiefly by area strip mining.

Two characteristics of Kansas coal that control its economic development are the thinness of the coal beds and the high sulfur content of the coal. All coal reserves in Kansas are medium-to-high-sulfur coal (more than 1 percent sulfur) (Allen, 1925); 3 to 5 percent sulfur content is common in most commercial coals. Most of the coal beds are thin, that is, less than 71 cm (28 in.); in the future, most of the coal mined in Kansas probably will be recovered by surface-mining methods. Several areas of strippable coal and most of the Weir-Pittsburgh coal beds are of intermediate thickness, that is, 71–107 cm (28–42 in.).

The demonstrated reserve base of bituminous coal for strip mining totals 905 million metric tons (998 million short tons) under less than 30 m (100 ft) of overburden (Brady and others, 1976). For coal beds having an overburden-to-coal thickness ratio of 30:1 or less, the demonstrated reserve base is 477 million metric tons (526 million short tons). In addition, 1,647 million metric tons (1,816 million short tons) of inferred coal reserves is under 30 m (100 ft) of overburden; of this total, 720 million metric tons (794 million short tons) has a stripping ratio of 30:1 or less. General distribution of the areas in Kansas having coal reserves is shown in figure 9.

During the 1970's, coal-mining activity has been almost entirely in southeastern Kansas in Linn, Bourbon, Crawford, and Cherokee Counties, where coal of the Cherokee Group is surface mined. Total recorded coal production for Kansas is approximately 260 million metric tons (287 million short tons) through 1976 (fig. 10). Of this total, nearly 68 percent was won by subsurface mining methods; since 1963, however, all coal mining in Kansas has been by surface methods, mainly area strip mining.

PETROLEUM

Mississippian and Pennsylvanian rocks in the subsurface of Kansas have been extremely important as sources and reservoirs of oil and natural gas. The first well drilled for oil and gas west of the Mississippi River was completed in 1860, producing oil from Middle Pennsylvanian sandstone at about 100 m (300 ft) near Paola in Miami County (Jewett and Abernathy, 1945). Among the latest and best oil and gas discoveries in the State are those in Mississippian and Pennsylvanian limestone in western Kansas.

Middle Pennsylvanian sandstone of the Cherokee Group was the focal point of early exploration in southeastern Kansas (fig. 11). The famous "shoestring sands" (Rich, 1923; Bass, 1937) of that area were economically very important in the early flush production of oil and gas and the accompanying industrial development. As primary production declined, these same fields were the sites of early waterflooding activity, which resulted in many techniques presently in use elsewhere. In recent times, the third, or tertiary, phase of operations has begun in these same fields. Many of them still contain millions of barrels of oil, which is producible only by

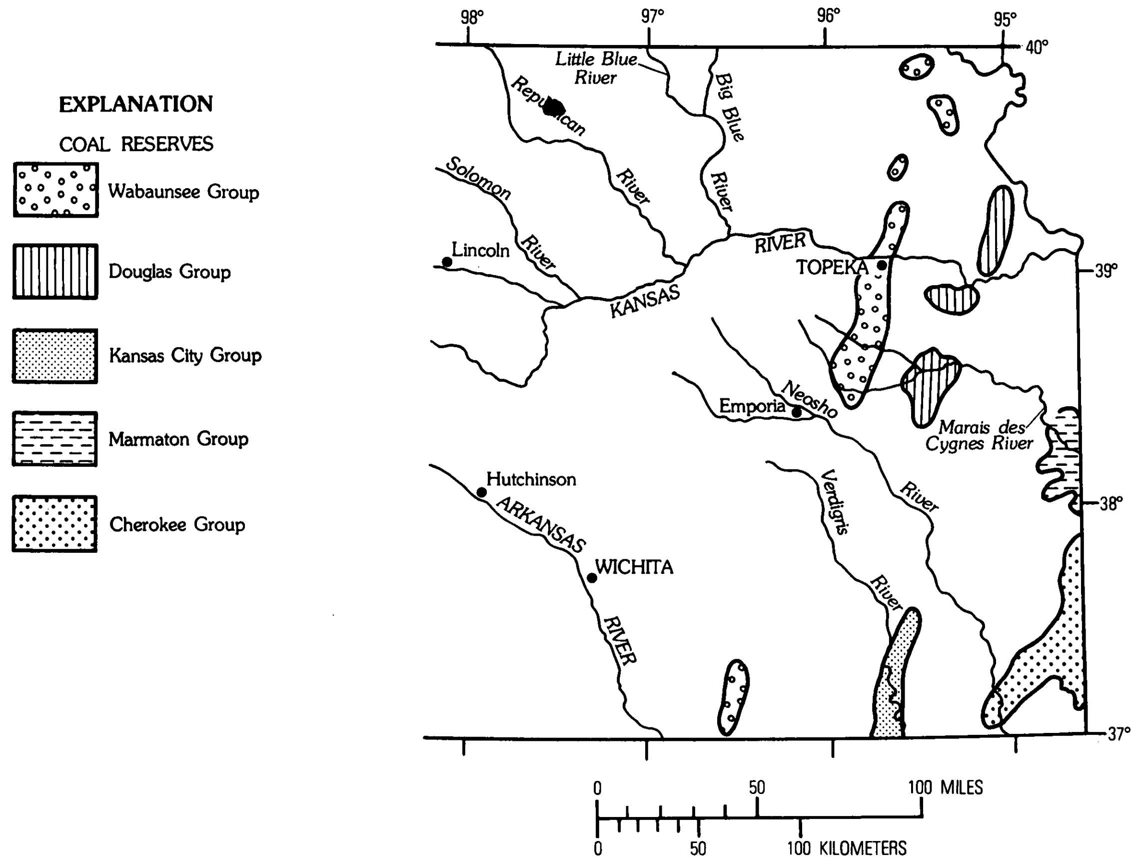

FIGURE 9.—Areas of strippable coal reserves beneath overburden of 30 m (100 ft) thickness or less. Formations in which coal reserves occur in the various areas are shown by patterns (Brady, and others, 1976).

methods such as thermal stimulation or the injection of surfactant chemicals. Cherokee sandstone also contains near-surface deposits of heavy oil and surface occurrences of tar sand in the Kansas-Missouri border area, which may be important as future energy resources (Ebanks and others, 1977).

Large amounts of oil and gas have also been produced from Upper Pennsylvanian rocks of central and western Kansas on or near major uplifts in the subsurface (fig. 11). These units of the Missourian and Virgilian Stages comprise limestone and sandstone reservoirs that usually have discontinuous porosity. The result is that many fields produce from combination structural-stratigraphic traps, at depths of 900–1,400 m (3,000–4,600 ft), both as fields having but one producing formation and as those in which the Pennsylvanian traps occur above deeper zones of production in Mississippian or Ordovician formations (Moore and Jewett, 1942; Merriam and Goebel, 1956). Almost one trillion cubic feet of gas has been produced from Upper Pennsylvanian lime-

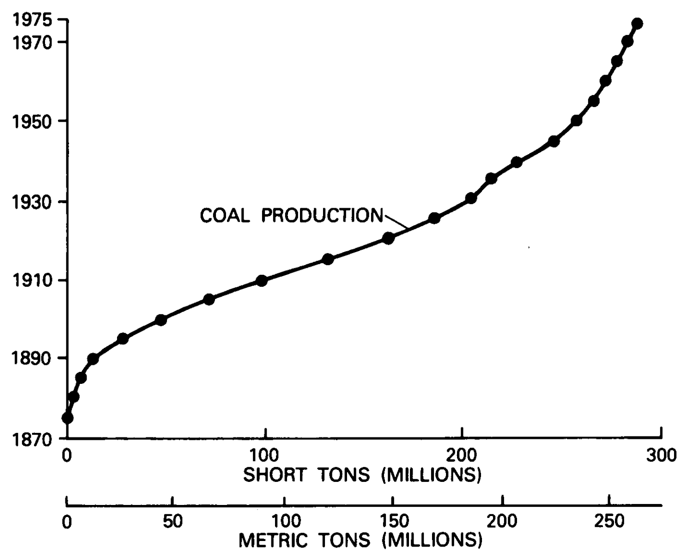

FIGURE 10.—Cumulative production of coal in Kansas. Almost all the coal produced has come from rocks of Pennsylvanian age.

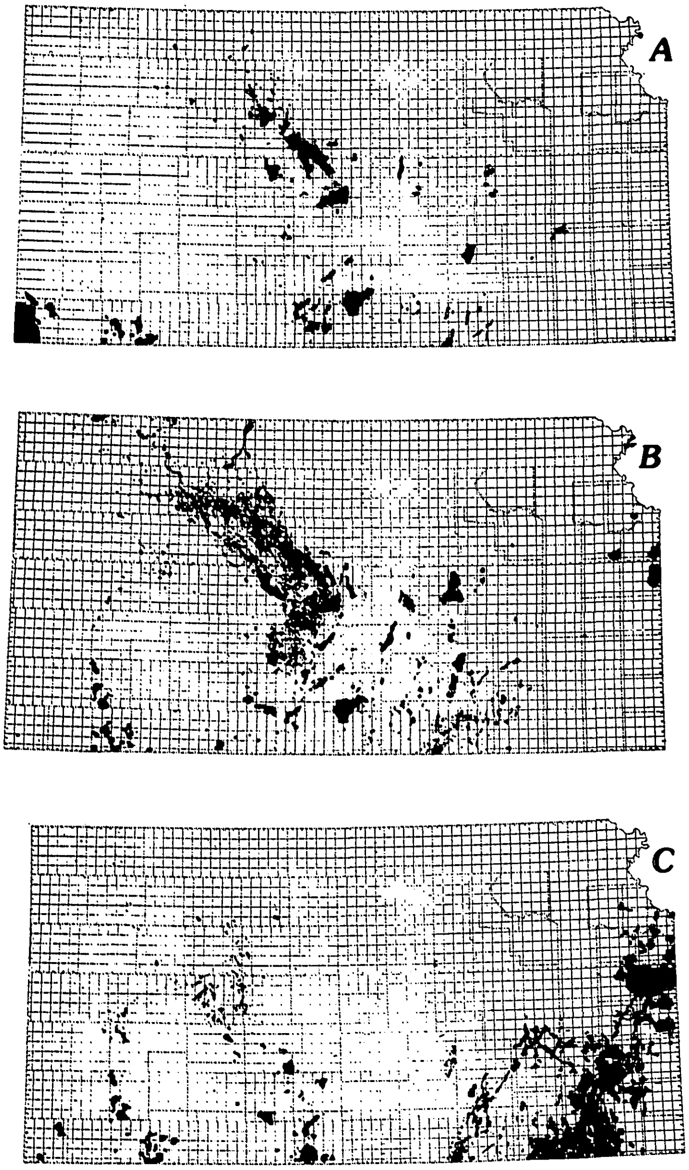

FIGURE 11.—Areas of production of oil from subsurface Upper and Middle Pennsylvanian rocks in Kansas: A. (upper) production from Wabaunsee, Shawnee, and Douglas Groups; B. (middle) production from Lansing and Kansas City Groups; C. (lower) production from Marmaton and Cherokee Groups (Ebanks, 1974, figs. 10, 11; Ebanks, 1975, fig. 12).

stone in one especially important field, the Greenwood Field, on the Colorado-Kansas border, partly underlying the famous Permian Hugoton gas field (Beene, 1977).

Lower Pennsylvanian sandstone and Mississippian carbonate oil and gas reservoirs are principally in southwestern and south-central counties (fig. 12), where they are important in local structural or stratigraphic traps in basinal areas at depths of 1,350–1,800 m (4,500–6,000 ft). The variability of

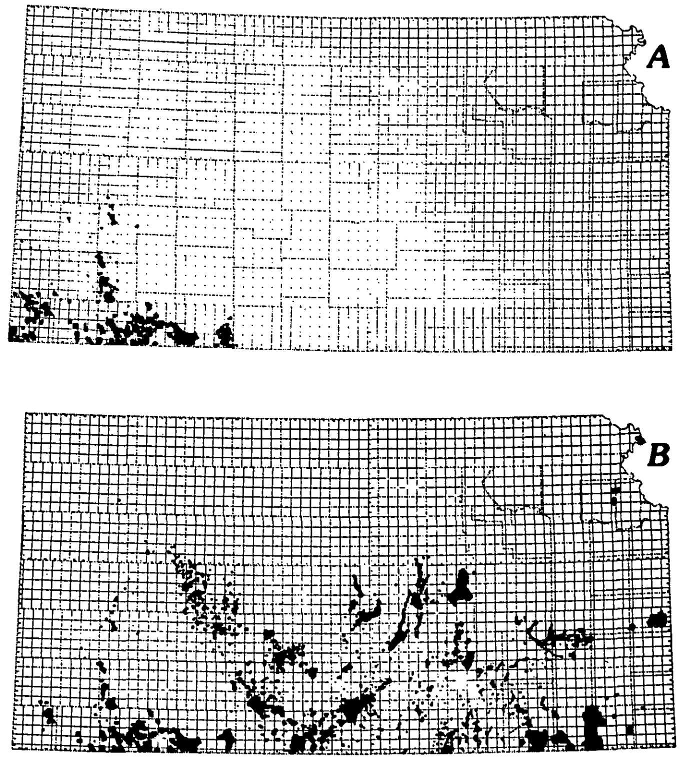

FIGURE 12.—Areas of production of oil from subsurface Lower Pennsylvanian and from Mississippian rocks in Kansas: A, (upper) production from "Atoka" and "Morrow" rocks; B, (lower) production from Mississippian (undifferentiated) rocks (Ebanks, 1974, fig. 13; Ebanks, 1975, fig. 8).

limestone and dolomite lithofacies in the Mississippian sequence and the complex diagenesis that has taken place in these rocks provide almost endless opportunities for imaginative exploration for oil and gas.

Oil and gas in Mississippian rocks tend to be in "trends" of favorably porous and permeable lithofacies within the overall interval of limestone and dolomite. In south-central Kansas, the most important types are those related to paleogeomorphic features and those in which fracturing of cherty rocks of the Osagian Stage has enhanced the permeability of the reservoir. Farther northwest, production is found in subunconformity traps in which lower Meramecian dolomite and dolomitic limestone are present beneath Middle Pennsylvanian shale. To the southwest, in slightly younger Meramecian beds, the oil and gas are trapped in oolitic and bioclastic limestone. Farther south, near the Oklahoma border, sandy limestone of Chesterian age forms oil and gas traps over local structurally high areas or in porosity-pinchout traps (Kansas Geological Society, 1956, 1959, 1965).

Cumulative oil production in Kansas at the end of 1976 was about 4.7 billion barrels, and cumulative gas production amounted to more than 23 trillion cubic feet. Of these total amounts, approximately 40 percent of the oil and 10 percent of the gas are estimated to have been produced from Pennsylvanian and Mississippian formations (Beene, 1977). The brightest prospects for future discoveries are also in these producing zones.

METALLIC ORES

Large amounts of zinc and lead were mined from the Mississippian rocks in southeastern Kansas. An early summary of the geology, mineralogy, and mining techniques used here was given by Haworth and others (1904). Ore bodies in Kansas, along with the zinc and lead deposits of southwestern Missouri and northeastern Oklahoma, make up the large Tri-State District (Brockie and others, 1968). This district was the major producer of zinc in the world for many years and has produced more than 2 billion dollars worth of lead and zinc. The most important field in this district was the Picher, in Oklahoma and Kansas (Lyden, 1950; McKnight and Fischer, 1970).

The Kansas part of the district has produced more than 2.6 million metric tons (29 million short tons) of zinc having an estimated value of 436 million dollars, and 590 thousand metric tons (650 thousand short tons) of lead worth nearly 91 million dollars (data from Martin, 1946, and U.S. Bureau of Mines Yearbooks, 1946–75). Cumulative production of these metals is shown in figure 13.

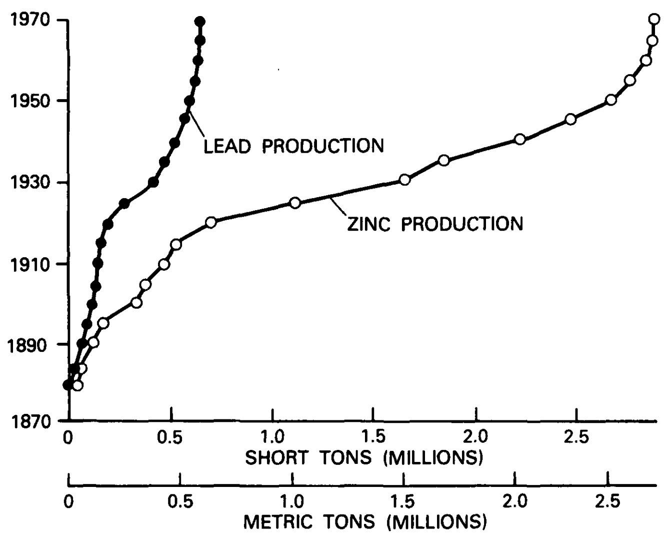

FIGURE 13.—Cumulative production of recoverable lead and zinc from mines in Kansas (data from Martin, 1946; U.S. Bureau of Mines minerals yearbook, 1946–70).

Ore bodies of the Tri-State District are restricted almost entirely to rocks of Mississippian age, specifically, to the Keokuk and Warsaw formations (fig. 4) (Moore, 1928; Brockie and others, 1968). Important studies of the relationship of the ores to the stratigraphy and structure include works by Fowler and Lyden (1932), Fowler (1938), and Moore and others (1939). Early geophysical investigations of the district were reported by Jakosky and others (1942), and later research into these indirect methods of exploration was reported by Hambleton and others (1959). In general, exploratory drilling has been necessary to the discovery of deeply buried ore bodies.

Within the Tri-State area, sphalerite and galena are the commercial ore minerals; however, many other minerals are associated with the ores, including chalcopyrite, wurtzite, and enargite (McKnight and Fischer, 1970, p. 101–124). Forms of the ore bodies have been described by Brockie and others (1968, p. 421–425) as assuming three basic shapes: (1) irregular, relatively narrow, long ore "runs" of varying heights; (2) circular "runs"; and (3) flat-lying, generally tabular bodies called "sheet ground" that cover large areas. The most important ore bodies in the district were the elongated "long runs."

Kansas production of lead and zinc ended in 1970. Low-grade ore and new antipollution standards contributed to the cessation of operations in the Tri-State District. Recently, renewal of interest in exploring for mineral deposits in the subsurface west of the areas previously mined has resulted in extensive drilling. Results of this drilling are not presently known.

LIMESTONE AND OTHER NONMETALLIC MINERALS

Limestone and shale of the Middle and Upper Pennsylvanian Series have been used extensively in Kansas. Limestone is used throughout eastern Kansas for different construction projects, especially as concrete aggregate and road metal, also as agricultural lime. In addition, limestone is used for cement manufacture at five different locations.

At least 20 different limestone units of Pennsylvanian age are presently used for crushed stone in Kansas. Total tonnage of crushed stone from Pennsylvanian rocks in 1976 is estimated at 12.1 million metric tons (13.4 million short tons), which has a value of approximately 29 million dollars. Important limestone units used for crushed stone are listed below:

Limestone Member	*Limestone Formation*
Ervine Creek	Deer Creek
Plattsmouth	Oread
Stoner	Stanton
Captain Creek	Stanton
Argentine	Wyandotte
Raytown	Iola
Bethany Falls	Swope
Laberdie	Pawnee

Three limestone members are presently mined by underground methods, the Bethany Falls, Argentine, and Plattsmouth. All present underground limestone mines are using, or plan to use, the mined space for commercial storage, and this use is considered in the pillar placement and design of their mine plan. Underground mined space is being used extensively for storage in the Kansas City and Atchison areas. Figure 14 shows locations of the large mines and quarries in Kansas.

Dimension stone was extensively quarried from Pennsylvanian rocks in Kansas in the past. Local limestone and sandstone were used to construct many of the buildings seen in the towns of eastern Kansas; the more important units of Middle and Late Pennsylvanian age are listed below (Risser, 1960; Grisafe, 1976):

Rock member	*Formation*	*Area of main use*
Utopia Limestone	Howard Limestone.	Eastern Kansas, sidewalks.
Hartford Limestone.	Topeka Limestone	Topeka area.
Big Springs Limestone.	Lecompton Limestone.	Lecompton area.
Kereford Limestone.	Oread Limestone	Atchison area.
Toronto Limestone	Oread Limestone	Atchison area.
Iatan Limestone	Stranger Formation.	Leavenworth area.
Raytown Limestone	Iola Limestone	Several localities.
Westerville Limestone.	Cherryvale Shale	Kansas City.
Bandera Quarry Sandstone.	Bandera Shale	Eastern Kansas many areas.

Of all the units quarried, the Bandera Quarry Sandstone was the most extensively worked; it was produced commercially in Crawford, Bourbon, Labette, and Neosho Counties. At present, Pennsylvanian rocks are used only as local rubble-stone; no commercial cut-stone quarries are now in existence. All the dimension stone presently produced in Kansas is from quarries in Lower Permian rocks.

Portland cement is manufactured at five different locations in Kansas, and all the plants use Pennsylvanian limestone. In addition, portland cement was formerly produced at 11 other locations, and natural cement was produced at one plant, all from Pennsylvanian limestone. The Raytown Limestone Member of the Iola Limestone is used for cement manufacture at two locations in Kansas. In addition, the Argentine Limestone Member of the Wyandotte Limestone, the Drum Limestone, and the Stanton Limestone are each utilized at other cement plants in Kansas. Figure 14 shows locations of cement plants in Kansas. Production from Pennsylvanian limestone and shale has accounted for an estimated cumulative total of 82.3 million metric tons (90.7 million short tons) of cement through 1976, having a total value of nearly 1,270 million dollars (Schoewe, 1958; U.S. Bureau of Mines Yearbooks 1957–1975; and Kansas Geological Survey estimates.)

Brick, sewer tile, pottery, and lightweight aggregate are all products made in Kansas from clay and shale of Pennsylvanian age. Extensive shale deposits are present in eastern Kansas, which led to widespread use of the shale for brick manufacture. Between 1868 and 1888, nearly every town in eastern

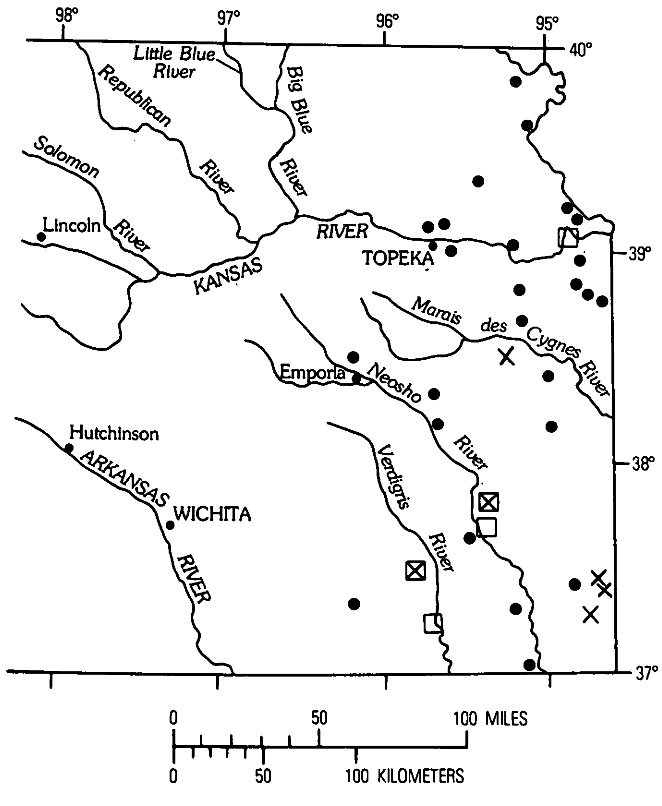

FIGURE 14.—Locations of large quarries and mines from which limestone is obtained, and locations of plants where cement and clay products are manufactured in Kansas. ● Limestone quarry or mine (100,000 metric tons, plus); □ cement plant; × clay products plant.

Kansas had a new plant starting the manufacture of brick (Douglas, 1910).

Six plants are now in operation in Kansas that utilize Pennsylvanian shale and clay for the manufacture of different clay products (fig. 14). As mentioned above, shale is also used in the manufacture of cement at the five portland cement plants. The geologic units used for the different clay products include the Krebs Formation and Cabaniss Formation of the Cherokee Group, the Lane Shale and the Bonner Springs Shale of the Kansas City Group, and the Weston Shale Member of the Stranger Formation in the Douglas Group.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Oklahoma

By ROBERT O. FAY, S. A. FRIEDMAN, KENNETH S. JOHNSON, JOHN F. ROBERTS, WILLIAM D. ROSE, *and* PATRICK K. SUTHERLAND

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areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in Oklahoma

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—OKLAHOMA

By ROBERT O. FAY,[1] S. A. FRIEDMAN,[1] KENNETH S. JOHNSON,[1] JOHN F. ROBERTS,[1] WILLIAM D. ROSE,[1] and PATRICK K. SUTHERLAND[2]

ABSTRACT

The Carboniferous rocks of Oklahoma represent both the Mississippian and Pennsylvanian Systems. Sedimentary rocks make up almost all the Carboniferous in the geologic record, although minor occurrences of igneous and metamorphic rocks have been noted. Volcanic ash, pyroclastic flows, and tuff beds are present in Mississippian rocks of the Ouachita Mountains; some low-grade metasedimentary rocks occur in the core area of the Ouachitas.

Oklahoma's three major mountain systems—the Wichita, Arbuckle, and Ouachita—extend through the southern part of the State. They were formed during the Pennsylvanian Period and are adjacent to the Anadarko, Arkoma, Ardmore, and Marietta deep sedimentary basins, which contain the thickest Carboniferous sedimentary rocks. The northern part of the State contains the more stable cratonic shelf areas and the southwest flank of the Ozark uplift in the northeast.

Carboniferous outcrops are limited to the eastern half of Oklahoma. Pennsylvanian rocks are exposed in the Arkoma basin and nearby shelf areas to the north and west as well as in parts of the Arbuckle, Ouachita, and Ozark uplifts and in the Ardmore basin to the south. Mississippian rocks are more limited in extent and are exposed only in the Arbuckle, Ouachita, and Ozark uplifts. In general, Mississippian rocks consist of carbonate rocks and some shales and sandstones, although their lithology varies markedly from one depositional basin to another. The greatest thickness of Mississippian strata is the 3,300 m of flysch sedimentary rocks making up the Stanley Group of the Ouachita basin. Pennsylvanian rocks consist mostly of shale, but beds of sandstone, limestone, conglomerate, coal, and underclay are also present; these rocks attain their greatest thickness, 5,500 m, in the Arkoma basin.

The Pennsylvanian-Permian boundary now is tentatively placed at the top of the Herington Limestone Member of the Oscar Formation by the Oklahoma Geological Survey. This usage places rocks of Gearyan (Wolfcampian) age in the uppermost Pennsylvanian System rather than in the lowermost Permian.

The principal economic resources of Oklahoma's Carboniferous deposits are coal and petroleum products. In 1977, Oklahoma ranked 20th in the Nation in coal production and produced a record 5.3 million short tons. In 1976, the State ranked third in the Nation in natural-gas production (1,710,-586 million cubic feet) and fourth in crude-oil production (150,627,000 barrels); an estimated 69 percent of the gas and 60 percent of the oil produced came from Carboniferous rocks.

Less significant economic resources that have been produced from Carboniferous rocks include metallic ores (zinc, lead, and copper) and nonmetallic minerals (limestone, shale, tripoli, sandstone, chat, and iodine). Two major Carboniferous aquifers, the "Boone" (Mississippian) and the Vamoosa (Pennsylvanian), produce large quantities of ground water.

INTRODUCTION

By KENNETH S. JOHNSON

Oklahoma, a region of complex geology, contains a great thickness of Carboniferous sedimentary rocks and underwent major orogenic activity during the Pennsylvanian Period. In the southern part of the State are three major mountain systems—the Wichita, Arbuckle, and Ouachita (fig. 1)—which largely mark the southern boundary of the North American craton. The mountain regions, all of which were formed during Pennsylvanian orogenies, are adjacent to a series of deep sedimentary basins—the Anadarko, Arkoma, Ardmore, and Marietta—which received 5,000 to 7,000 m of Carboniferous sediments. The northern half of the State, on the other hand, contains the more stable cratonic shelf areas and the southwest flank of the Ozark uplift in the northeast. Much of the data presented in this report was collected in the comprehensive studies by Branson (1962), Ham and Wilson (1967), and Frezon and Dixon (1975). The reader is referred to these reports as well as to an atlas by Johnson and others (1972) that contains a series of generalized maps, cross sections, and a text on the geology and mineral resources of Oklahoma.

Outcrops of Carboniferous rocks are limited to the eastern half of Oklahoma (figs. 2, 3). Pennsyl-

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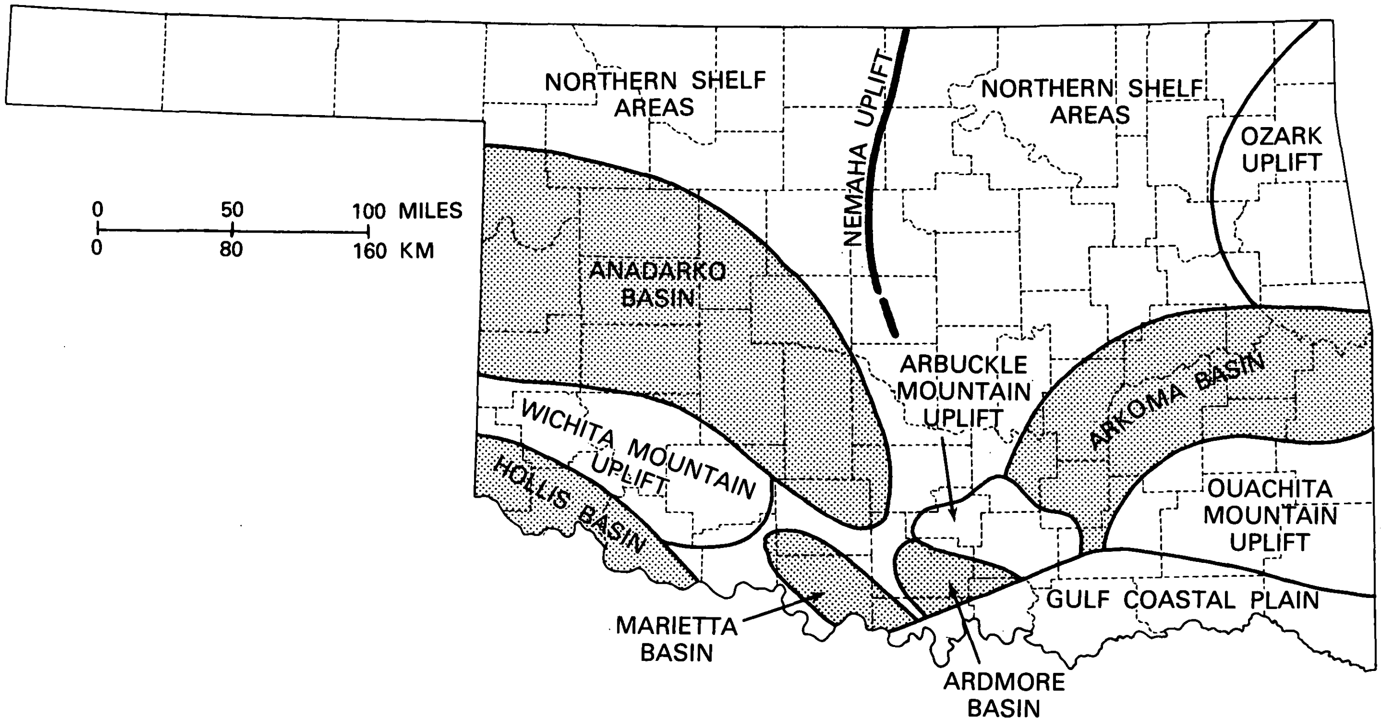

FIGURE 1.—Major geologic provinces of Oklahoma (from Johnson and others, 1972, p. 1).

vanian strata are extensively exposed in the Arkoma basin and nearby shelf areas to the north and west, as well as in parts of the Arbuckle, Ouachita, and Ozark uplifts and in the Ardmore basin to the south. Mississippian strata are more limited and are exposed only in the Arbuckle, Ouachita, and Ozark uplifts.

Carboniferous strata are well exposed in most parts of eastern Oklahoma. The area was not covered by Pleistocene glaciers; thus, only soil and vegetation conceal the beds of sandstone, shale, and limestone that make up almost all the stratigraphic section. Average annual precipitation ranges from about 85 cm in the central part of the State to 110–140 cm in the east.

Regional eastward tilting of the midcontinent of the United States during uplift of the Rocky Mountains in Late Cretaceous and Early Tertiary time established the dominant east-flowing river systems that cross the strike of Carboniferous strata in Oklahoma. The simple dendritic pattern is locally replaced by radial drainage along the flanks of the Ozark uplift and by poorly defined trellis drainage in much of the Ouachita Mountains and the Arkoma basin.

Topographic relief in most of eastern Oklahoma is low to moderate; west-dipping cuestas overlook broad shale plains. Bedrock exposures are abundant, and the many highways and maintained dirt roads provide easy access to them. In the Ozark uplift, the Ouachita Mountains, and parts of the Arkoma basin, the relief is moderate to high (as much as 500 m locally in the Ouachitas). Although bedrock is widely exposed in these provinces, ready access to these exposures is limited largely to the few highways and dirt roads that cross this more rugged terrain.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Oklahoma Geological Survey. The following series designations of the Carboniferous are now used by the Oklahoma Geological Survey: For the Mississippian, they are Kinderhookian and Osagean (Lower) and Meramecian and Chesterian (Upper). For the Pennsylvanian, they are Morrowan and "Atokan" (Lower); Desmoinesian and Missourian (Middle); Virgilian (Upper); and Gearyan (Upper?). Terms such as Springeran, Bendian, Lampasan, Atokan, and Wolfcampian appear to overlap other series and are in controversy. These terms have been used in a group or rock sense in Oklahoma and are placed in quotation marks in this report to indicate questionable usage and the need for restudy.

As noted in the table of contents, this report was prepared by Robert O. Fay, S. A. Friedman, Kenneth S. Johnson, and John F. Roberts, staff members of the Oklahoma Geological Survey; and by Patrick K. Sutherland, School of Geology and Geophysics, University of Oklahoma. William D. Rose, Oklahoma Geological Survey, coordinated the report.

HISTORY OF GEOLOGIC STUDIES

By Robert O. Fay

The early history of geologic studies in Oklahoma began with the Indians, who knew about salt, lead ore, coal, oil seeps, and limestone. The Indians kept no records, but they directed the early explorers to most of the deposits. The Spanish (1541–1803), French (1682–1803), and American (1803–11) explorers recorded many of their findings. For Spanish history, see Shipp (1881), Winship (1896), Thoburn (1916), and Thomas (1928). For French history, see Le Page du Pratz (1758), Parkman (1869), Hyde and Conard (1899), Goodspeed (1904), Bennitt and Stockbridge (1905), Thoburn (1916), Lewis (1924a,b,c), Lyon (1934), Hyde (1948), Beers (1957), and Boone (1968). For American history, see Austin (1804), John Sibley (1805), Lewis and others (1806), Pike (1810), G. C. Sibley (1812), Stoddard (1812), Coues (1895), Thwaites (1908), Thoburn (1916), Barker (1924), Foreman (1930, 1932), Jackson (1966), and Stout (1976).

On April 30, 1803, the Province of Louisiana was purchased from the French by the United States. On March 26, 1804, the land south of lat 33° N. was designated Territory of Orleans, and New Orleans, its capital; the land to the north was designated District of Louisiana, and St. Louis, its capital. On April 30, 1812, the Territory of Orleans became the State of Louisiana. On March 2, 1819, the Arkansas Territory was created, which included present-day Arkansas and Oklahoma; the land to the north was Missouri Territory. On August 10, 1821, Missouri became a State, and the land to the west was known as Missouri Territory. On May 28, 1828, the western part of Arkansas Territory was designated Indian Territory. On June 15, 1836, Arkansas became a State. On November 16, 1907, Indian Territory became the State of Oklahoma. (For more details, see Williams (1904) and Herndon (1922).)

Fort Smith was established in 1817 and became the center for southwestern explorations. Bearss and Gibson (1969) gave a history of Fort Smith. Schoolcraft (1819, 1821), Nuttall (1821), James (1823), Hinton (1834), Featherstonehaugh (1835), Shumard (1853), and Marcou (1854) published some of the earliest accounts of the geology of what is now Oklahoma and adjacent areas.

In his description of rocks near Fort Smith, James (1823, p. 410) first used the name Carboniferous Limestone in what is now Oklahoma: "Conybeare and Phillips [1822] apply this name to the limestone of the English coal measures (p. 340, pl. 1). *Compact limestone* is a name obviously inapplicable to the whole series of calcareous beds, occurring in connexion [sic] with the coal." Hinton (1834) described the general geology of the Ozarks, the Ouachitas, and the Great Plains, summarizing from previous sources. Shumard (1853) described the geology from Fort Smith to the Arbuckles and the region southward, showing the Carboniferous Limestone, below, and the Coal Measures, above. He described the red beds around the Wichita Mountains. Marcou (1854) described the geology along the Canadian River, distinguishing the Carboniferous Limestone, the Coal Measures, and the overlying New Red Sandstone.

The completion of the Missouri, Kansas, and Texas Railroad across Oklahoma in 1872 was followed by the opening of the coal mines at McAlester. Three oil wells were drilled near Chelsea in 1889. Asphaltite was known in the Ouachitas in 1890, and the first oil production in Oklahoma was recorded in 1891. In 1893, tripoli was found near Peoria, Ottawa County.

Chance (1890) studied the Choctaw coal fields. Hill (1891) wrote on the Ouachitas. Haworth and Kirk (1894) studied the Neosho River section. C. R. Keyes (1894, unpub. data) wrote the first formal oil and gas prospectus for John D. Rockefeller. Stevenson (1896) wrote on the geology of Indian Territory. Drake (1897) mapped eastern Oklahoma in 1896, noting accurately many details of the coal geology. Vaughan (1899) published on the Arbuckle Mountains. Taff (1899) described and mapped the McAlester-Lehigh area and described asphalt in the Choctaw Nation, following a treaty of 1897 to map the segregated coal lands in the Choctaw and Chickasaw Nations. White (1899) and Girty (1899) wrote about Taff's collections of invertebrate and plant fossils from near McAlester.

Taff (1899, 1902) continued his work with the help of G. I. Adams, S. H. Ball, J. W. Beede, S. W. Beyer, G. H. Girty, C. N. Gould, R. D. Mesler, G. B. Richardson, M. K. Shaler, C. D. Smith, E. O. Ulrich, and C. D. White. Almost 20 publications resulted on

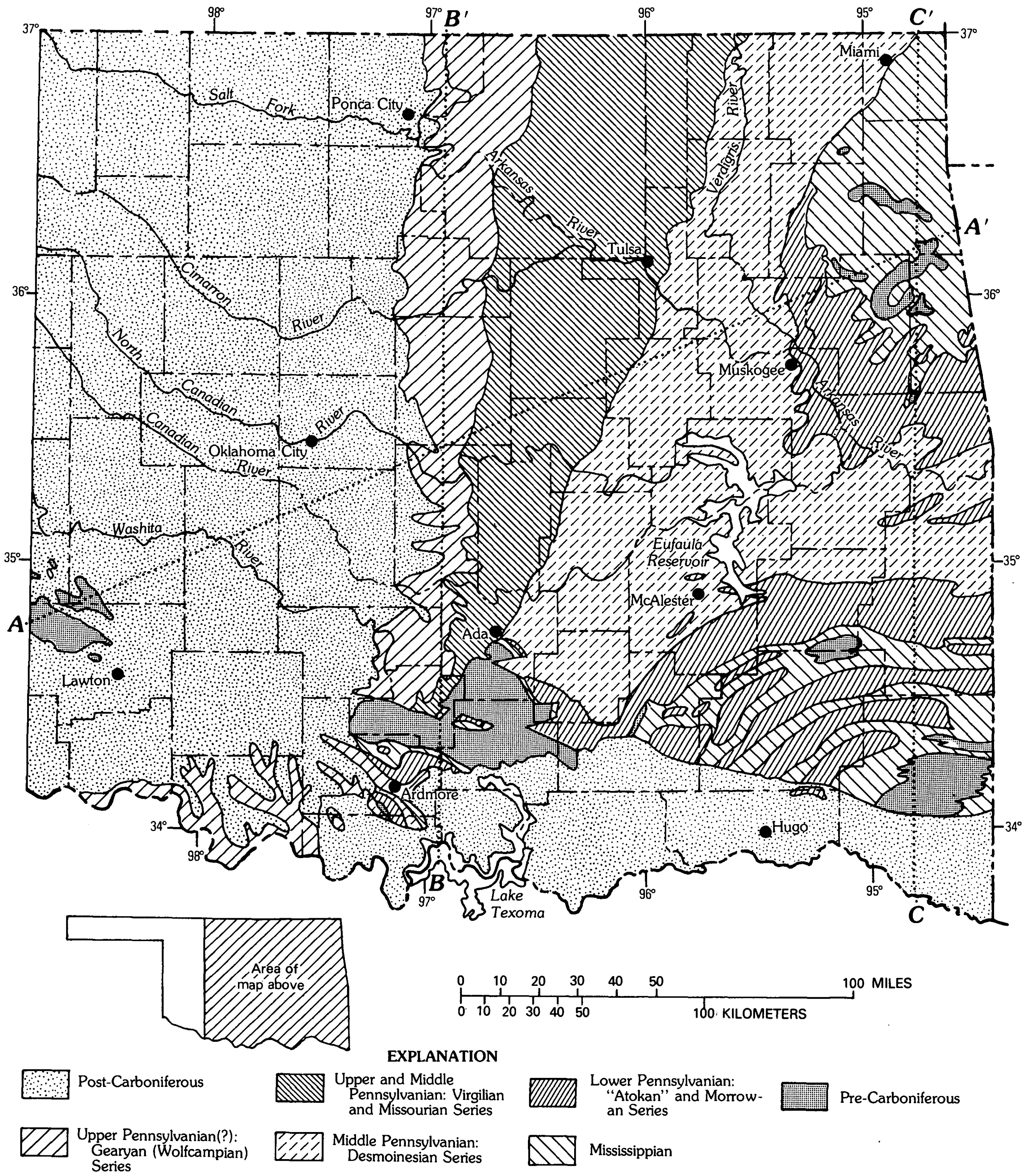

FIGURE 2.—Generalized geologic map showing outcrops of Carboniferous rocks in Oklahoma (from Miser, 1954; Johnson and others, 1972, p. 4). Cross sections shown in figure 3.

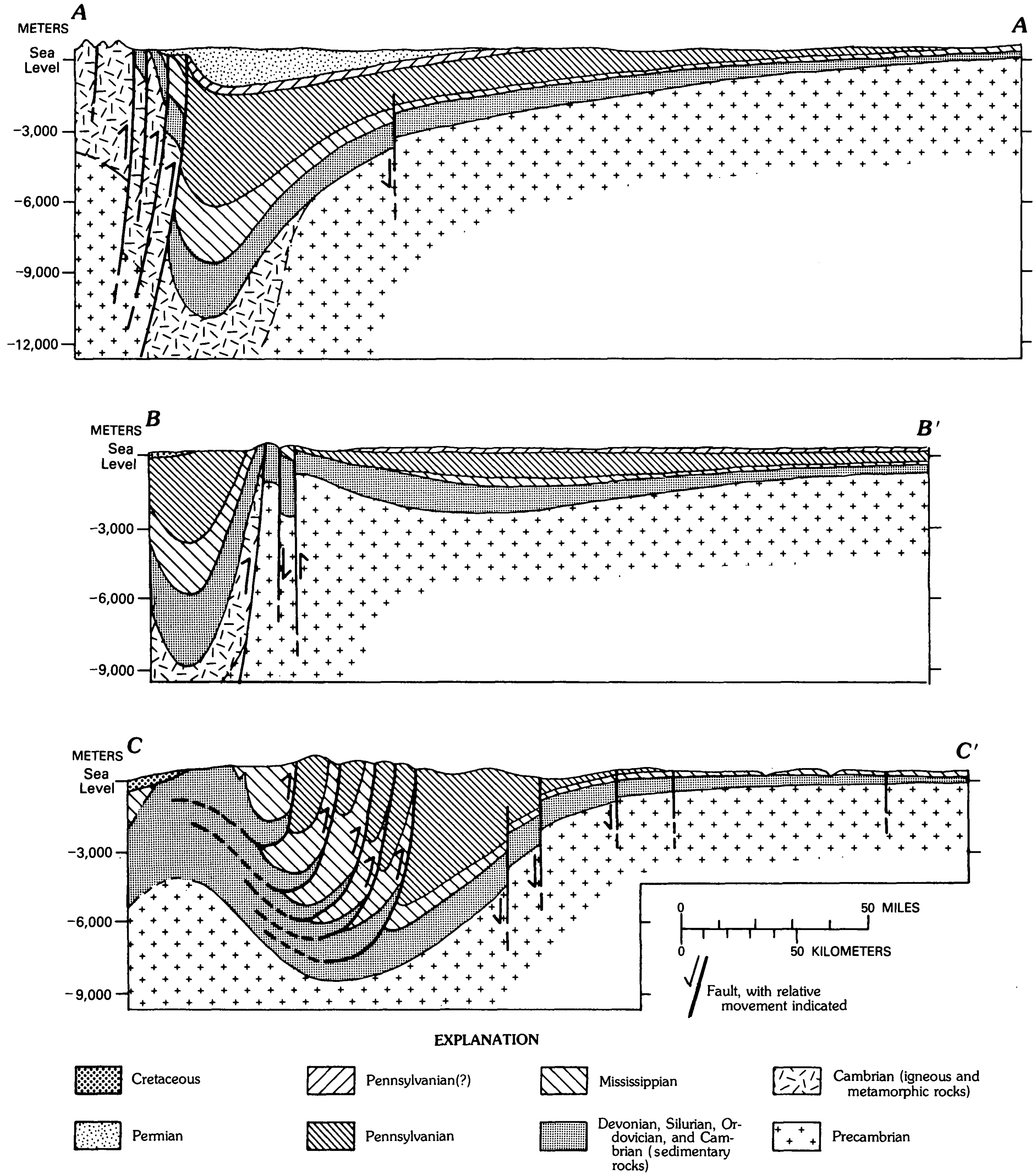

FIGURE 3.—Generalized cross sections through major geologic provinces of Oklahoma (from Johnson and others, 1972, p. 5). Lines of cross sections shown in figure 2.

the Ozarks, the Arkoma basin, the Ouachita Mountains, the Arbuckle Mountains, the Ardmore basin, the Criner Hills, and the Wichita Mountains. Maps of the McAlester, Windingstair, Tuskahoma, and Antlers Quadrangles were never published but were utilized by Miser in preparation of the first geologic map of Oklahoma (1926).

From 1900 to 1910, more than 30 additional articles were published on all phases of the Carboniferous. In 1907, the Miami-Picher zinc field was discovered in Mississippian rocks, and most of the 344 gas wells and 7,850 oil wells in Oklahoma were producing from the Carboniferous strata. The new Oklahoma Geological Survey was established July 25, 1908.

From 1910 to 1930, many changes took place. In April 1912, the Cushing field was discovered; oil production from this field amounted to 3 percent of the world's total for that year. In 1913, most of the 1,052 producing gas wells and 20,620 oil wells in Oklahoma were producing from Carboniferous rocks. Everett Carpenter organized the Empire Gas and Fuel Co. in 1913, hiring 250 geologists and initiating subsurface studies in Oklahoma. The Oklahoma Geological Survey first used automobiles in 1915; previously horses and mules were used. In June 1915, the Oklahoma Corporation Commission first regulated oil production. The American Association of Petroleum Geologists was organized on January 7, 1916, in Norman, Okla. Core drilling was first used in Oklahoma in 1919. The secondary recovery of oil by waterflooding began in 1920. Geophysical work was begun in 1921 and was based upon a method that was used to locate cannons in World War I; the first geophysical company consisted of W. P. Haseman, head of the Physics Department at the University of Oklahoma, and D. W. Ohern, Irving Perrine, and J. C. Karcher. By 1924, micropaleontology had become popular, rotary drilling had been introduced in Oklahoma, and well cuttings were being preserved. In 1929, electric logs were first used. These developments enabled geologists to have a three-dimensional concept of Carboniferous rocks in Oklahoma. A flood of published information followed.

From 1910 to the present, many hundreds of articles have been written on the Carboniferous of Oklahoma. Gould (1925) gave the first stratigraphic index. Branson and Jordan (1964) and Branson and others (1967) published indexes to mapping in Oklahoma. Now, more than 100 surface maps and 250 subsurface maps exist on the Carboniferous of Oklahoma. McKnight and Fischer (1970) and Gibson (1972) provided histories of the Tri-State mineral district and gave extensive bibliographies. From 1958 to the present, K. S. Johnson, E. A. Ham, and others on the Oklahoma Geological Survey staff have compiled annual bibliographies of Oklahoma geology, which have been published in Oklahoma Geology Notes.

Concerning nomenclature and regional stratigraphy, the following are excellent references: Williams (1891), Wilmarth (1925), Moore (1940), Branson (1962), and Frezon and Dixon (1975).

Various workers have assigned different locations for the Pennsylvanian-Permian boundary over the years (see Moore, 1940, figs. 3, 4). On the new geologic maps of Oklahoma, which have been published at a scale of 1:250,000 by the Oklahoma Geological Survey and the U.S. Geological Survey as part of a series of hydrologic atlases, rocks of Gearyan ("Wolfcampian") age have been considered uppermost Pennsylvanian rather than Permian. This usage follows the work of Dott (1932), Green (1936), Branson (1962), Clendening (1971), and Wilson and Rashid (1975). In figure 2 of the present report, Gearyan rocks are included in the map unit showing rocks of questionable Pennsylvanian age.

GEOLOGIC SETTING

By KENNETH S. JOHNSON

UNDERLYING ROCKS

The base of the Carboniferous throughout most of Oklahoma lies within a dark-gray to black cherty shale of Late Devonian and Early Mississippian age. The shale is called the Woodford Shale in most parts of the State, but the term Chattanooga Shale is used in the northeast on the flanks of the Ozark uplift. In the Ouachita Mountains, the base of the Carboniferous is in the Lower Silurian-Lower Mississippian Arkansas Novaculite, a siliceous chertlike sedimentary rock, the upper part of which is generally equivalent to the Woodford Shale. The Woodford Shale or equivalent strata are absent in parts of the Panhandle, in the Hollis basin, over the Wichita uplift, and over other smaller uplifts; in these areas, younger Mississippian and even Pennsylvanian strata are at the base of the Carboniferous locally.

At the base of the Woodford is a major pre-Upper Devonian unconformity that extends over most of the midcontinent region. The Woodford therefore rests with low-angle unconformity upon strata of Ordovician, Silurian, and Early Devonian age (Tarr and others, 1965). Silurian and Devonian rocks underlie the Woodford Shale or Arkansas Novaculite in the deeper parts of most basins, whereas Ordovician strata underlie the pre-Woodford unconformity on the north flanks of the Arkoma and Anadarko basins and in parts of the Arkoma, Ardmore, and Hollis basins.

In areas where the Woodford Shale or Arkansas Novaculite is present, the Devonian-Mississippian contact is transitional and can be only approximated by micropaleontologic study of pollen and (or) conodonts. Black shale and siliceous sediments were deposited without apparent interruption in most parts of the State from Late Devonian through Early Mississippian (Kinderhookian) time. In the Panhandle and in the Hollis basin, limestones of Osagean through Meramecian age are at the base of the Carboniferous. Mississippian strata were eroded during subsequent Pennsylvanian uplift of the Wichita block and smaller blocks extending northward from the Arbuckle Mountains across central Oklahoma (Hunton arch and Oklahoma City uplift-Nemaha uplift). In these areas, Middle and Upper Pennsylvanian clastic sedimentary rocks lie unconformably upon Cambrian through Devonian sedimentary rocks or, in much of the Wichita uplift, upon a Cambrian basement of granite, rhyolite, and gabbro.

OVERLYING ROCKS

Carboniferous strata are now widely exposed in the eastern third of Oklahoma. In the west, however, Pennsylvanian(?) rocks of Gearyan age are overlain by Lower Permian strata consisting chiefly of red-bed clastic rocks. In southeastern Oklahoma, Permian rocks are absent, and the Carboniferous is overlain by sands of the Trinity Group of Early Cretaceous age.

The contact with overlying Permian strata in western Oklahoma appears to be conformable. The stratigraphic relations and lithologies of rocks on both sides of the boundary indicate that Late Pennsylvanian depositional conditions continued into the Permian without appreciable break, except for a change into red beds.

Cretaceous strata in the southeast rest with angular unconformity upon the folded Carboniferous rocks of the Ouachita system and the Ardmore basin.

STRUCTURAL EVENTS

Carboniferous time in Oklahoma was characterized by formation of deep sedimentary basins and, in the southern half of the State, by formation of the mountain systems. All sedimentary basins—including the Anadarko, Arkoma, Ardmore, Marietta, Hollis basins and the Ouachita geosyncline—had their principal period of downwarping from Late Mississippian through Pennsylvanian time. These basins typically are elongate and now contain some 3,000–12,000 m of sedimentary rocks; Carboniferous strata constitute 50–75 percent of the total thickness in each basin (fig. 3). The three mountain belts—the Wichita, Arbuckle, and Ouachita—were the sites of folding, faulting, and uplift during several orogenic pulses that took place throughout the Pennsylvanian Period. Principal studies upon which we have relied heavily for discussion of structural events are those by Cline and others (1959), Flawn and others (1961), Branson (1962), Ham and others (1964), Ham and Wilson (1967), American Association of Petroleum Geologists and others (1968, 1975), Ham (1969), Frezon and Dixon (1975), and Decker and Black (1976).

During the first half of the Mississippian Period, shallow seas covered all of Oklahoma. Limestone and interbedded chert were the predominant sediments laid down upon the Upper Devonian-Lower Mississippian Woodford Shale in most areas, whereas deposition of the Arkansas Novaculite continued in the Ouachita geosyncline. The widespread beds of Lower Mississippian limestone are the youngest or last of the thick sequence of carbonate rocks that attest general crustal stability in Oklahoma during early and middle Paleozoic time.

In the last half of the Mississippian Period, shale and sandstone were predominant; major sites of deposition were the rapidly subsiding basins in southern Oklahoma. Principal formations of southern Oklahoma (excluding the Ouachitas) are the Sycamore Limestone, Delaware Creek Shale, and Goddard Shale. These strata have a total thickness of 500–2,000 m in the Ardmore and eastern Anadarko basins and nearby areas. The greatest thickness of Mississippian strata is the 3,300 m of flysch sedimentary rocks making up the Stanley Group of the Ouachita basin. Mississippian strata in central and north-central Oklahoma have been largely removed by Early Pennsylvanian uplift and erosion, and the remaining Mississippian rocks consist generally of 50–200 m of cherty limestone that thickens to the west and reaches a thickness of 1,500 m in the western Anadarko basin.

The Pennsylvanian Period was the principal time of crustal unrest in Oklahoma—a time of both orogeny and basinal subsidence in the south and of epeirogenic movement in the north. Preexisting sedimentary rocks and the underlying basement rocks of the Wichita, Arbuckle, and Ouachita Mountain areas were complexly folded, faulted, and thrust upward into major mountains, while the nearby basins subsided more rapidly and received the greatly increased sediment load eroded from the highlands (fig. 3). Orogenies took place during all epochs of the Pennsylvanian Period, but different areas were affected to different degrees by each pulse.

Pennsylvanian rocks comprise mostly marine shale, but beds of sandstone, limestone, conglomerate, coal, and underclay are also present. The strata are commonly 600–1,500 m thick but are as much as 5,000 m thick in the Anadarko basin, 4,500 m in the Ardmore basin, 4,000 m in the Marietta basin, and 5,500 m in the Arkoma basin. In fact, each of these basins, as well as the eastward continuation of the Arkoma basin in Arkansas, contains a greater thickness of Pennsylvanian strata than does any other comparable area in the United States.

The major Pennsylvanian orogeny, commonly called the Wichita orogeny (Morrowan and early "Atokan"), was characterized by strong folding and uplift of as much as 3,000–4,500 m in the Wichitas and in the Criner Hills south of Ardmore. Conglomerate and "granite wash" (a local name given to coarse arkosic detritus) were commonly deposited near major uplifts. These coarse sedimentary rocks grade into sandstone and shale toward the middle of the basins. A broad, north-trending arch across central Oklahoma was raised above sea level during this time; along its axis was a narrow belt of fault-block mountains (Nemaha uplift) extending northward from the Oklahoma City area into Kansas. The uplift was accompanied by erosion that removed part or all the pre-Pennsylvanian sediments from the mountain uplifts and the central Oklahoma arch. In fact, the unconformity at the base of Pennsylvanian rocks is the most profound Paleozoic unconformity in Oklahoma and can be recognized everywhere but in the deeper parts of major basins.

Principal pulses of folding and uplift in the Ouachita Mountains began in Mississippian time and continued into the Permian; these structural movements are referred to as the Ouachita orogeny. The thick sequence of dark-gray shales, cherts, and flysch sedimentary rocks in the Ouachita geosyncline was complexly folded and thrust faulted. The Ouachitas contain a series of south-dipping thrust faults in the northern belt, where the Choctaw fault outlines the frontal (north) edge of the mountain system. In the southern belt, the major faults dip north. After the deformation of the Ouachita trough, basinal downwarping shifted northward into the Arkoma basin during "Atokan" and Desmoinesian time and then ceased after the folding and faulting of the Arkoma basin. Of special importance in the Arkoma basin and northeastern Oklahoma are the coal beds formed during Desmoinesian time.

The last major Pennsylvanian orogeny, called the Arbuckle orogeny, was one of strong compression and uplift during Virgilian time. It affected all mountain areas of the south and is represented by most of the prominent folding in the Ardmore, Marietta, and Anadarko basins. Much of the thrusting in the Ouachita Mountains probably also took place in late Virgilian time. Thus, by the end of the Pennsylvanian Period, the mountain systems of Oklahoma were substantially as we know them today, although subsequent gentle uplift and accompanying erosion have cut more deeply into underlying rocks.

Post-Carboniferous structural movements were largely confined to epeirogenic raising and lowering of broad regions in Oklahoma. In Early Permian time, the Wichitas, Arbuckles, Ouachitas, and Ozarks were still fairly high land areas, and they supplied sand and mud to shallow seas that covered the Anadarko basin and other parts of western Oklahoma. By Late Permian time, the Wichitas were largely buried by sediment derived mainly from the lowland areas of the Ouachitas and Ozarks in the east. Small faults and flexures in Permian sedimentary rocks attest minor movement of preexisting major faults, principally along the margins of sedimentary basins.

During the Cretaceous Period, southern and western parts of Oklahoma were gently depressd to accept the last incursion of marine waters into the State. Formation of the Rocky Mountains in Late Cretaceous and Early Tertiary time elevated the western part of the State, imparting an eastward tilt to the land surface that has persisted to the present.

Major drainage systems of today were initiated during Pleistocene time. Continental glaciers extended southward only to northeastern Kansas, but meltwater from Rocky Mountain glaciers and snowfields provided much of the streamflow that helped strip away post-Carboniferous strata and thus exhume the major mountain systems.

MISSISSIPPIAN AND LOWER PENNSYLVANIAN STRATIGRAPHY

By Patrick K. Sutherland

Strata of Mississippian and Early Pennsylvanian age crop out in five regions in Oklahoma: (1) the southwestern Ozark region, in the northeast; (2) the frontal Ouachita Mountains; (3) the central Ouachita Mountains, in the southeast; (4) the northeastern Arbuckle Mountains, in the south-central part of the State; and (5) the Ardmore basin and southwestern Arbuckle Mountains, in the south (fig. 4). These areas show marked differences in lithologic character, thickness, and depositional pattern.

OZARK REGION

Mississippian and Lower Pennsylvanian strata in the Ozark region of northeastern Oklahoma are dominated by carbonate rocks. Unconformities are numerous. Represented in figure 5 is a typical shallow-water platform facies, which is thinner than the clastic, geosynclinal facies that crops out in the Ouachita Mountains. The maximum thicknesses recorded for Mississippian formations in the Ozark region of Oklahoma total about 250 m, but the total preserved thickness for the Mississippian in any local area does not exceed 150 m. The maximum recorded thickness for the Lower Pennsylvanian Morrowan Series is 94 m, and that for the "Atokan" Series, about 185 m.

The Upper Mississippian Pitkin Formation is truncated northward by pre-Pennsylvanian erosion and is absent north of T. 18 N. The succeeding Morrowan Series is truncated by pre-Atokan erosion in T. 20 N. In T. 22 N., Desmoinesian strata overlap the "Atoka" Formation and rest directly upon eroded Mississippian strata.

KINDERHOOKIAN, OSAGEAN, AND MERAMECIAN SERIES

The Chattanooga Shale, as much as 20 m thick in northeastern Oklahoma, ranges in age from Late Devonian to Early Mississippian (Kinderhookian) and correlates with the Woodford Shale in southern Oklahoma (Hass, 1956).

The Chattanooga is overlain unconformably by the "Boone" Group, which consists of beds of chert and limestone. This group ranges in age from latest Kinderhookian to early Meramecian but is predominantly Osagean. Included are the St. Joe, Reeds Spring, and Keokuk Formations. Their maximum recorded thicknesses are 12, 55, and 75 m, respectively (Huffman, 1958); the thickness of each averages much less, however, and all are missing in the southern part of the Ozark outcrop area, where the Moorefield Formation rests directly on the Chattanooga. The Keokuk is particularly distinctive lithologically, consisting of white- to buff-weathering chert that forms distinctive fractured rubble surfaces. It contains abundant fossils, mostly in the form of molds and casts. Various formations of the "Boone" Group are overlain unconformably by the Moorefield Formation, the Hindsville Limestone, or, locally, the Fayetteville Shale.

The Moorefield Formation, predominantly of Meramecian age, consists mostly of argillaceous limestones, but other facies include oolitic and pelmatozoan grainstones and calcareous siltstones. Four facies were given local member names by Huffman (1958). The formation has a maximum thickness of about 30 m but is missing in Craig and Ottawa Counties, where the Hindsville rests unconformably on the "Boone" cherts.

CHESTERIAN SERIES

The Hindsville Limestone, Fayetteville Shale, and Pitkin Limestone, all of Chesterian age, form a continuous depositional sequence in the Ozark region. The Hindsville is a widely distributed fossiliferous limestone that rests unconformably on the Moorefield Formation or on the "Boone." It has a maximum thickness of 15 m, but it averages 8–10 m (Huffman, 1958).

The Hindsville Limestone is overlain conformably by the Fayetteville Shale. The Fayetteville consists predominantly of black or gray-green shale, but it is interbedded locally near the base or near the top with dark nodular layers of carbonate mudstone. The formation has a maximum thickness of about 50 m.

The Pitkin Limestone conformably overlies the Fayetteville Shale and shows marked local variations in facies, ranging from crossbedded oolites to skeletal grainstones to dark carbonate mudstones. The formation is typically about 10 m thick but has a maximum observed thickness of 25 m (Huffman, 1958). The formation is unconformably overlain by the Pennsylvanian Sausbee Formation.

POST-CHESTERIAN EROSIONAL SURFACE

The Pitkin Limestone is regionally truncated northward in Oklahoma along a highly irregular line; the most northerly exposures are in T. 18 N., at the southern edge of Mayes County. Farther

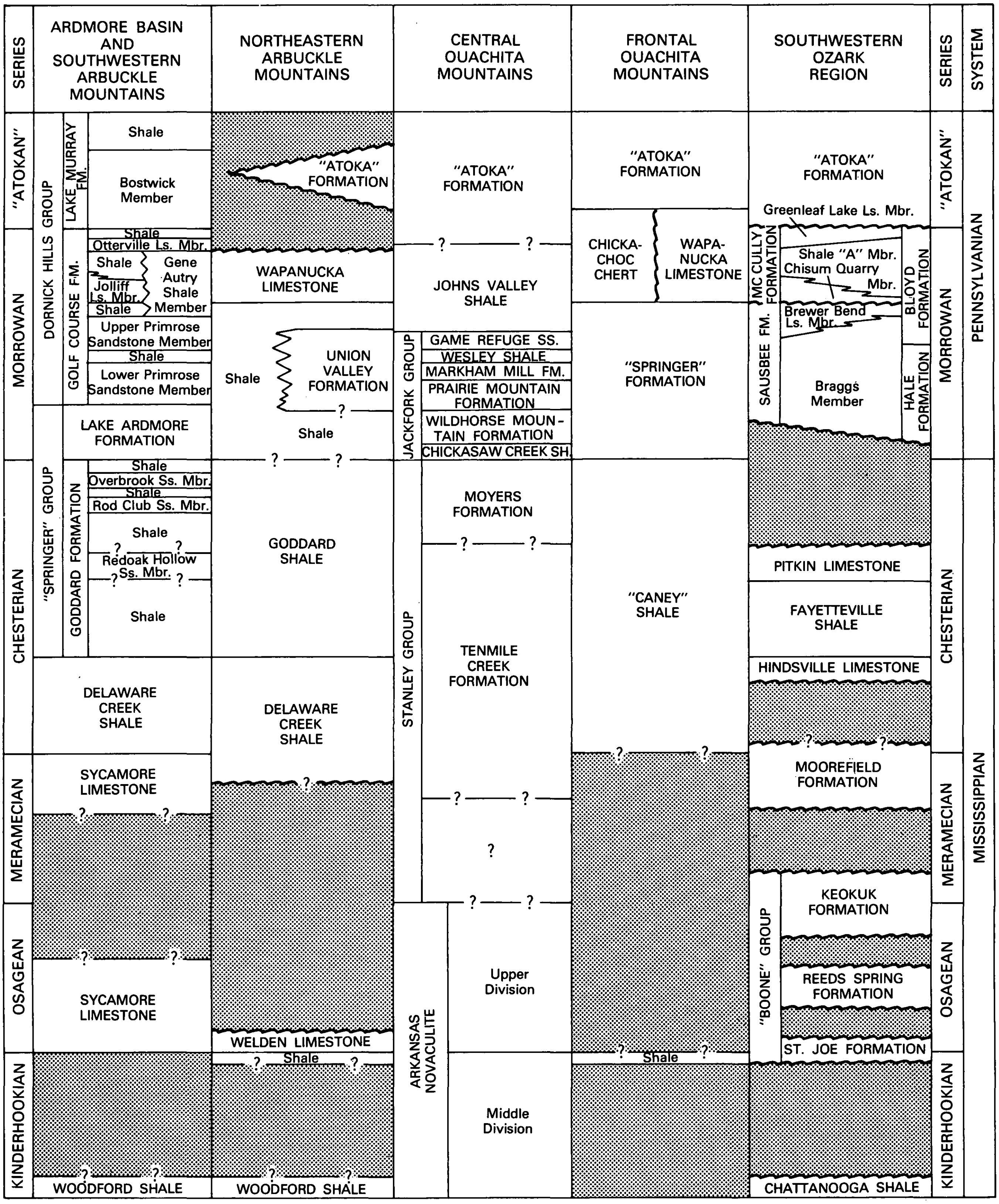

FIGURE 4.—Correlation chart of Mississippian and Lower Pennsylvanian rocks in five outcrop regions of Oklahoma.

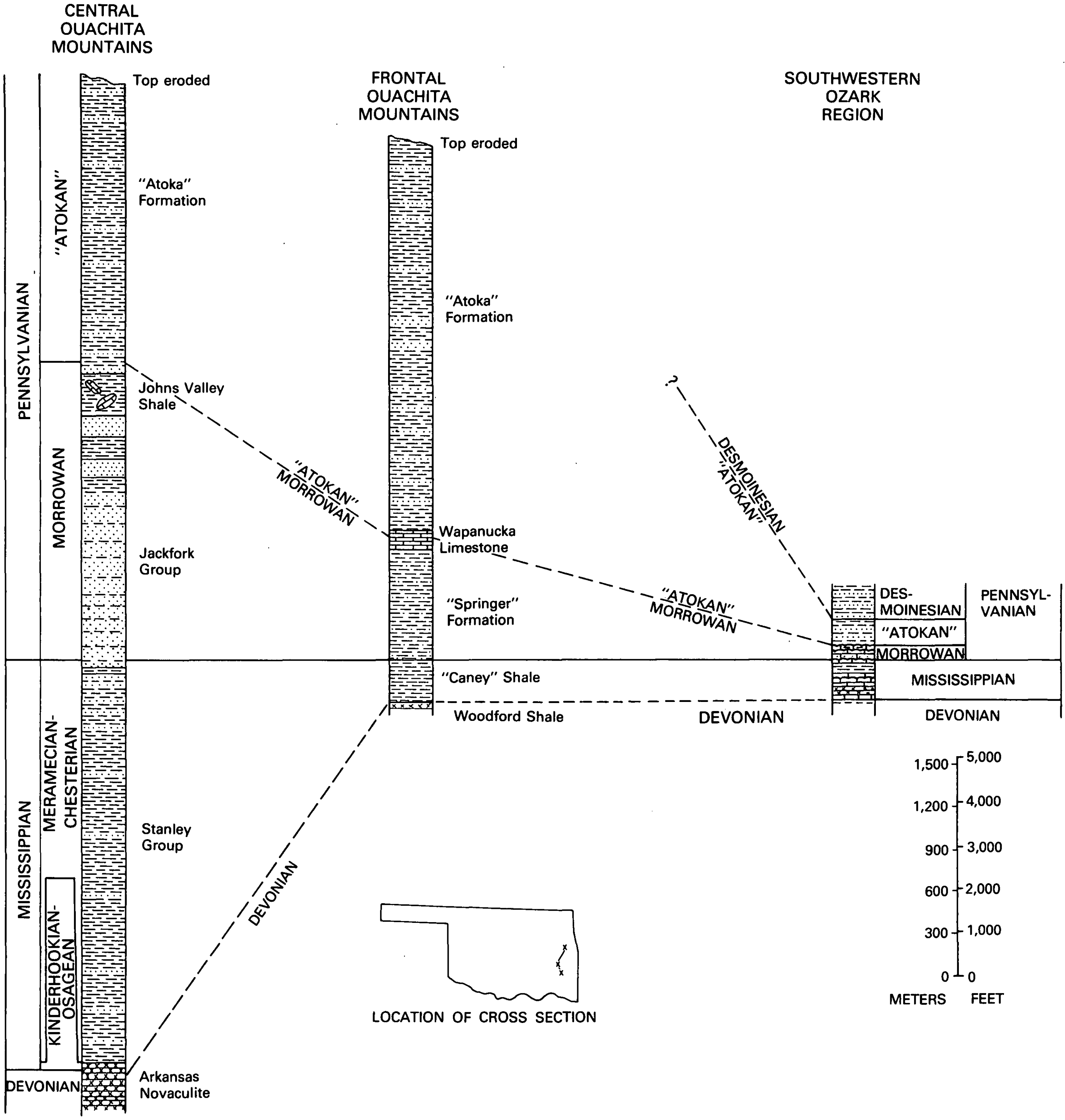

FIGURE 5.—Stratigraphic cross section from central Ouachita Mountains northward to southwestern Ozark region.

north, the Pennsylvanian rests directly on the Fayetteville Shale; locally it rests on the Hindsville Limestone. The post-Mississippian unconformity in northeastern Oklahoma has a regional relief of as much as 25 m (Sutherland and Henry, 1977b).

The magnitude of the post-Pitkin unconformity decreases eastward. In Searcy County, Ark., about 240 km to the east, the Imo Formation of late Chesterian age partly fills the gap (Saunders and others, 1977). The Imo overlies the Pitkin Formation and is in turn overlain unconformably by strata of Morrowan age.

In Arkansas, reworked Pitkin pebbles are found in the Chickasaw Creek Formation at the base of the Jackfork Group.

MORROWAN SERIES

In northwestern Arkansas, the Pennsylvanian Morrowan Series is divided into the Hale and Bloyd Formations (in ascending order). These units can be recognized in Oklahoma only in the eastern half of Adair County. Farther west, the lithologic distinction is lost, as there is a marked westward increase in the percentage of limestone and a corresponding decrease in the percentage of terrigenous-rock types. Sutherland and Henry (1977a) divided this carbonate facies into the Sausbee and McCully Formations on the basis of a regional disconformity at the top of the Sausbee. This break coincides in Washington County, Ark., with a regional unconformity at the base of the Dye Shale Member of the Bloyd Formation. The Sausbee Formation consists typically of skeletal grainstones interbedded with shale (Braggs Member), overlain by beds of algal wackestone and mudstone (Brewer Bend Limestone Member). The formation has a maximum thickness of 61 m. The overlying McCully Formation is composed of interbedded limestones and shales and has a maximum recorded thickness of 23 m.

"ATOKAN" SERIES

The "Atokan" Series, which consists of the "Atoka" Formation, crops out in a wide belt along the south and west flanks of the Ozark dome in northeastern Oklahoma. It consists of interbedded thick shales and thinner sandstones and a few beds of thin discontinuous impure limestone. Wilson (1935) gave member names to six of the ridge-forming sandstones in Muskogee County, but Blythe (1959) was unable to differentiate most of these members in areas north of Muskogee County. The "Atoka" Formation is about 185 m thick in Muskogee County.

The "Atoka" Formation is truncated northward in T. 23 N., R. 19 E., by a regional unconformity at the base of the overlying McAlester Formation.

FRONTAL OUACHITA MOUNTAINS

Describing and interpreting Carboniferous strata in the frontal Ouachitas is complicated by both faulting and lateral facies changes in each of the fault blocks from north to south. The column in figure 4 labeled "Frontal Ouachita Mountains" is a composite section for the several fault blocks north of the Ti Valley fault. The block directly south of the Choctaw fault, the leading edge of the frontal Ouachitas, exposes, in ascending order, the "Caney" (Delaware Creek) Shale, "Springer" (Goddard) Formation, Wapanucka Limestone, and "Atoka" Formation. Farther south, the block south of the Katy Club fault exposes the "Caney," "Springer," Chickachoc Chert, and "Atoka." The block south of the Pine Mountain fault, still north of the Ti Valley fault, exposes the Woodford Shale (Devonian), "Caney," "Springer," and "Atoka."

KINDERHOOKIAN SERIES

Rocks of Kinderhookian age have been reported from only a single locality in the frontal Ouachitas. Hass and Huddle (1965) recovered conodonts of this age from the basal 0.15 m of shale that directly overlies the Woodford Shale at a locality near Pine Top School in Pittsburg County, Okla., between the Pine Mountain and Ti Valley faults.

CHESTERIAN SERIES

The "Caney" Shale of Chesterian age reaches a maximum thickness in the frontal Ouachita Mountains of possibly 275 m, but in the frontal block directly south of the Choctaw fault, only the upper 15 m or so is exposed. Farther south, in the fault block between the Pine Mountain and Ti Valley faults, the "Caney" Shale rests directly on the Woodford Shale.

The Chesterian age assignment for the "Caney" Shale is based on several occurrences of goniatites and microfossils.

MORROWAN SERIES

Overlying the "Caney" Shale is the "Springer" Formation, which contains, at a few localities, goniatites of Morrowan age that correlate with the Hale Formation and the Brentwood Limestone Member of the Bloyd Formation in northwestern Arkansas (Gordon and Stone, 1977). Hendricks and others (1947) stated that the "Springer" apparently rests conformably on the "Caney" but that both units are poorly exposed. The Mississippian-Pennsylvanian boundary has in fact not been established precisely in this outcrop belt. They also reported a thickness of as much as 760 m for the "Springer" in the fault block directly southeast of the Choctaw fault. The "Springer" sequence was recently classed as Goddard (Upper Mississippian) in the Ouachita Mountains by the Oklahoma Geological Survey (see Hart, 1974, sheet 1).

The Wapanucka Limestone overlies the "Springer" in the frontal fault ridges of the Ouachitas, where it consists of interbedded spiculiferous packstones, carbonate mudstones, pelmatozoan and oolitic grainstones, and shales and is typically about 90 m thick. Basinward (southward), successive fault blocks expose changes to the Chickachoc Chert facies, with which the Wapanucka is correlative. This facies consists predominantly of shale interbedded with as much as 10 layers of dark-gray to black spiculite and a few thin beds of spiculiferous limestone. The thickness is typically 180–215 m. Conodonts have been recovered from this facies, as have rare goniatites (Gordon and Sutherland, 1975). The conodonts make possible a correlation between the Chickachoc and Wapanucka (Sutherland and Grayson, 1977). These units are equivalent in age to the Dye Shale and Kessler Limestone Members of the Bloyd Formation plus the lower part of the "Atoka" Formation in northwestern Arkansas and the McCully Formation and lower part of the "Atoka" Formation in northeastern Oklahoma.

"ATOKAN" SERIES

The uppermost parts of the Wapanucka Limestone and the Chickachoc Chert are "Atokan" in age, on the basis of conodonts (Sutherland and Grayson, 1977), and these units are overlain conformably in the frontal Ouachitas by the "Atoka" Formation. In this area, the "Atoka" consists mostly of gray silty, micaceous shale containing a few beds of medium-grained sandstone. The "Atoka" is poorly exposed, and only the lower part is preserved in the frontal Ouachitas, where Hendricks and others (1947) estimated the maximum thickness preserved to be 2,750 m.

CENTRAL OUACHITA MOUNTAINS

KINDERHOOKIAN AND OSAGEAN SERIES

The Kinderhookian and Osagean Series are presumed to be represented in the Ouachita Mountains of Oklahoma; the location depends upon the distribution in Oklahoma of Hass' (1951) middle and upper divisions of the Arkansas Novaculite. Most of the Arkansas Novaculite is Devonian in age, but Hass (1951) recorded a Kinderhookian age on the basis of conodonts for the top 8.5 m of the middle division of the formation at Caddo Gap, Montgomery County, Ark. In addition, he assigned a tentative latest Kinderhookian or Osagean age for another conodont collection made 25 m below the top of the 38-m-thick upper division of the formation at a locality in Polk County, Ark. Thus, Hass considered the upper 47 m of the formation to be Mississippian in age in Arkansas. Hass (1951) quoted H. D. Miser as stating that the upper division of the Arkansas Novaculite occurred in Oklahoma only in McCurtain County. The only conodonts recovered by Hass (1951) from the Arkansas Novaculite on Black Knob Ridge, near Atoka, Okla., were Devonian in age.

MERAMECIAN AND CHESTERIAN SERIES

Rocks of the Stanley Group are conformable with the underlying Arkansas Novaculite in McCurtain County, Okla. (Honess, 1923), and are generally so in the Potato Hills and at Black Knob Ridge, although a local conglomerate is at the base at some localities on Black Knob Ridge (Goldstein and Hendricks, 1962). Nowhere can an unfaulted sequence of the Stanley be seen, but a maximum thickness of 3,300 m has been estimated in the central Ouachitas. The group thins abruptly westward and northward toward the frontal Ouachitas (Cline, 1960). Harlton (1938) divided the Stanley Group into the Tenmile Creek, Moyers, and Chickasaw Creek Formations on the basis of the occurrence of several thin siliceous shales that are apparently widespread and locally mappable. The Stanley is composed predominantly of black shale, but sandstone beds are more common in the upper part. The Chickasaw Creek Formation was originally the basal unit of the Jackfork (Taff, 1902), and this usage is now followed by the Oklahoma Geological Survey (see Briggs, 1973, p. 8, 9).

The Stanley Group contains few fossils. Hass (1950) collected conodonts of Meramecian age from the lower part of the group in both Oklahoma and Arkansas. Higher conodonts, collected from 23 to 45 m above the base of the Stanley in Arkansas, were believed by Hass to be of Meramecian age but are now considered to be of early Chesterian age (Gordon and Stone, 1977). Plant fossils of Chesterian age have been recovered from the upper middle part of the group in Arkansas, and marine invertebrate fossils of Chesterian age have been recovered from erratic blocks of the Stanley in the frontal belt of the Ouachitas in Arkansas (Gordon and Stone, 1977). Also, the assignment of a Chesterian age for most of the Stanley is supported by its stratigraphic position. It is gradational with the overlying Jackfork Group. Chesterian plants, probably reworked, have been collected from the basal beds of the Jackfork (see the following section).

MORROWAN SERIES

The Jackfork Group is conformable with the underlying Stanley Group. Sandstone is the prevailing rock type, although shales make up as much as 40 percent of the group in some areas (Shelburne, 1960). The Jackfork is more resistant to erosion than the underlying Stanley Group, which is composed mostly of shale, and is one of the main ridge-forming units in the Ouachitas. The Jackfork Group is typically 1,750 m thick in the central Oklahoma Ouachitas. Cline (1960) and Shelburne (1960) recorded as much as 1,980 m in northern McCurtain County, Okla. These authors reported that the group thins abruptly northward toward the frontal Ouachitas. The Jackfork was divided by Harlton (1938) into the following formations, listed in ascending order: Wildhorse Mountain, Prairie Mountain, Markham Mill, Wesley, and Game Refuge. However, current usage by the Oklahoma Geological Survey includes the Chickasaw Creek Formation at the base of the Jackfork, according to Taff's (1902) original definition (see Briggs, 1973, p. 8, 9).

The Jackfork Group contains few fossils, and considerable disagreement exists in the literature regarding its precise age. Of particular importance is the recovery from the lowermost beds of the group, a short distance west of Talihina, Okla., of plants of Chesterian age (Gordon and Stone, 1977). These plants are possibly reworked (Gordon and Stone, 1977, p. 81), however, and the sedimentary rocks with which they are associated may be Morrowan in age. Collections of marine invertebrate fossils from the middle and upper parts of the Jackfork Group in the vicinity of Little Rock, Ark., include several poorly preserved but identifiable Morrowan species of goniatites and brachiopods (Gordon and Stone, 1977). The basal Chickasaw Creek Formation in Arkansas contains reworked Pitkin pebbles.

The Johns Valley Shale overlies the Jackfork Group conformably. It is well known for its great variety of erratic limestone boulders, mostly of the Arbuckle facies, that range in age from Cambrian to Early Pennsylvanian and for its huge slump blocks, some more than 900 m in length, of "Caney" Shale. The formation is typically 130–275 m thick in the central Ouachitas in Oklahoma (Cline, 1960). Gordon and Stone (1977) recorded a maximum thickness of 565 m, but they did not give a locality.

An indigenous fauna from the lower part of the Johns Valley Shale, from localities in both Oklahoma and Arkansas, includes cephalopods of the *Branneroceras branneri* zone (Gordon and Stone, 1977). This fauna occurs also in the upper part of the "Springer" Formation in the frontal Ouachitas and in the Brewer Bend Limestone Member of the Sausbee Formation in northeastern Oklahoma. From the middle part of the Johns Valley Shale, Gordon and Stone (1977) recorded the *Axinolobus modulus* goniatite zone. This zone occurs in the frontal Ouachitas in the lower part of the Wapanucka Limestone and in northeastern Oklahoma in the Chisum Quarry Member of the McCully Formation. Distinctive fossils have not yet been reported from the upper part of the Johns Valley.

MORROWAN AND "ATOKAN" SERIES

The "Atoka" Formation in the central Ouachitas consists of interbedded gray shale and fine-grained sandstone commonly having convolute bedding and sole markings. Shelburne (1960) recorded no more than 25-percent sandstone in the Boktukola syncline and a maximum preserved thickness in that area of 2,065 m. The top of the formation is eroded.

The "Atoka" Formation is virtually unfossiliferous in the central Ouachitas. However, L. R. Wilson (oral commun., 1978) reported that all the unit sampled contained palynomorphs of Morrowan age.

NORTHEASTERN ARBUCKLE MOUNTAINS

KINDERHOOKIAN AND OSAGEAN SERIES

The Woodford Shale in the northeastern Arbuckle Mountains is similar to that found in the southwestern Arbuckle Mountains, except that the percentage of chert decreases northeastward. In the northeast, the formation is composed predominantly of platy dark shale interbedded with some chert layers. The regional thickness is about 100–125 m (Ham, 1969). In the northeastern Arbuckle Mountains, the formation is apparently all Late Devonian in age at most localities, but a Kinderhookian conodont fauna has been recovered from the top 0.3 m or less at a few localities (Hass and Huddle, 1965).

The Sycamore Limestone, conspicuous in the southwestern Arbuckles, is absent from the Mill Creek syncline northeastward in the Arbuckle Mountains (Ham, 1969). The same stratigraphic position is locally occupied by the Welden Limestone. It is a buff to gray thick-bedded argillaceous limestone that is moderately fossiliferous. It has a maximum recorded thickness of 1.5 m, but it is typically 0.6 m thick (Barker, 1950). Cooper (1939) described a conodont fauna from this shale that contained 105 different species. This fauna was assigned a late Kinderhookian age by Ormiston and Lane (1976). Probably this pre-Welden shale rests unconformably on the underlying Woodford, and most of the Kinderhookian Series is missing in this area. Ormiston and Lane (1976) gave an early Osagean age to the Welden Limestone on the basis of conodonts.

Branson and Mehl (1941) described a conodont fauna from the basal part of the overlying Delaware Creek Shale (they termed it Caney). They apparently believed that, for the most part, those conodonts represented reworked specimens. Most of their specimens apparently came from the basal sandy, shaly glauconitic zone that, according to Barker (1950), directly overlies a major unconformity. Ormiston and Lane (1976) placed the Branson and Mehl conodont fauna in the lower Osagean. Ham (1969) stated that the regional unconformity at the base of the Delaware Creek Shale in the northeastern part of the Arbuckle Mountains explains the thinness of strata of possible Osagean and Meramecian age and, in part, the extreme thinness of the total Mississippian sequence in this area. Where the Welden Limestone is locally cut out below this unconformity, the basal sandy, glauconitic zone of the Delaware Creek Shale rests directly on the Woodford (Barker, 1950).

MERAMECIAN AND CHESTERIAN SERIES

The Delaware Creek Shale in the northeastern Arbuckle Mountains differs from that found in the southwestern Arbuckles by being friable and not siliceous. It contains the same local abundance of small phosphatic nodules and large calcareous septarian concretions. The formation is much thinner than that in the southwest, reaching a maximum thickness of only 49 m.

The lower 7.5 m of the Delaware Creek Shale is calcareous and fossiliferous locally on the Lawrence uplift (Barker, 1950). Elias (1956) named this occurrence the Ahloso Member of the Caney Shale.

Gordon and Stone (1977, fig. 4) recorded four goniatite zones for the Delaware Creek Shale in this area that ranged in age from late Meramecian to early Chesterian. The lowest calcareous part of the formation is apparently older than the lowest part of the formation in the southwestern Arbuckle Mountains, and it appears to be equivalent in age to the upper part of the Sycamore Limestone in that area.

The Goddard Shale in the northeastern Arbuckle Mountains is more or less similar to the Goddard in the southwestern Arbuckles. However, the percentage of sandstone decreases northeastward, and the thick sandstone members present in the Ardmore basin are lacking. There is also a marked decrease in its interval, which is only about 75 m in the northeastern Arbuckles. Gordon and Stone (1977, fig. 4) recorded three goniatite zones in this area that range in age from middle to late Chesterian.

MORROWAN SERIES

The Mississippian-Pennsylvanian boundary has not been defined in the northeastern Arbuckle Mountains. It falls somewhere within the poorly exposed shales, which are here included partly in the upper part of the Goddard Shale and partly in the lower part of the Morrowan, below the Union Valley Formation or the Wapanucka Limestone (fig. 4).

The Union Valley Formation crops out in a limited area on the Lawrence uplift, in the northernmost Arbuckle Mountains, and it covers an extensive area in the subsurface of the Arkoma basin, northeast of the Arbuckles. Where exposed on the Lawrence uplift, it consists of 45–80 m of sandstone overlain by 3.5–7.5 m of sandy limestone (Barker, 1950).

The limestone at the top of the Union Valley is highly fossiliferous locally. Gordon and Stone (1977, fig. 4) recorded the *Branneroceras branneri* goniatite zone from this interval. This limestone is of middle Morrowan age and is correlative with the Brewer Bend Limestone Member of the Sausbee Formation in northeastern Oklahoma and the upper Primrose Sandstone in the Ardmore basin.

On the northeast flank of the Arbuckle Mountains, the Union Valley Formation is missing on the outcrop, and the Wapanucka Limestone is underlain by 100 m or so of poorly exposed and poorly known shales.

The Wapanucka Limestone in this area includes a wide variety of carbonate-rock types, including crossbedded oolites and skeletal grainstones and packstones. In the more westerly exposures, significant shale interbeds are present. The maximum observed thickness for the Wapanucka in this area is about 55 m, not including underlying shales (Rowett and Sutherland, 1964). The Wapanucka is truncated westward along the south margin of the Franks graben. The Wapanucka is highly fossiliferous and is late Morrowan in age.

"ATOKAN" SERIES

Morrowan rocks are overlain unconformably in the northeastern Arbuckle Mountain area by the

"Atoka" Formation, which consists of thick shales interbedded with thinner fine-grained sandstones. The "Atoka" onlaps westward and is itself truncated farther west by an unconformity at the base of overlying Desmoinesian formations. It ranges in thickness in this area from about 910 m to a featheredge.

ARDMORE BASIN AND SOUTHWESTERN ARBUCKLE MOUNTAINS

KINDERHOOKIAN SERIES

The Woodford Shale consists of interbedded dark shale and chert and is 107–122 m thick over most of its outcrop area in the Arbuckle Mountains. It is Late Devonian in age except for the top 0.3 m or less at a few localities, from which a Kinderhookian conodont fauna was recovered by Hass and Huddle (1965).

OSAGEAN AND MERAMECIAN SERIES

The Sycamore Limestone overlies the Woodford Shale and occurs only in the southwest segments of the Arbuckle Mountains (Ham, 1969). It consists of fine-grained silty limestone interbedded with thin layers of dark-gray shale. It is 115 m thick on the south limb of the Arbuckle anticline and 64 m thick on the north limb (Fay, 1969). The Sycamore Limestone has produced very few identifiable megafossils, and age determinations thus far available are based on a few conodont faunas. These have been reported only from the lowest and the highest strata. Ormiston and Lane (1976) recovered conodonts from four samples from Fay's (1969, p. 68) measured section on the north limb of the Arbuckle anticline. Three of their samples came from the lowermost 17 m of the Sycamore Limestone (64 m total thickness) and indicate an early Osagean age for this part of the formation. The lowermost sample, of earliest Osagean age, was taken 1 m above the base of the formation; it carries the same fauna as that from the Welden Limestone in the northeastern Arbuckle Mountains (Ormiston and Lane, 1976).

The fourth sample described by Ormiston and Lane was from a zone 14 m below the top of the Sycamore Limestone, in the same measured section. They reported conodonts from this zone to be latest Meramecian or early Chesterian in age. The unfossiliferous middle part of the Sycamore Limestone is assumed to be Osagean and (or) Meramecian in age.

CHESTERIAN SERIES

For more than half a century the usage and age of the "Caney" Shale in southern Oklahoma has been controversial. The type locality of the "Caney" in the Ouachita Mountains is in fact a large erratic block in the Johns Valley Shale of Early Pennsylvanian age. The term Caney has now been generally abandoned in both the Arbuckle and Ouachita areas.

The name Delaware Creek was proposed by Elias (1956) as a member of the "Caney" Shale in the northeastern Arbuckle Mountains. This term was more recently used at formation rank throughout the Arbuckle Mountains (Hart, 1974, sheet 1). The Delaware Creek Shale rests directly on the Sycamore Limestone at the northern margin of the Ardmore basin. It consists in this area of dark-gray, partly siliceous shale that weathers to a lighter color than is normal for the formation in the northeastern Arbuckle Mountains. Small phosphatic nodules and large calcareous septarian concretions are locally abundant, but limestone and sandstone are absent (Ham, 1969). The formation ranges in outcrop thickness in this area from 69 m on Oil Creek to 134 m on Henryhouse Creek (Elias, 1956); it is 228 m in the subsurface of the Ardmore basin (Hart, 1974). Gordon and Stone (1977, fig. 4) recorded three goniatite zones in the unit in this area, all early Chesterian in age.

The "Springer" Group overlies the Delaware Creek Shale in the Ardmore basin. It includes both the Goddard Formation (Chesterian) and the Lake Ardmore Formation (Morrowan). The Goddard, named by Westheimer (1956), conformably overlies the Delaware Creek Shale at the north margin of the Ardmore basin. As defined by Elias (1956), the Goddard included only the beds of shale and thin sandstone between the top of the Delaware Creek and the base of the Rod Club Sandstone. Hart (1974, sheet 1) extended the Goddard Formation upward to the base of the Lake Ardmore Formation; in this usage, which is followed here, it includes the Rod Club and Overbrook Sandstones and interbedded shales. The total thickness of the Goddard Formation, as here used, amounts to 1,100 m in the Ardmore basin (Hart, 1974). Named sandstone members distributed upward within the formation include the Redoak Hollow, 8 m thick; the Rod Club, 75–122 m thick; and the Overbrook, 14–30 m thick (Tomlinson and McBee, 1962). These noncalcareous sandstones are all fine grained and well sorted. The lower Goddard shales differ from the Delaware Creek Shale in being more friable and less cliff forming and in containing abundant thin sideritic layers and concretions (Tomlinson and McBee, 1962).

Gordon and Stone (1977, fig. 4) recorded four goniatite zones in that part of the Goddard that lies below the Rod Club Sandstone Member, and these range in age from middle to early-late Chesterian.

MISSISSIPPIAN-PENNSYLVANIAN BOUNDARY

The Mississippian-Pennsylvanian boundary currently is placed arbitrarily at the base of the Lake Ardmore Formation in a sequence of apparently continuous deposition. This assignment is uncertain because few fossils are present in the interval from the base of the Rod Club Sandstone Member of the Goddard Formation to the base of the Primrose Sandstone Member of the overlying Golf Course Formation. Its position is most precisely limited by conodont faunas described by Straka (1972), who recorded conodont faunas of Chesterian age in the Goddard Formation as high as the Rod Club Sandstone. He recovered no diagnostic conodonts from the top of the Rod Club to the base of the Lake Ardmore Formation, but he described what he considered to be definite Pennsylvanian conodonts from the Target Limestone Lentil in the lower part of the Lake Ardmore Formation. He reported the occurrence of the same conodont fauna from the lowermost beds of the type Morrowan Series in northwestern Arkansas. Thus, the Mississippian-Pennsylvanian boundary apparently lies within the 130-m interval that includes the Overbrook Sandstone Member of the Goddard Formation and extends into the lower beds of the Lake Ardmore Formation.

MORROWAN SERIES

The Lake Ardmore Formation consists, in the northern part of the Ardmore basin, of 152 m of shales interbedded with three beds of ridge-forming fine-grained sandstone (Tomlinson and McBee, 1962). Each of these sandstone intervals is 9–21 m thick. The Target Limestone Lentil, referred to in the previous section, is about 15 m above the base; it is only about 1 m thick and is poorly exposed along a distance of 3 km or so in the northern part of the Ardmore basin.

The Golf Course Formation includes the members listed as follows, in ascending order: Primrose Sandstone, Gene Autry Shale (in the north only), Jolliff Limestone (in the south only), and Otterville Limestone. The Golf Course attains a thickness of about 610 m (Tomlinson and McBee, 1962).

The Primrose Sandstone Member is 46–76 m thick and differs from the underlying beds of sandstone of the "Springer" Group in being distinctly calcareous in several zones. Two goniatite occurrences indicate an early to middle Morrowan age, the upper being the *Branneroceras branneri* zone (Gordon and Stone, 1977, fig. 4).

North of Ardmore, the Primrose is overlain by the Gene Autry Shale Member, which is about 360 m thick. The Gene Autry contains the late Morrowan *Axinolobus modulus* goniatite fauna in the lower part.

South of Ardmore, a thin shale overlies the Primrose, which is in turn overlain by the distinctive Jolliff Limestone Member. The Jolliff is highly variable in character and thickness. A basal limestone-cobble conglomerate as much as 9 m thick occurs typically, as do irregularly distributed higher conglomerates, packstones, and carbonate mudstones. The formation varies in thickness from 6 to 39 m by lateral replacement of the lower part of the overlying shale (Cromwell, 1975).

The unnamed shale between the Jolliff and Otterville Limestone Members is as much as 137 m thick. The Otterville is either covered or missing in the area west of Ardmore, and the covered (and apparent shale) interval between the Jolliff and the base of the Bostwick Member of the Lake Ardmore Formation is as thick as 360 m (Cromwell, 1975).

The Otterville ranges from 2.5 to 6 m in thickness and consists of fossiliferous oolitic grainstones. It is overlain by an unnamed shale as much as 91 m thick that extends to the base of the Bostwick.

The Jolliff Limestone Member contains the *Axinolobus modulus* goniatite fauna. The interval from the Jolliff to the shale above the Otterville is late Morrowan in age and correlates with the whole of the McCully Formation in northeastern Oklahoma.

"ATOKAN" SERIES

This series in the Ardmore basin includes that part of the Lake Murray Formation that extends from the base of the Bostwick Member to the base of the Lester Limestone. The Bostwick Member, composed of beds of conglomerate, sandstone, limestone, and intercalated shale, has a maximum thickness of 152 m (Cromwell, 1975). Limestone-cobble conglomerate in the south grades into chert-pebble conglomerate in the central part of the Ardmore basin. This part of the sequence is characterized by marked lateral facies changes. The Bostwick is the most conspicuous ridge-forming unit in the southern Ardmore basin.

The "Atokan" age assignment of the Bostwick is based primarily on fusulinids. Waddell (1966) defined his Fusulinid Zone I as including the Bostwick

Member and part of the overlying shale. This zone is characterized by the occurrence of several species of *Fusulinella*.

Overlying the Bostwick, an unnamed shale, which has a maximum thickness of 228 m, was included in the "Atokan" Series by Waddell (1966). He placed the base of the Desmoinesian Series at the base of the Lester Limestone Member of the Lake Murray Formation, on the basis of the lowest occurrence in the Ardmore basin of the genera *Fusulina* and *Wedekindellina*.

MIDDLE AND UPPER PENNSYLVANIAN STRATIGRAPHY

By S. A. Friedman

As defined in this report, the Middle Pennsylvanian consists of the Desmoinesian and Missourian Series, and the Upper Pennsylvanian consists of the Virgilian and Gearyan(?) Series. As explained previously, Gearyan rocks are now tentatively referred to the uppermost Pennsylvanian by the Oklahoma Geological Survey; traditionally, most workers have placed them in the Lower Permian.

Strata of Desmoinesian age crop out in the Arkoma basin, the northern Oklahoma shelf, and the Ardmore basin. Strata of Missourian and Virgilian age are exposed mainly in the shelf area and in the Ardmore basin but are absent because of erosion in the Arkoma basin. Rocks of Gearyan age are present in the western part of the northern shelf area. Differences in lithology, thickness, structure, and depositional environments help in distinguishing these strata in the shelf and basin areas. Figure 6 shows the general stratigraphic relationships of these Middle and Upper Pennsylvanian rocks.

Earlier workers believed that widespread unconformities are present within the Pennsylvanian of Oklahoma. During the last 10 years, detailed stratigraphic mapping has shown that local disconformable surfaces exist at the bases of channels filled with sandstone and chert conglomerate and that progressive overlap and offlap northward and northwestward are responsible for the erroneous interpretation of regional unconformities within these units.

Some units of the Desmoinesian Series contain many thick beds of sandstone in the Arkoma basin that do not persist northward into the northern Oklahoma shelf. Beds of Desmoinesian, Missourian, and Virgilian limestone are more numerous and thicker north of Tulsa in the shelf area than they are south of Tulsa and farther south in the Arkoma basin area. Some of the upper Desmoinesian sandstones and shales are thick in the western part of the Arkoma basin but thin northward, where they are replaced by limestones and shales in the shelf area. The entire Desmoinesian sequence is only one-seventh as thick on the shelf as it is in the Arkoma basin.

Generally, rocks of the Desmoinesian Series have been assigned to three groups, in ascending order, the Krebs, Cabaniss, and Marmaton (fig. 6). Rocks of the Missourian Series have been assigned to two groups, in ascending order, the Skiatook and Ochelata (fig. 6).

ARKOMA BASIN

DESMOINESIAN SERIES

In any one region of the Arkoma basin, the maximum thickness of Desmoinesian shale, sandstone, and relatively thin beds of coal and underclay is 2,000 m. The series is thickest in synclines just north of the Choctaw fault, where the basal Hartshorne Formation attains a maximum thickness of 910 m of sandstone, shale, and a major coal bed. Geologists believe that the Hartshorne was deposited within a deltaic system whose source was to the east and that most of the McAlester and Boggy Formations were deposited within deltaic systems that originated to the north, on the northern Oklahoma shelf. The Thurman Sandstone and the Stuart Shale apparently had southern sources, and they overlap underlying strata in the northern shelf area. At least two deltaic systems in the Senora Formation have eastern sources.

The Hartshorne Formation conformably overlies the "Atoka" Formation ("Atokan" Series) and is only 7.5 m thick at the northern part of the basin. In places at the southwestern end of the basin, the Hartshorne is absent because of pre-Desmoinesian erosion. Plant compressions (mostly ferns) and casts and molds of upright tree trunks are found in the shale and sandstone members, and roots and rootlets, in the underclay of the Hartshorne Formation. The Hartshorne coal is 1.5–3 m thick at the western end and 2 m thick at the southeastern corner of this basin. The Hartshorne coal probably is absent north of Muskogee in the shelf area.

The McAlester Formation conformably overlies the Hartshorne Formation and has a maximum thickness of 760 m in the southeastern part of the basin; it is only 60 m thick in the northern part. This formation consists of many beds of medium- to dark-gray silty or clayey shale, gray sandstone,

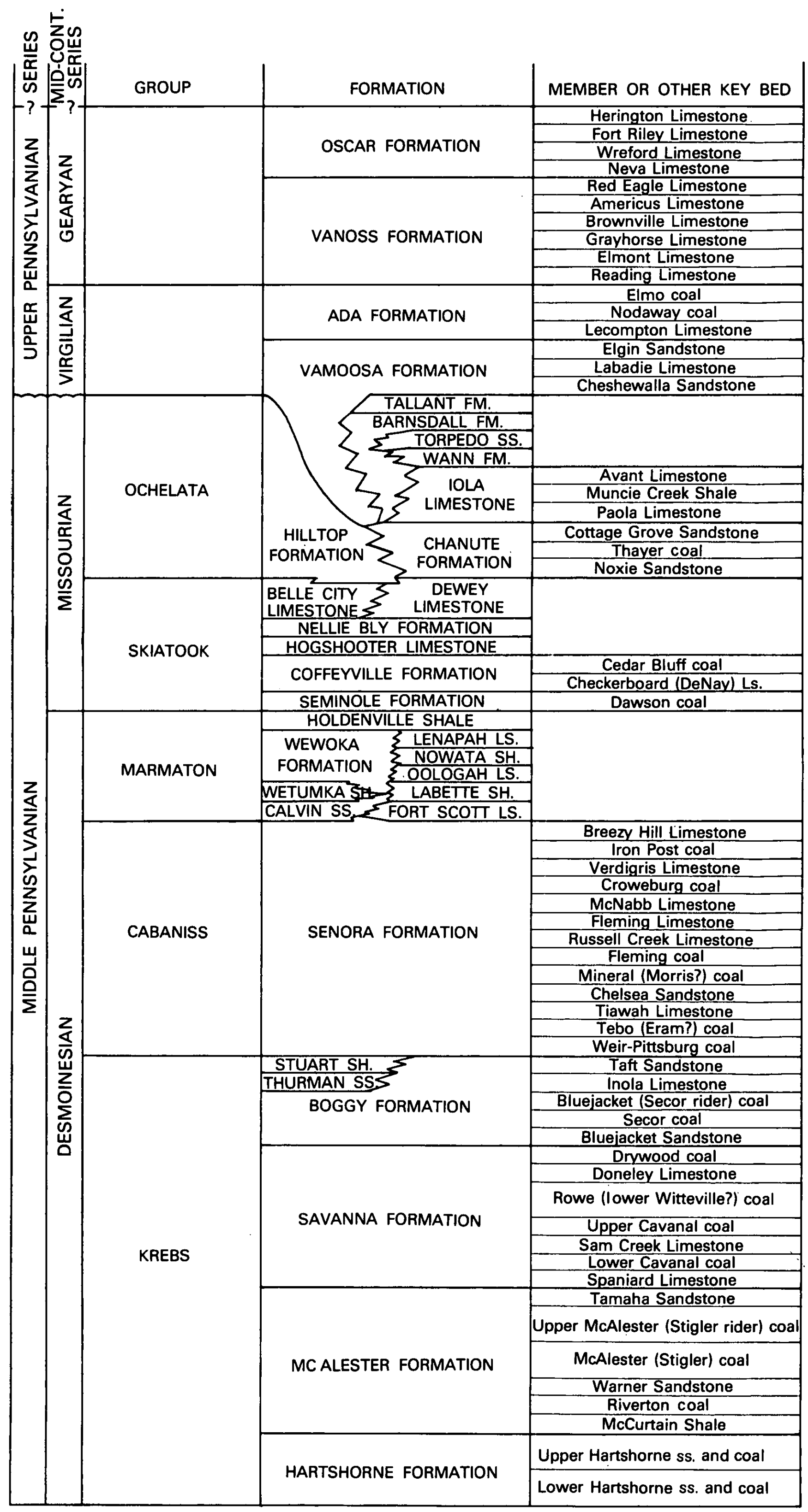

FIGURE 6.—Generalized stratigraphic chart of Middle and Upper Pennsylvanian rocks in Oklahoma.

coal, and underclay in cyclic sequences. The McAlester (Stigler) coal occurs in the upper third of the formation. The coal is as much as 1.5 m thick in the western part of the basin, 1 m thick at McAlester in the central part, and about 0.5 m thick in the eastern part. The coal thins in the northern part of the basin and is absent on the shelf area north of Muskogee.

The Savanna Formation conformably overlies the McAlester Formation, except at places where local disconformities were formed by channel sandstones. The Savanna consists mainly of shales and sandstones and attains a maximum thickness of about 600 m in the southeastern and southwestern parts of the Arkoma basin. Along the northern part of the basin, this formation is only 76 m thick. Thin coal and underclay beds are present in the Savanna Formation. Upright tree trunks have been preserved in some of these beds. In the basin, the lower and upper Cavanal, Rowe, and lower Witteville coals are important coal beds of the Savanna.

The Boggy Formation consists mainly of sandstones and shales, and it conformably overlies the Savanna Formation except at places where the basal Bluejacket Sandstone Member occupies erosional channels. The Boggy reaches a maximum thickness of 900 m where fully preserved in the McAlester area, and it is at least 600 m thick at each end of the Arkoma basin. It contains one major coal (the Secor) and underclay bed in the basin. The Boggy is only 60 m thick in the northern part of the basin.

The overlying Thurman Sandstone underlies the Stuart Shale, at the top of the Krebs Group, in the western part of the basin. Both formations contain chert-conglomerate, sandstone, and shale beds, but the Thurman contains a greater quantity of sandstone. The maximum thicknesses of each formation are found in the west-central part of the basin, where the Thurman is 76 m thick and the Stuart, 115 m. The Stuart overlaps the Thurman northward, where it overlies the Boggy Formation. Both the Stuart and the Thurman are confined to the Arkoma basin.

The overlying Senora Formation in the northwestern part of the Arkoma basin consists of 45–275 m of sandstones and shales.

The overlying Calvin Sandstone, Wetumka Shale, and Wewoka Formation are composed of sandstones and shales that are mostly restricted to the basin area and thin northward. The Calvin is composed mostly of sandstone and shale that have a maximum thickness of about 120 m. The Wetumka conformably overlies and overlaps the Calvin northward and is composed mostly of shales and some sandstones and siltstones. Its maximum thickness is about 60 m. The Wewoka conformably overlies and overlaps the Wetumka and the Calvin northward and attains a maximum thickness of about 215 m; it consists of limestone, chert conglomerate, sandstone, and calcareous shale. Northward, the Wewoka clastic rocks are fine grained, and the formation thins. The Holdenville Shale is present in a small area in the northwesternmost part of the Arkoma basin, where it overlies the Wewoka Formation conformably and attains a thickness of 79 m.

NORTHERN OKLAHOMA SHELF

Undoubtedly an unconformity is present at the base of the Desmoinesian Series in places that had been exposed to pre-Desmoinesian erosion in the northern Oklahoma shelf and Arkoma basin areas; it is progressively younger northward. Nondeposition and erosion resulted in progressive northward and northwestward thinning or absence of strata of Desmoinesian through Gearyan age that crop out in the shelf area.

DESMOINESIAN SERIES

The Hartshorne Formation is 15–30 m thick in the southernmost part of the northern shelf or homocline area. It is less than 15 m thick in T. 14 N., south of Muskogee, and is probably absent at, and north of, Tulsa.

The McAlester Formation conformably overlies the Hartshorne at most places but overlaps it northward in the shelf area. The contact between the Hartshorne and McAlester has been controversial, but recent workers have agreed to use the top of the Hartshorne coal or the upper Hartshorne coal as the boundary (McDaniel, 1961; Oakes and Knechtel, 1948; American Association of Petroleum Geologists and others, 1978).

At the southern border of the shelf area, in T. 10 N., Muskogee County, the McAlester Formation is 137 m thick (Oakes, 1977), and in T. 15 N., at Muskogee, it is only 55 m thick (Oakes, 1977). Northward, the basal McCurtain Shale Member probably is absent. The Riverton coal, miscorrelated with the upper Hartshorne coal by Branson (1962, p. 441), is probably a thin coal underlying the Warner Sandstone Member (Branson and others, 1965). The principal coal in the McAlester is the Stigler, which is not present north of Muskogee. In Ottawa County, only the Warner Sandstone Member is recognized, and the McAlester Formation is possibly only about 18 m thick at the Kansas border.

The overlying Savanna Formation consists mostly of 30–140 m of shale and sandstone in the northern shelf area. The principal key beds of the Savanna are the Spaniard Limestone, Sam Creek Limestone, Rowe coal, Doneley Limestone, and Drywood coal. The Savanna is only 30 m thick at the Kansas border but is 140 m thick in southern Muskogee County.

The Boggy Formation overlies the Savanna Formation conformably, although channel-fill sandstone in the basal Bluejacket Sandstone Member indicates a local disconformity. The Boggy is about 210 m thick in Muskogee County, in the southern part of the shelf area. Northward in Craig County, the Boggy is only about 44 m thick. Key beds of the Boggy, besides the basal Bluejacket Sandstone, are, in ascending order, the Secor coal, Bluejacket coal, Inola Limestone, and Taft Sandstone. The Secor coal has not been identified north of Wagoner County, and the Bluejacket coal has been recognized in Craig County. These coals are possibly correlative, as each overlies the Bluejacket Sandstone Member.

Apparently the Stuart Shale overlies the Boggy Formation conformably in the southern parts of the shelf, but it is not present north of T. 13 N.

At most other places in the northern Oklahoma shelf area, the Senora Formation conformably overlies the Boggy. The Senora consists of approximately 46–305 m of sandstone, shale, coal, underclay, and limestone in conspicuously cyclical sequences. In the southern part of the shelf area, key beds are the Eram, Morris, and Croweburg coals. In the central part, key beds are the Chelsea Sandstone; the Mineral, Croweburg, and Iron Post coals; and the McNabb and Verdigris Limestones. In the northern part, key beds are the Weir-Pittsburg, Tebo, Mineral, Fleming, Croweburg, and Iron Post coals; and the Tiawah, Russell Creek, McNabb, Verdigris, and Breezy Hill Limestones. The limestones are light to dark gray, fine grained, massive, and dense; they contain marine invertebrates and algae. The coal beds are thin, but they do attain thicknesses of 0.25–1 m, where they constitute recoverable coal reserves and have been mined. Some of the limestone beds are as much as 9 m thick in the northern part of the shelf area.

The northward change of lithofacies to beds of fine-grained clastic rocks and limestone and the northward thinning in the Marmaton Group are no less dramatic than in the underlying Cabaniss Group (Senora Formation). But the Marmaton is thinner, has fewer named units, and is easier to map. It is about 210 m thick at the Kansas border and about 245 m thick at the Arkansas River.

South of the Arkansas River, in the shelf area, the Marmaton Group comprises, in ascending order, the Calvin Sandstone, Wetumka Shale, Wewoka Formation, Lenapah Limestone, and Holdenville Shale. North of the Arkansas River, the Fort Scott Limestone was reported to underlie the Calvin Sandstone (Oakes, 1952). However, some earlier workers confused the Breezy Hill with the lower member of the Fort Scott. Thus, the Breezy Hill Limestone possibly extends as far south as Tulsa, Wagoner, and Muskogee Counties.

The Labette Shale of the shelf is probably equivalent in part to the Calvin Sandstone and Wetumka Shale. The Oologah Limestone conformably overlies the Labette Shale and is in turn overlain by the Nowata Shale. The Lenapah Limestone conformably overlies the Nowata. The Wewoka Formation consists of beds of sandstone and calcareous shale, believed to be lateral equivalents of the Labette Shale, Oologah Limestone, Nowata Shale, and Lenapah Limestone. The Holdenville Shale conformably overlies the Wewoka and the Lenapah and is the top formation of the Marmaton Group. Key units in this group are shown in figure 6. The Calvin, Wetumka, Wewoka, Nowata, and Holdenville thin northward, and the Fort Scott, Labette, Oologah, and Lenapah thicken northward.

MISSOURIAN SERIES

The Seminole Formation, the basal formation of the Skiatook Group, probably overlies the Holdenville Shale conformably. In places where chert conglomerate and sandstone fill channels at its base, the Seminole is in local disconformable contact with the Holdenville, Lenapah, and Nowata Formations. The local disconformities led earlier writers to select this contact as the Desmoinesian-Missourian boundary because they believed it was a regional unconformity (Oakes, 1952, p. 48–54). The Seminole is 36 m thick in the southern part of the shelf area and about 6 m thick at the Kansas border. It contains a few thin beds of coal in addition to the Dawson coal, a commercial seam in Tulsa and Rogers Counties.

The Coffeyville Formation conformably overlies the Seminole Formation. At its base is the dark-gray finely crystalline fossiliferous Checkerboard Limestone Member. The Checkerboard is probably the equivalent of the DeNay Limestone of the southern part of the northern shelf area; it is present from Okmulgee County northward into Kansas. The

Coffeyville consists mainly of 45–140 m of shale, sandstone, and chert conglomerate. The thickest part of this formation is in Tulsa County and vicinity. It is present in the central and northern part of the northern shelf-homocline area and is equivalent to the Francis Formation (restricted) in the southern part of the shelf area (Hart, 1974, sheet 1), where it is only 6–21 m thick. The thin Cedar Bluff coal is present below a sandstone at the top of the formation in Washington County (Oakes, 1940, p. 38).

The Hogshooter Limestone conformably overlies the Coffeyville Formation. Locally, the Hogshooter attains a thickness of 15 m, but at most places the formation is only 1.5 m thick. It contains invertebrate fossils and phosphatic nodules. Generally, the Hogshooter is thicker in the northern part of the shelf area, in Tulsa and Washington Counties, than in the southern part, where it is absent locally.

The Nellie Bly Formation, 25–167 m thick, conformably overlies the Hogshooter Limestone. The Nellie Bly consists of shale, sandstone, and, in the southern part of the shelf, chert and limestone conglomerate. The formation thins northward and contains thin limestone beds at the top.

The Dewey Limestone forms the top unit of the Skiatook Group. It conformably overlies the Nellie Bly and grades southward into the Belle City Limestone and the base of the Hilltop Formation of the overlying Ochelata Group. The Dewey consists of limestone and shale 6–18 m thick near the Kansas border and calcareous sandstone and sandy limestone in the southern part of the shelf area. The Belle City Limestone, 3–6 m thick, is found only in the southern part of the shelf area.

The Chanute Formation, 4–58 m thick, is the basal unit of the Ochelata Group and contains the members listed as follows, in ascending order: Noxie Sandstone, Thayer coal, and Cottage Grove Sandstone. In the northern part of the shelf area, it also contains a basal limestone conglomerate and a gray shale. At some places, the Chanute (probably in channels) disconformably overlies the Dewey Limestone, the Nellie Bly Formation, and the Hogshooter. Apparently the Chanute grades laterally into the Hilltop Formation between Interstate Highway 40 and the Arbuckle Mountains (Hart, 1974, sheet 1; Bingham and Moore, 1975, sheet 1).

The overlying Iola Limestone, 1–30 m thick, contains three members, the Paola Limestone, Muncie Creek Shale, and Avant Limestone, in ascending order. The Iola conformably overlies the Chanute in the northern part of the shelf area. Southward, the Iola apparently grades laterally into the lower part of the Barnsdall Formation.

The Wann Formation, 15–122 m thick, conformably overlies the Iola Limestone and consists of units of shale, sandstone, and thin fossiliferous limestone. The Wann thins and apparently grades laterally southward into the Barnsdall Formation. The Wann is overlain conformably by the Torpedo Sandstone, which consists of 0–18 m of sandstone and 0–29 m of shale. The Torpedo apparently thins and grades laterally southward into the Barnsdall. The Barnsdall crops out in the central and northern part of the shelf area, where it is 14–61 m thick; it grades into the Chanute Formation, Iola Limestone, and Wann Formation. South of Interstate Highway 40, the Hilltop Formation, which is about 21 m thick, occurs at about the same position as the Barnsdall, and the two formations intergrade.

The Tallant Formation conformably overlies the Barnsdall Formation and consists of 23–61 m of gray and brown shales and sandstones. It thins southward and is cut out south of T. 13 N. by channels in the overlying Vamoosa Formation, the basal unit of the Virgilian Series.

VIRGILIAN SERIES

The outcropping part of the Upper Pennsylvanian Virgilian Series consists of the Vamoosa Formation, 76–210 m thick, overlain by the Ada Formation, 30–122 m thick.

The Vamoosa Formation consists of brown and gray shales and light-gray limestones in the northern part of the northern Oklahoma shelf and light-brown sandstones and chert conglomerates that fill channels in the southern part of the shelf. A maximum thickness of 210 m of the formation crops out in Okfuskee County, but the Vamoosa is not as thick in Osage and Creek Counties. The formation overlaps the entire Ochelata Group and, thus, probably overlies the Missourian Series unconformably (Ries, 1954, p. 81). However, a basal shale of the Vamoosa overlies the Tallant Formation conformably at some places.

The Ada Formation consists of beds of sandstone, limestone, chert and limestone conglomerates, and brown shale. The Ada conformably overlies the Vamoosa except near Ada, where a large channel cuts out the Vamoosa. In parts of the shelf area to the south, the Ada is the correlative of the Collings Ranch Conglomerate (Hart, 1974, sheet 1).

GEARYAN SERIES

The Gearyan Series is included tentatively at the top of the Pennsylvanian System in Oklahoma, because its flora and fauna are similar to those of other Upper Pennsylvanian strata and it conformably overlies the strata of the Virgilian Series. The Gearyan comprises the Vanoss Formation and the Oscar Formation, which conformably overlies the Vanoss.

The Vanoss Formation consists of beds of shale, red-brown arkosic sandstone, limestone, limestone and chert conglomerates, and coal. It is 76–183 m thick. Thin units of limestone and of shale are common north of the Canadian River in the northern part of the shelf area, and units of conglomerate and sandstone, in the southern part. However, the formation is thickest in the northern part. Many of the lithologic units in this formation are red.

The Oscar Formation conformably overlies the Vanoss and consists of limestone, shale, arkosic sandstone, and conglomerate beds. Its thickness increases from 91 m in the southern part of the shelf to 213 m in the northern part. The Oscar arkosic sandstones and conglomerates are present near the Arbuckle Mountains. As currently recognized by the Oklahoma Geological Survey, the Herington Limestone Member is the highest stratigraphic unit of the Oscar Formation and, therefore, of the Pennsylvanian System in Oklahoma. It is conformably overlain by Permian strata.

ARDMORE BASIN

Only the Desmoinesian and Missourian Series are represented in the Ardmore basin, the former by the Deese Group and the latter by the Hoxbar Group (see Hart, 1974, sheet 1). The stratigraphy of these rock groups in this basin is complex, little described, and little understood. Lithologies grade in short distances from limestone and chert conglomerates to sandstones and shales containing many thin beds of limestone.

The Deese Group attains a total thickness of about 2,950 m and is considered equivalent to the lower Franks Conglomerate and to the lower part of the upper Franks Conglomerate (Hart, 1974, sheet 1). The Hoxbar Group is about 850 m thick and is probably equivalent to the upper part of the upper Franks Conglomerate (Hart, 1974, sheet 1).

IGNEOUS AND METAMORPHIC ROCKS

By Kenneth S. Johnson

Igneous and metamorphic rocks of Carboniferous age are sparse in Oklahoma. Veins, as open-space fracture fillings, consist mainly of quartz and contain lesser amounts of barite and calcite; some of the veins contain lead, zinc, and copper sulfides. The veins are chiefly in the Ouachita Mountains, but some also are present in the Wichita, Arbuckle, and Ozark uplifts. Introduction of the vein material probably took place during late stages of deformation, late in the Pennsylvanian Period; Miser (1959) showed a similarity between the abundance of quartz veins and degree of metamorphism in the Ouachita Mountains. Some workers believe that the zinc and lead deposits of the Tri-State district in northeastern Oklahoma also may have been emplaced in Middle or Late Pennsylvanian time.

Volcanic ash and pyroclastic flows are present in the Stanley Shale (Upper Mississippian) of the Ouachita Mountains. Tuff beds that range in thickness from 7 to 40 m were deposited in the deep-marine basin as interbeds with flysch sediments. Whole-rock samples analyzed for rubidium and strontium isotopes yielded an isochron age of 310 ± 15 m.y. (Mose, 1969), and associated conodonts show a late Mississippian age.

Low-grade metasedimentary rocks in the core area of the Ouachita Mountains were apparently metamorphosed mainly during the Pennsylvanian orogeny. Shales, sandstones, and impure limestones of Ordovician age were dynamically metamorphosed under considerable shearing stress at relatively low temperatures into slate, phyllite, schist, metasandstone, and graphitic marble (Goldstein, 1975). Age determinations from micas range from Devonian to Early Permian; the younger Pennsylvanian and Early Permian ages probably represent the time of deformation in the Ouachita Mountains.

ECONOMIC RESOURCES

COAL

By S. A. Friedman

Coal-bearing strata of Middle and Late Pennsylvanian age in eastern Oklahoma occur in an area of approximately 36,000 km² in the southern part of the western region of the interior coal province of the United States (Campbell, 1917) (fig. 7).

An area of approximately 20,000 km² contains at least 24 coal beds, 20 of which have been formally named. These beds are 0.25–2 m thick, and they are of commercial value, or have been during the past 104 years. Figure 6 shows the positions of the coal beds within the Middle and Upper Pennsylvanian stratigraphic framework. The identified bituminous-coal resources are shown by county in figure 7. Low- and medium-volatile bituminous coals occur in the eastern part of the Arkoma basin, and

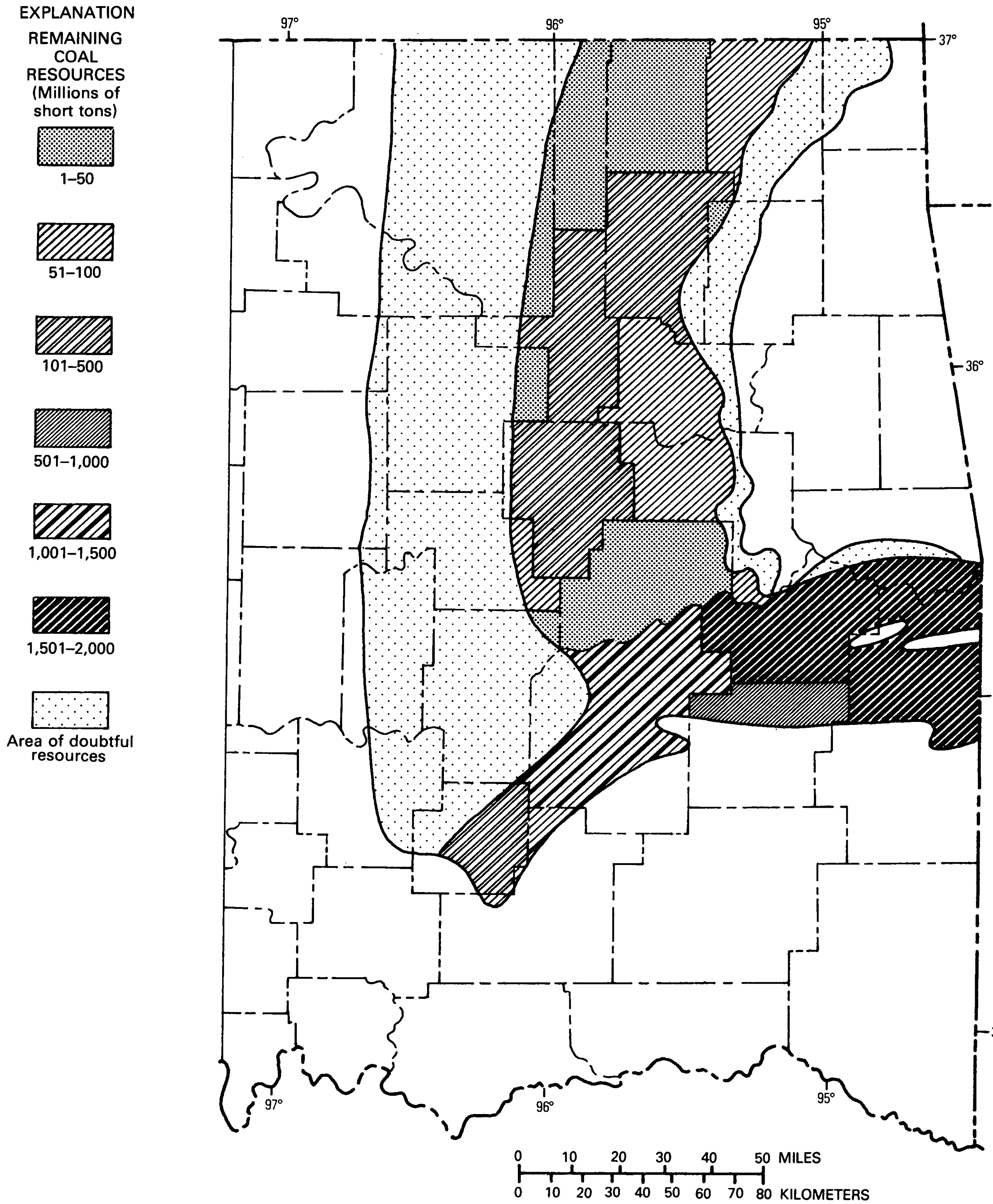

FIGURE 7.—Distribution of remaining coal resources in Oklahoma part of western region of interior coal province.

high-volatile bituminous coals are present in the western part of the basin and in the northern Oklahoma shelf area.

Most of the coals are moderately dull to moderately bright banded, and the lithotypes are thin banded. Vitrain is sparse, and bright attritus is abundant. The thickest vitrain (5 mm) occurs in the Croweburg coal, and the brightest attritus, in the Stigler coal.

The most recent estimate by the Oklahoma Geological Survey indicates that 7,700 million short tons of coal constitutes the original identified resources; 500 million tons has been mined and lost in mining, leaving 7,200 million tons as remaining resources, of which only 2,300 million tons is considered to be net recoverable reserves (Friedman, 1974). Most of the resources are contained in the Hartshorne coals, in the base of the Desmoinesian Krebs Group (fig. 6). The Hartshorne, McAlester, Stigler, and Croweburg coals contain most of the low-sulfur (1-percent sulfur or less) coal and all the coking coal in the State. Other coals, in isolated occurrences, are low in sulfur, and many coals show a free swelling index of 5–9 but are high in sulfur and ash. Most Oklahoma coals yield low ash-fusion temperatures (800° C–1,160° C).

Desmoinesian coal beds are found in the Arkoma basin in a clastic sequence 1,800 m thick and in the northern Oklahoma shelf area in a sequence 300 m thick. In the Arkoma basin, coal-bearing strata crop out on the flanks of plunging, broadly folded synclines and of some sharply folded and faulted anticlines. Folds commonly plunge only 1°–3°, but coal beds on the flanks of the folds dip 3°–65°. In the northern Oklahoma shelf area, the strata show a west-northwestward regional dip of ½°–2° but also show superimposed folds and structures associated with channel-fill sandstone that cause beds to attain dips of approximately 5°.

The presence of coal in Oklahoma had been noted as early as 1821, according to Trumbull (1957, p. 361), but mining on a commercial scale did not begin until the Missouri-Kansas-Texas Railroad was built through McAlester in 1872. Coal production in Oklahoma increased gradually until 1900. Then it increased sharply, although in spurts, culminating in 4.85 million tons of coal in 1920 for a record high that was not surpassed for 57 years. Most of the coal production during these times was from underground mines in the Arkoma basin, including the Henryetta district.

Production decreased sharply during the economic depression of the 1930's, and it increased during the wartime economy of the 1940's, peaking at 3.2 million tons in 1946, when the Henryetta district led with 37 percent of the total. Thirty years later (1976), that figure was exceeded when coal production reached 3.6 million tons. In 1977, a new record high was set at 5.3 million tons.

Strip mining has been the principal method of coal mining in Oklahoma since 1942 and has been the exclusive method of mining since 1972, except for one tandem auger mine. Several companies have indicated plans for underground coal production in the Arkoma basin by 1980.

Approximately 35 strip mines produced coal from 12 beds in Oklahoma in 1977. During 1976, 28 strip mines produced coal from 12 beds; 48 percent of this production was from the Iron Post coal (table 1). From 18 to 32 percent of the coal produced from 1972 through 1976 was used for coke manufacture in Texas, Colorado, Ohio, Pennsylvania, and Japan. Steam electricity-generating plants in Kansas and Missouri accounted for the increased coal use in 1972–77; thus, the percentage of coking-coal production in Oklahoma (table 2) decreased.

Less than 1 percent of the State's 1976 production was from coal beds dipping more than 18°. Most (91 percent) of the production was from beds dipping less than 5°, and 8 percent was from beds dipping 5°–18°. The thickness of coal beds mined in 1976 was 0.25–1.8 m, and the maximum thickness of overburden at strip mines was 36 m.

TABLE 1.—*Oklahoma coal production, by bed, 1976*

Coal bed	Short tons	Percentage of total
Iron Post	1,719,596	48
Croweburg	665,899	19
Mineral	450,441	13
Stigler	239,601	7
Upper Hartshorne and Lower Hartshorne (combined)	228,159	6
Secor and Secor rider	95,843	3
Weir-Pittsburg	58,233	2
Rowe	27,790	1
Cavanal	19,447	1
McAlester	12,541	<1
Totals	3,517,550	100

TABLE 2.—*Estimated uses of Oklahoma coal, 1972–76*

Year	Coking coal production		Steam coal and other coal production
	Short tons	Percentage of total	Short tons
1972	680,000	27	1,850,211
1973	606,828	28	1,587,842
1974	582,455	25	1,792,230
1975	910,004	32	1,940,423
1976	665,924	18	2,960,757

Most Oklahoma coal mines produced less than 100,000 short tons in 1976. Only Peabody Coal Co.'s Rogers County No. 2 mine produced more than 1 million tons (in 1976 and in 1977), and this coal (the Iron Post) was consumed in steam electricity-generating plants out of State. Garland Coal and Mining Co. ranked second in production but produced less than 50,000 tons, all of which was used for coke manufacture out of State.

In 1976, Oklahoma ranked 20th in coal resources and production in the United States. The reported cumulative coal production from Oklahoma from 1873 through 1977 is 212 million short tons.

PETROLEUM

By John F. Roberts

In Oklahoma, the oil produced from Carboniferous rocks is estimated to constitute 60 percent of total cumulative production in the State, as well as 60 percent of current production and remaining recoverable reserves. Natural gas produced from the Carboniferous is estimated to constitute 69 percent of total cumulative production, current production, and remaining recoverable reserves in the State. These estimates are based on those furnished by the American Petroleum Institute and the American Gas Association.

The recorded cumulative production of crude oil in Oklahoma to January 1, 1977, is 11,487,173,000 barrels (42 U.S. gallons per barrel). The value of this oil is estimated to be $27,015,173,000.

Oil production during 1976 was 150,627,000 barrels, valued at $1,432,463,000. Peak production was 277,775,000 barrels, valued at $397,200,000, during 1927. Annual production and value through 1976 are shown in figure 8.

Natural-gas production in Oklahoma from 1906 through 1976 totaled 40,030,023 million ft^3, and its value was $6,289,820,000. Natural-gas production during 1976 was 1,710,586 million ft^3, and its value was $858,714,000. Peak natural-gas production was in 1972 (1,806,887 million cubic feet, valued at $294,523,000). These statistics are shown in figure 9.

During 1976, Oklahoma's total petroleum products amounted to 94 percent of the State's total mineral production and had a value of $2.5 billion. This value includes $208 million from natural gasoline, hydrocarbons that have been recycled by repressuring of reservoirs, and liquefied petroleum gases not included in crude-oil production and values given in figure 8.

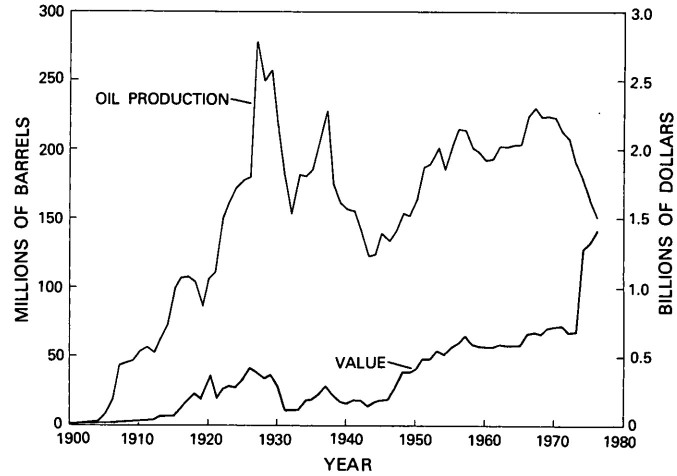

Figure 8.—Crude-oil production and value in Oklahoma, 1891–1976.

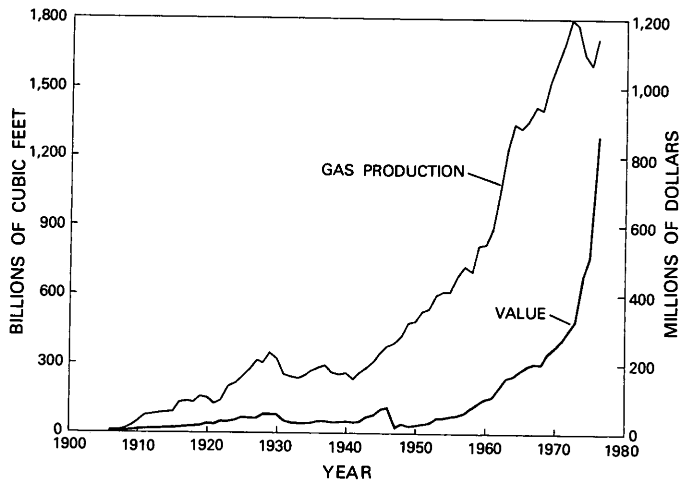

Figure 9.—Natural-gas production and value in Oklahoma, 1906–76.

Oklahoma liquid-petroleum production accounted for 5.4 percent of the United States total for 1976; natural-gas production accounted for 8.5 percent.

Since 1889, when Oklahoma's first producing well was drilled, more than 300,000 wells have been drilled in the State. Oil and (or) gas has been produced from more than 218,000 of these wells. In 1976, 12 crude-oil refineries, 83 natural-gas-processing plants, and 8 petrochemical plants were operating in Oklahoma; at least 43,200 persons were directly engaged in these operations. Thus, the petroleum industry continues to have a significant effect on the economy of Oklahoma. Figure 10 shows the extent of the State's producing areas as well as the location of giant oil and gas fields.

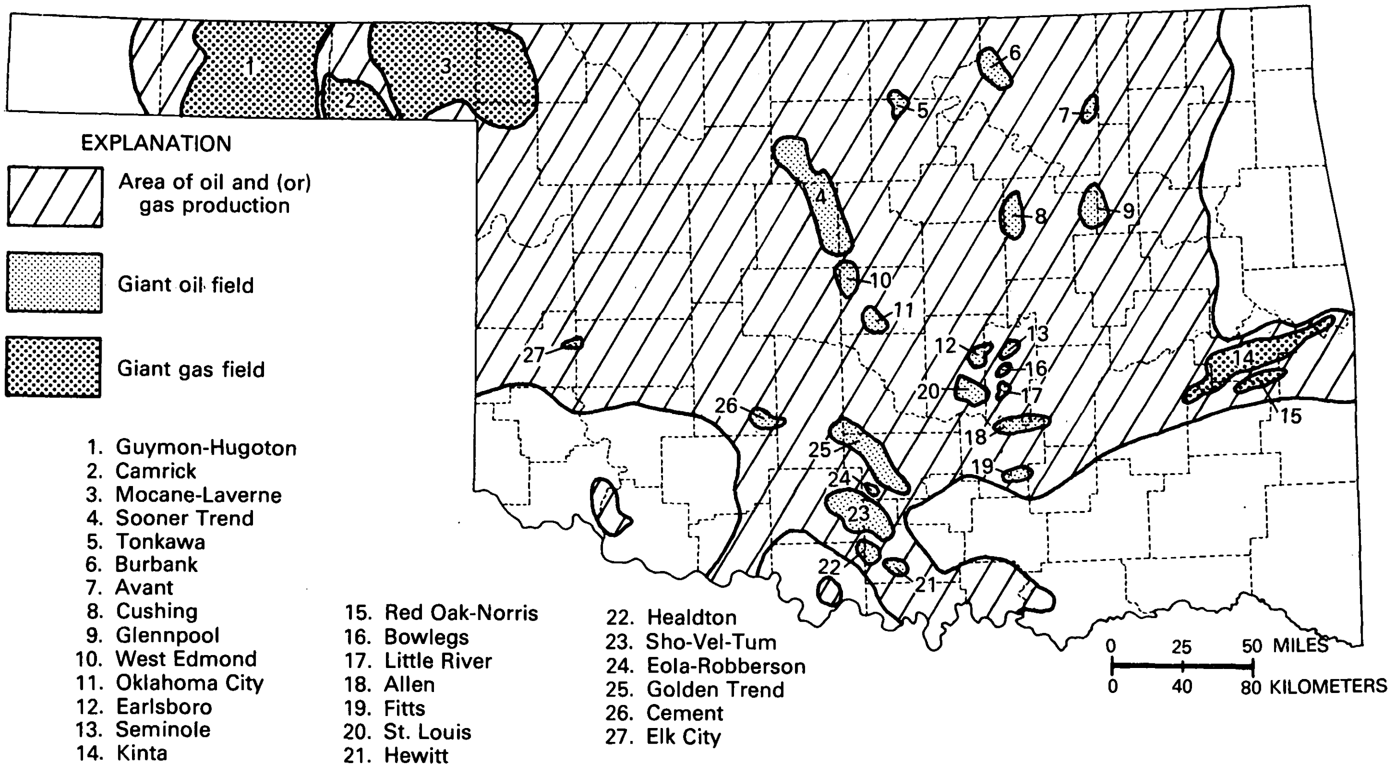

FIGURE 10.—Generalized oil and gas map of Oklahoma showing extent of productive areas and giant fields (from Johnson and others, 1972, p. 7).

Man's deepest penetration into the earth—9,583 m—was accomplished by the drilling of the Lone Star Producing Co. 1 Rogers Unit in sec. 27, T. 10 N., R. 19 W., Washita County, western Oklahoma. At this depth, in rocks of the Arbuckle Group (Ordovician), molten sulfur entered and surged up through the drill pipe; upon cooling, it crystallized in the drill pipe and caused the bottom part of the hole to be lost. The well was then plugged back to 3,962 m and completed as a gas well in Pennsylvanian rocks. The well was drilled near the axis of the deep Anadarko basin in western Oklahoma.

Oil and (or) gas has been produced from 72 of the State's 77 counties (see fig. 10). In addition to the areas of basement-rock outcrops in the Arbuckle Mountains in south-central Oklahoma and the Wichita Mountains in the southwest, the three areas that seem unlikely to produce significant hydrocarbons are the Hollis basin, the Ozark uplift, and the Ouachita Mountains.

Within the Hollis basin, in extreme southwestern Oklahoma, exploratory efforts have been largely unsuccessful in obtaining commercial production. One reason may be the lack of suitable reservoir beds. Future efforts in this rather sparsely drilled area might lead to recognition of areas where the conditions are better for the accumulation of petroleum, however.

In northeastern Oklahoma, the southwest flank of the large Ozark uplift exposes Mississippian limestones covering thin remnants of older Paleozoic sedimentary rocks over shallow Precambrian granites. Significant petroleum accumulation seems unlikely, although many outcrops and quarries in the limestone beds show heavy residual oil in joints and fissures.

In southeastern Oklahoma, the Ouachita Mountain system extends westward from Arkansas and curves southwestward to disappear under the Cretaceous overlap. South of the Choctaw fault, a structurally complex series of Pennsylvanian shales, siltstones, and sandstones is underlain by an equally complex sequence of Mississippian and older Paleozoic calcareous shales, limestones, and cherts. In the past, several asphalt mines were worked commercially in these rocks. In the western part of the region, a few noncommercial shallow oil wells have been completed in Carboniferous rocks. A few gas wells of questionable commercial value are scattered throughout the region; they produce from beds of thick, low-porosity and low-permeability sandstone and siltstone of Carboniferous age as well as from fractured chert of older formations. The economic potential of this sparsely drilled region depends on successful exploratory discoveries and on the feasibility of new methods to extract gas from these rocks of low permeability. Gas, not oil, is the mode of hydrocarbon occurrence in the deeper zones throughout the region, owing to the low-grade metamorphism that has driven out the heavier hydrocarbons.

In Oklahoma, hydrocarbons are produced from every series of the Carboniferous, but Pennsylvanian rocks account for approximately 75 percent of the Carboniferous total. The most prolific series is the Desmoinesian of Middle Pennsylvanian age. Both oil and gas are produced from Desmoinesian rocks, predominantly from sandstones associated with the cyclic sedimentation that also contains the commercial coals mined in northeastern Oklahoma. In this part of the State, the depths to these sandstone beds are as much as about 1,000 m. Equivalent beds of productive sandstone are found in most producing areas of Oklahoma, at depths of more than 3,000 m in the west-central part. None of these individual sandstone beds is continuous over a broad area. The sandstones originated as a series of deltaic deposits combined with accompanying nearshore and offshore bars and channel sands. Some production has been obtained from carbonate rocks that may have been reefal in part; these rocks are generally oolitic or oolicastic.

Next in productive importance are the many beds of sandstone within the "Atokan" and Morrowan Series of Early Pennsylvanian age. These older deposits are not as widespread as the Desmoinesian because more parts of the State served as areas of nondeposition and (or) erosion during this time. Producing sandstone beds extend to depths of 4,000 m and more in the western part of Oklahoma. These beds are not as clearly deltaic in nature as those of the younger sandstone but are more like channel and offshore deposits. The Morrowan Series contains some persistent carbonate rocks, but where productive, the facies is sandy.

Hydrocarbon accumulation in these Pennsylvanian sandstones typically is due to combinations of structural and stratigraphic traps. Facies changes from permeable sandstones to impermeable shales are the predominant entrapping mechanism. Anticlinal structures, such as local uplifts or differential compaction, contribute to accumulations. These structural features aid exploration techniques both in subsurface mapping and in seismic interpretations. Within a given reservoir, minor faulting may or may not limit production.

In the Mississippian System, limestone is the predominant rock type, followed by shale and sandstone. Sandstone in the uppermost part of the Chesterian Series, the "Springer," produces over wide areas in southern and west-central Oklahoma from several thick sections indicative of large offshore bars. In north-central Oklahoma, another widespread sandstone contributes considerable production.

Hydrocarbon accumulation in Meramecian carbonate is attributable locally to intercrystalline porosity combined with extensive vertical fractures. The fractures are due to deep-seated movement, as structure maps indicate no appreciable faulting.

Production from Osagean rocks is limited to north-central Oklahoma, where deeply eroded limestone beds have furnished an erratic reservoir called the "Mississippi Chat." Hydrocarbons are produced from highly fractured, overturned, and faulted Kinderhookian chert in a small area in southern Oklahoma.

In 1884, the Cherokee and Choctaw Indian Nations entered into agreements with eastern U.S. interests to drill for oil in northeastern and eastern parts of their territories. In 1889, oil was found in three of these wells in the northeastern area, near Chelsea. Some wells were also drilled in Muskogee in 1894.

The first well to produce oil in commercial quantities, however, was drilled in 1897 in what is now a park at Bartlesville. The well, the Cudahy Oil Co. 1 Nellie Johnstone, was drilled to 402 m in a Pennsylvanian sandstone. It produced commercially until 1946.

During the early years of this century, many large fields were discovered and developed in northeastern Oklahoma, all producing oil and (or) gas from beds of Pennsylvanian sandstone. The locations were chosen by random methods or by offsetting existing production. Many natural-gas fields were discovered in east-central Oklahoma. Exploration then began in south-central Oklahoma. Here, locations were chosen near the many known oil seeps and asphaltic deposits. Considerable production was obtained from Permian sandstones at depths of less than 100 m, but deeper drilling soon found the more productive Pennsylvanian sandstones. By 1907, the new State of Oklahoma was the Nation's leading producer of petroleum.

During the years 1908–20, the production of oil increased and exceeded demand, which resulted in depressed prices. Voluntary efforts by producers to limit production succeeded to a small extent, so that prices stabilized at higher levels. During this period, operators discovered that the services of geologists could significantly increase their successes in locating new fields and in developing older fields. Most of the production at this time came from Carboniferous rocks, but deeper drilling was adding pre-Carboniferous production.

Exploration and development from 1920 to 1930 resulted in the addition of many prolific fields, but prices remained stable until 1929, when the nearly simultaneous discoveries and development of the Seminole, Oklahoma City, and East Texas fields depressed prices to all-time lows.

During the 1930's, the practice of forced prorationing was instigated in Oklahoma. Regulatory agencies in other States adopted this policy, and prices were stabilized. At about this time, the Interstate Oil Compact Commission was formed, with headquarters in Oklahoma City. During the rest of this decade, significant discoveries and extensions continued to be made, production continued to exceed demand, and prorationing continued to drop production to even lower levels.

During the war years, demand for all petroleum products increased dramatically. Demand for petroleum slackened in the postwar years but increased gradually as reserves in Oklahoma and the Nation continued to decline.

State oil production reached a postwar peak in 1967, and its steady decline since that time is a direct reflection of new discoveries not keeping pace with depletion. This production decline cannot be reversed, but it will be slowed by future discoveries and by the development of better methods to recover a greater percentage of the oil remaining in the reservoirs.

METALLIC ORES

By Kenneth S. Johnson

The world-famous Tri-State zinc and lead deposits of northeastern Oklahoma (fig. 11) and adjacent parts of Kansas and Missouri are found in Mississippian carbonate rocks on the west flank of the Ozarks (Brockie and others, 1968; McKnight and Fischer, 1970). Sphalerite and galena occur in blanket deposits and in breccia zones that are flat-lying or that follow the predominant northeast- and northwest-trending joint systems. The host rock is cherty limestone and dolomite of the Keokuk and Warsaw Formations (Osagean and Meramecian), which sometimes are collectively referred to as the "Boone" Formation or Group. Controversy still exists about the origin and age of mineralization and paragenesis, although workers generally favor theories of emplacement coincident either with Middle and Late Pennsylvanian orogenies in the Arbuckle and Ouachita Mountains or with Late Cretaceous igneous activity of the Mississippi Valley region.

The Tri-State district, which has produced more than $2 billion worth of zinc and lead since 1848, is one of the greatest mining districts in the world. Approximately 5.2 million tons of zinc and 1.3 million tons of lead have been produced in Ottawa County, which contains the Oklahoma part of the district. Oklahoma led the United States in zinc production almost every year from 1918 through 1945. The recoverable grade of ore produced each year since 1907 ranged from about 1.5 to 4.5 percent zinc and from about 0.4 to 1.5 percent lead. The depletion of higher grade ores and decline of the market price caused Oklahoma production to fall sharply in recent years, until the last commercial mine ceased operations and was abandoned late in 1970. Production came from underground mines in Oklahoma, generally at depths of 30–100 m, and all that remains to indicate this past mining are the abandoned mines and the large piles of chat, or crushed stone, removed during milling of the ores.

Other small vein deposits of lead, zinc, and (or) copper minerals scattered in the Ouachita, Arbuckle, and Wichita Mountains and in the Ozark region are perhaps of Middle or Late Pennsylvanian age.

NONMETALLIC MINERALS

By Kenneth S. Johnson

Principal nonmetallic-mineral resources in Mississippian and Pennsylvanian rocks are limestone and shale. Resources being produced in lesser quantities are tripoli, sandstone, chat, iodine, and ground water (fig. 11).

Limestone.—Thick Mississippian limestone units crop out extensively in the Ozarks and in scattered parts of the Arbuckle Mountains. Pennsylvanian limestone units are thinner, typically 2–15 m thick, and they cap a series of subparallel escarpments in northeastern Oklahoma and in small parts of the Ouachita and Arbuckle Mountains. These limestone units are not of exceptional purity, but they are widely used in the construction industry as a source of aggregate, cement, and dimension stone. Mississippian and Pennsylvanian limestones are currently being worked in about 30 quarries and 1 underground mine.

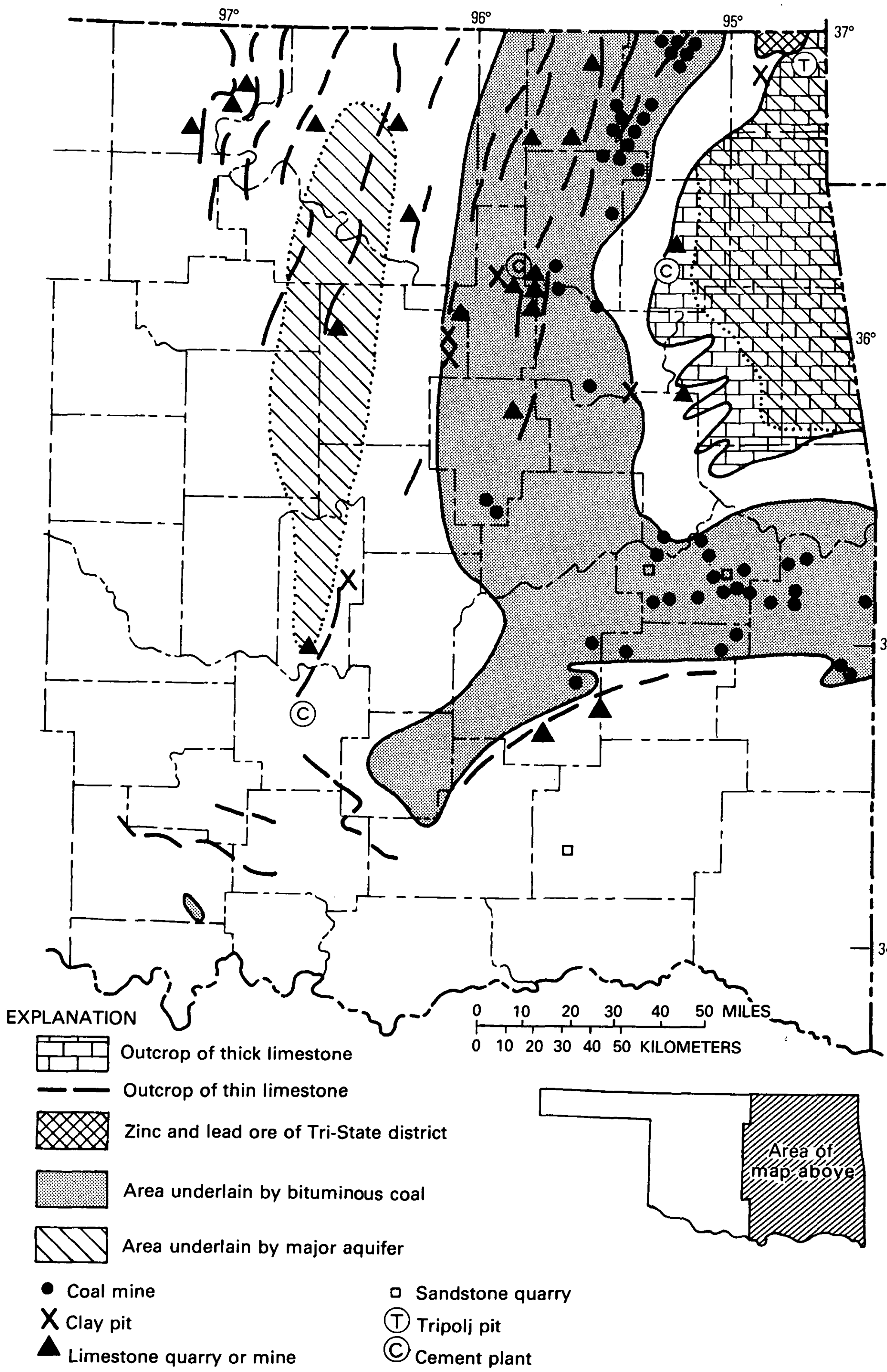

FIGURE 11.—Distribution of principal nonpetroleum mineral resources in Carboniferous rocks of Oklahoma (from Johnson, 1969; Johnson and others, 1972, p. 6).

Shale.—Pennsylvanian shale units that are 10–100 m thick and consist mainly of illite, chlorite, and quartz are a major source of raw material for the manufacture of fired-clay products in Oklahoma. Red brick and cement are the principal products made from these shales, but clay pipe, expanded lightweight aggregate, and pottery are also being produced. Nearly 10 shale pits are now being operated in Pennsylvanian shale beds, and 1 pit is being worked in a Mississippian shale. Underclay beneath Desmoinesian coal beds is generally 0.3–1.0 m thick, and a recent study indicates that it may be suitable locally for making light-colored brick and low-grade refractories.

Tripoli.—Moderately large deposits of tripoli, a lightweight form of silica rock used chiefly for abrasive and absorbent properties, are being mined from small open pits in Ottawa County in the northeast. Tripoli occurs as lenses 1–4 m thick in cherty limestones of the Mississippian "Boone" Group.

Sandstone.—Thin-bedded and well-cemented Pennsylvanian sandstone beds in parts of eastern Oklahoma have been quarried for building stone. Gray, brown, and buff stone is slabbed by splitting it along micaceous bedding planes.

Chat.—Chat is the name applied locally to the crushed chert, limestone, and dolomite produced as a byproduct of milling zinc and lead ores from the Mississippian "Boone" Group in the Tri-State district of northeastern Oklahoma. The large piles of this waste material are now being used extensively as crushed stone for road metal, railroad ballast, and concrete aggregate.

Iodine.—Early in 1977, iodine production began near Woodward in northwestern Oklahoma. Production is from natural brines in the basal Pennsylvanian (Morrowan) sandstones at a depth of about 2,100 m. The concentration of iodine in the brines ranges from 10 to 700 ppm and averages about 300 ppm. The 909,000 kg of iodine produced annually at this plant will supply more than one-third of the Nation's requirements.

Ground water.—Two major ground-water aquifers occur in Carboniferous rocks. The Mississippian Keokuk and Reeds Spring Formations (sometimes referred to the "Boone" Formation or Group) consist of fractured cherty limestone that yields large quantities of good water to shallow wells and springs in the Ozark uplift area of the northeast. The second major aquifer is the Pennsylvanian Vamoosa Formation (Virgilian), which consists of sandstone, shale, and conglomerate in the central and north-central part of the State.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Texas

By R. S. KIER, L. F. BROWN, JR., *and* E. F. MCBRIDE

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areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in Texas

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—TEXAS

By R. S. KIER,[1] L. F. BROWN, JR.,[2] and E. F. MCBRIDE[3]

ABSTRACT

Carboniferous rocks in Texas crop out in the Colorado, Brazos, and Trinity River valleys in central and north-central Texas and in the Trans-Pecos region of west Texas. In central and north-central Texas, Mississippian and Pennsylvanian strata are mainly shale, sandstone, and limestone deposited in fluvial-deltaic and interdeltaic environments, on open shelves and carbonate platforms, and in shelf-edge, slope, and basin environments. In west Texas, Carboniferous rocks are exposed in the Marathon and Solitario uplifts, and in the Sierra Diablo, Hueco, and Franklin Mountains. In the Marathon region, the principal site of Carboniferous deposition in west Texas, deepwater sandstone and shale (flysch) are capped by shallow-water shale, limestone, and conglomerate. Mississippian and Pennsylvanian rocks in the Franklin and Hueco Mountains are chiefly limestone, marl, and shale.

Carboniferous geology of central and north-central Texas is closely tied to the tectonic development of the Fort Worth (foreland) basin, the eastern shelf of the Midland basin, and the Red River uplift–southern Oklahoma mountains. In response to Late Mississippian and Early Pennsylvanian structural activity in the Ouachita geosyncline, the Fort Worth basin became well defined. Thick, westward-prograding terrigenous clastic wedges (Atoka Group) of probable fan-delta and related slope origin entered the basin along high-gradient paleoslopes from the Ouachita foldbelt. Fan deltas shifted westward over thin, relatively starved basinal Smithwick facies. Platform and shelf-edge carbonate environments (Marble Falls, Big Saline, Comyn, and Caddo) contemporaneously dominated the Concho platform. Eastern shelf edges of the Concho platform retreated periodically westward in response to westward-shifting fan-delta environments. Fan deltas reached the western flank of the Fort Worth basin late in the waning stages of Atoka deposition.

Decreased subsidence of the Fort Worth basin and diminished Atoka sediment supply marked the deceleration of Ouachita orogenic activity. Consequently, Strawn (Desmoines Series) deposition was dominated by fluvial-deltaic systems that overlapped shelf-edge carbonate facies (Marble Falls, Caddo, Big Saline) and prograded repeatedly across the shallow Concho platform. Youngest Smithwick prodelta-basinal facies were deposited in the path of the initial delta system to prograde over the Concho platform.

As the source areas were lowered by erosion and as paleogradients were diminished, less terrigenous sediment reached the Pennsylvanian coastline. Extensive and long-lived carbonate-bank and reef systems of the Canyon Group (Missouri Series) began to form on the stable platforms provided by abandoned Strawn deltas. Some of the carbonate banks, growing on the western edge of the structurally positive Eastern shelf of the rapidly subsiding Midland basin, extended upslope and intertongued with Canyon delta systems. Rejuvenation in the Ouachita foldbelt and eastern Fort Worth basin significantly increased sediment supply and initiated extensive lower Cisco (Virgil Series) delta-fluvial deposition. Cisco deltas prograded westward across the relatively stable Eastern shelf, overlapping Canyon carbonate facies and supplying sediment to thick, basinward-prograding slope and submarine-fan environments in the Midland basin. During deposition of the upper Cisco Group (Permian, Wolfcamp Series), sediment supplied from the east again diminished, and thick, low-relief, shelf-edge limestone banks became increasingly prominent.

The complex history of the Red River uplift-Oklahoma mountains structural elements is recorded by thick clastic wedges extending southward and southwestward into the subsurface of north-central Texas. These arkosic sediments represent fluvial and fan-delta deposition along steep paleoslopes adjacent to fault blocks in north Texas and southern Oklahoma. Fan-delta deposition was contemporaneous with limestone deposition on adjacent, structurally positive blocks.

The Marathon region of west Texas was the site of slope and deep basinal sedimentation during most of the Paleozoic. Radiolarian chert and shale in the upper part of the Caballos Novaculite (Lower Mississippian?) were deposited probably in water depths greater than 1,000 m. Black mud of the Tesnus Formation was followed by deposition of thin distal turbidities and shale that reflect progradation of a delta system from east to west. Siliciclastic detritus was derived almost entirely from Llanoria (Africa and South America).

Uplift of the western margin of the Ouachita geosyncline initiated an episode of calcareous flysch deposition (Dimple Formation). Sediment derived from carbonate banks on the shelf and from uplifted older rocks was transported into the basin as slides, debris flows, and turbidity currents. Renewed uplift of Llanoria brought a return to siliciclastic flysch deposition of the Haymond Formation beginning with black mud and followed by alternating turbidites and pelagites and by an olistostrome. Turbidite-pelagic deposition continued after formation of the olistostrome, but younger sandstone beds locally are burrowed, suggesting that the geosyncline was becoming shallower. The Haymond passes

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upward into shelf and slope deposits of the Gaptank Formation in the northern part of the outcrop belt.

Carboniferous rocks in central, north-central, and west-central Texas have contributed significantly to the economic development of Texas and to the Nation. Oil and gas production has dominated the economic picture, but industrial and ceramic clays, coal, and constructional limestone historically have been locally and periodically important. Potentially, uranium may be found within Carboniferous rocks in sufficient quantities to warrant further intensive exploration. Ground-water potential from Carboniferous rocks is poor and has not been significantly exploited. Resources of economic value have not been recognized in the Marathon basin or in the Franklin and Hueco Mountains.

INTRODUCTION

LOCATION AND EXTENT

Carboniferous rocks in Texas crop out in the Colorado, Brazos, and Trinity River valleys in central and north-central Texas and in Trans-Pecos Texas (figs. 1 and 2). Mississippian and lowermost Pennsylvanian rocks are exposed in isolated areas within and as relatively continuous exposures along the eastern, northern, and western margins of the central mineral region (Llano uplift; figs. 2 and 3). In north-central Texas, younger Pennsylvanian rocks are exposed in a north-northeast-trending belt from the central mineral region to the Red River (fig. 1). Carboniferous outcrops in central and north-central Texas cover approximately 15,675 km^2.

Within the Trans-Pecos region (fig. 1) of Texas, Mississippian and Pennsylvanian rocks are exposed in the Marathon uplift (400 km^2), Solitario uplift (4 km^2), Sierra Diablo Mountain (1 km^2), Hueco Mountains (20 km^2), and Franklin Mountains (10 km^2). Rocks that crop out in the Marathon and Solitario uplifts are similar, and only the Marathon uplift is discussed. Outcrops in the Sierra Diablo Mountain, which are poorly understood, are not included in this report.

In central and north-central Texas, a maximum of 1,333 m of Pennsylvanian rocks is exposed, but only 15 m of Mississippian rocks crops out locally in paleosinkholes. Approximately 2,800 m of Pennsylvanian rocks and 1,500 m of Mississippian rocks are exposed in the Marathon uplift of west Texas. In the Franklin Mountains, 965 m of Pennsylvanian rocks and 150 m of Mississippian rocks have been measured, and in the Hueco Mountains, approximately 400 m of Pennsylvanian rocks has been recognized. The basal part of the Mississippian section in the Hueco Mountains is not exposed.

This chapter was critically reviewed by Shirley J. Dutton, David K. Hobday, and Mark W. Presley of the Bureau of Economic Geology.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Bureau of Economic Geology, The University of Texas at Austin.

GENERAL GEOLOGY

In central and north-central Texas, Mississippian and Pennsylvanian strata consist of shale, sandstone, and limestone (fig. 4). Locally, conglomerate and coal deposits are found in the clastic sequences. Vertical distribution of rock types is generally predictable and provides the basis for stratigraphic classification. Pennsylvanian strata were deposited principally in deltas and interdeltaic embayments, on open shelves and carbonate platforms, and in slope and basin environments. Some Mississippian shale accumulated in relatively shallow starved basins.

Outcropping Mississippian and Pennsylvanian rocks in the Marathon uplift consist of about 3,600 m of deepwater (flysch) deposits capped by about 550 m of shallow-water shale, limestone, and conglomerate (fig. 4). Sandstone and shale make up 85 percent of the flysch units. One flysch sequence, however, the Dimple Formation, is characterized by calcarenite turbidites, chert, and shale. Olistostromes (boulder beds of submarine debris-flow origin) are found in all flysch deposits. Radiolarian chert beds in the upper part of the Caballos Novaculite may be of Mississippian age.

Mississippian and Pennsylvanian rocks exposed in the Franklin and Hueco Mountains (fig. 4) are chiefly limestone, cherty limestone, marl, and shale; chert conglomerate and sandstone are minor rock types. Mississippian sequences contain more shale and marl (15–20 percent) than Pennsylvanian sequences.

PREVIOUS WORK

The presence of Carboniferous rocks in Texas was first reported by Ferdinand Roemer (1848) on the basis of observations made during extensive travels in central Texas during German colonization. A few years later, Roemer (1852) published descriptions of Carboniferous fossils collected at two localities, probably in the Canyon Group and in the Marble Falls Formation (Gries, 1970). Other early explorers who reported Carboniferous-age rocks in Texas include Schumard (1854), Marcou (1856), Shumard (1860), Ashburner (1881), and Glenn (*in* Comstock, 1890).

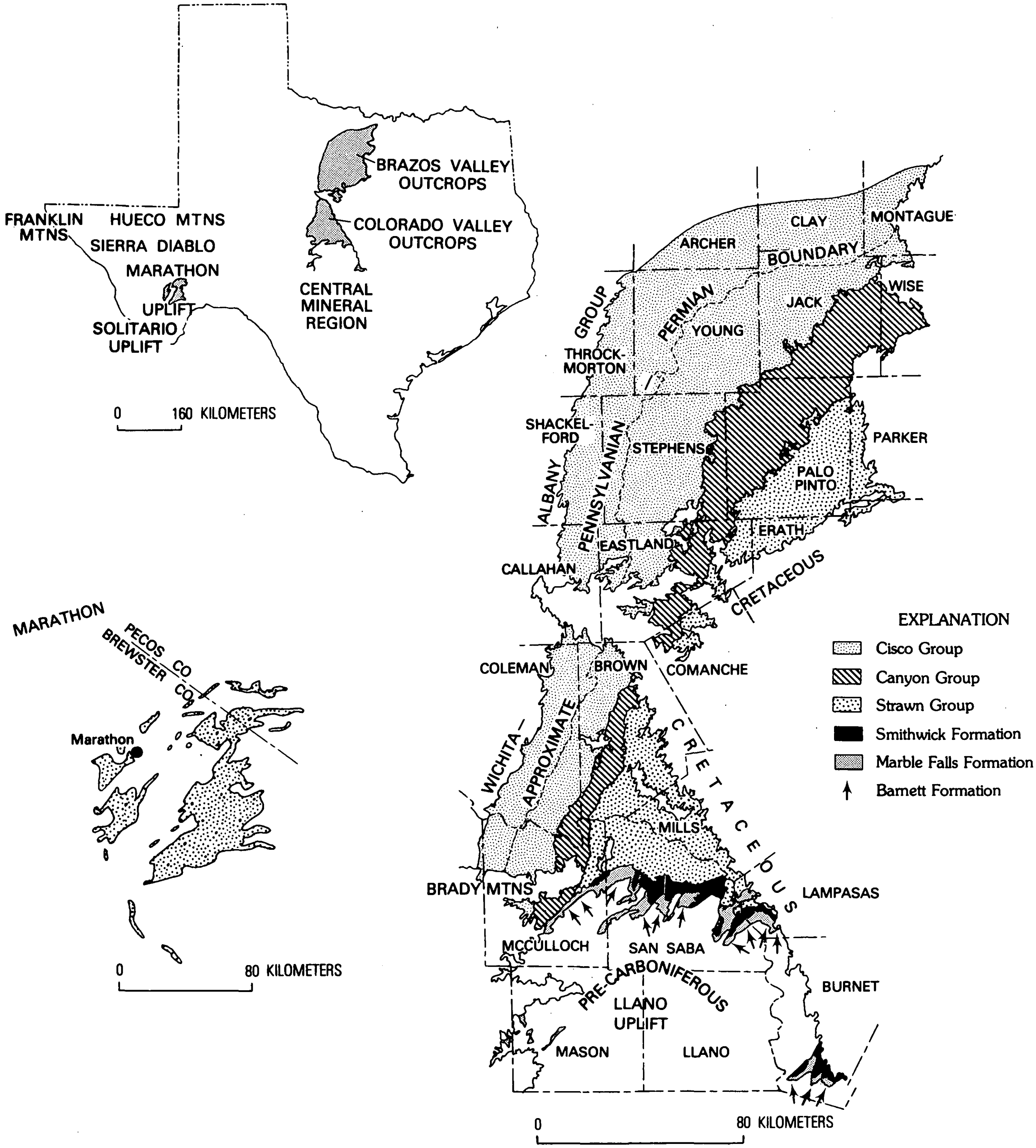

FIGURE 1.—Location of outcropping Mississippian and Pennsylvanian rocks in Texas.

Earliest systematic studies of Carboniferous rocks in Texas were published in annual reports of the third Geological Survey of Texas (1889–1901). In the First Annual Report, Dumble (1890) divided Carboniferous rocks of central Texas into "series," including the Bend series, Richland-Gordon Sandstones, Milburn-Strawn series, Brownwood-Ranger series, Wildrip-Cisco series, and Coleman-Albany

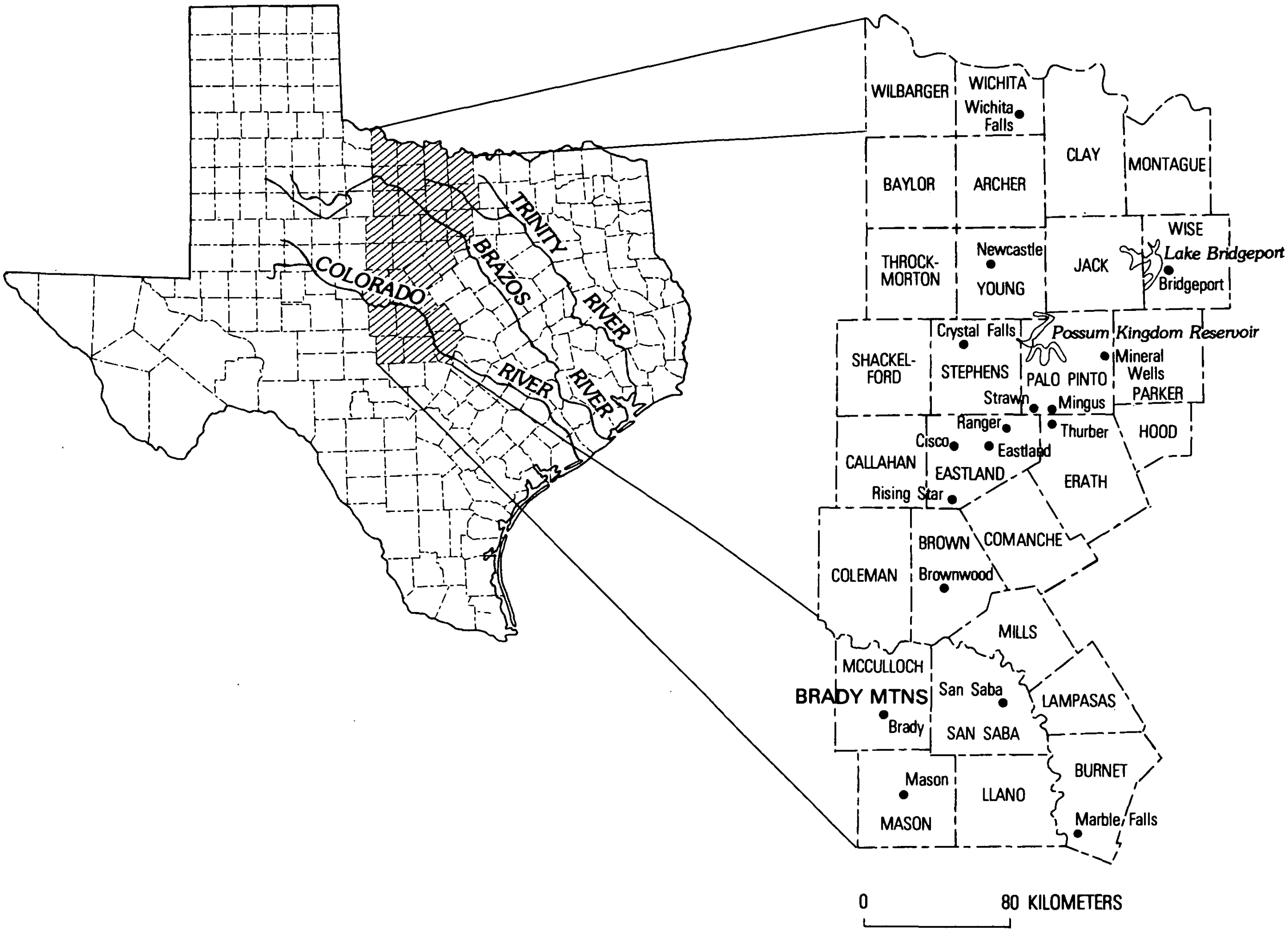

FIGURE 2.—Geographic index to central and north-central Texas.

series. Cummins (1890) and Tarr (1890) described the results of field explorations of the Colorado River valley outcrops. In the Second Annual Report, Cummins (1891) described the general geology of central and north-central Texas, defining five Carboniferous and four Permian "divisions." He also compared the stratigraphy of the central (Colorado River valley) coal fields with that of the northern (Brazos River valley) coal fields. In the Fourth (and last) Annual Report, Drake (1893) described the geology of the "Colorado Coal Field" and named or numbered numerous "beds," many of which subsequently became formal members and formations for central and north-central Texas.

Most early geologic investigations of Carboniferous strata in Texas were centered in central and north-central Texas (for example, Hill (1889, 1901), Paige (1911, 1912), Udden (*in* Udden and others, 1916), and Bridge and Girty (1937)). Plummer (1919) proposed a preliminary classification of Carboniferous rocks for the Brazos River valley. Later, Plummer and Moore (1921; see also Moore and Plummer, 1922) presented a more comprehensive lithostratigraphic classification of Pennsylvanian rocks in the Colorado and Brazos River valleys, including formations and groups. Sellards and others (1932) modified Plummer and Moore's classification, but continued to classify on the basis of lithologic characteristics.

Cheney (1940, 1947, 1948, 1949, and 1950; West Texas Geol. Soc., 1951; Cheney and others, 1945) initiated a major change in the approach to Pennsylvanian stratigraphic classification in Texas. Following concepts of Moore (1936) in Kansas, Cheney proposed a provincial time-stratigraphic classification of Pennsylvanian rocks in central and north-central Texas based on inferred faunal changes and unconformities which were thought to be regionally significant. The Strawn, Canyon, and Cisco were used as series names, and a new Lampassas series

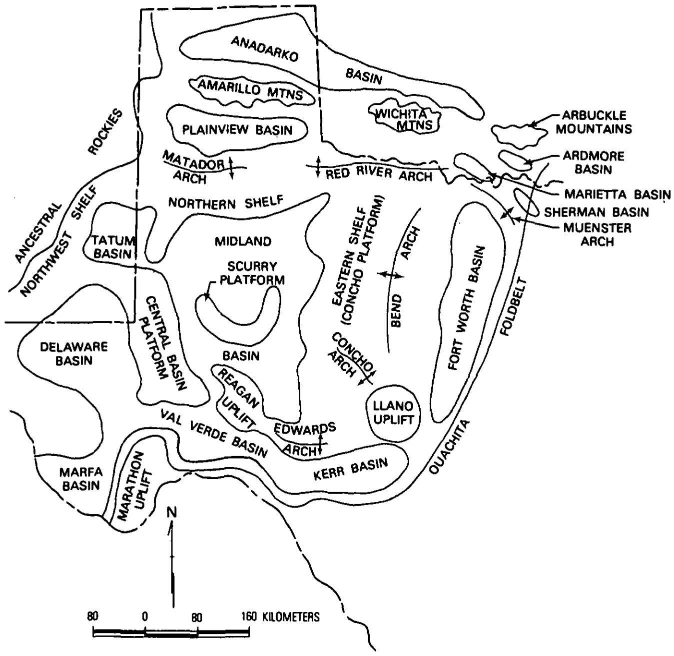

FIGURE 3.—Carboniferous tectonic elements in Texas and southern Oklahoma. Modified from Wermund and Jenkins (1970) and Galloway (1970).

was proposed to encompass strata of the Big Saline and Smithwick "Groups". Conceptually, this classification was intended to facilitate correlation between different basins or areas of exposure within a basin. Although several attempts have been made to apply Cheney's classification scheme in the field (Cheney and Eargle, 1951; Shelton, 1958), such application has proven difficult, if not completely inappropriate (see Brown, 1959; Brown and Goodson, 1972).

Other contributors to the stratigraphy of Mississippian and Pennsylvanian rocks in central and north-central Texas include Plummer (1945; 1947a, b; and 1950), Eargle (1960), Stafford (1960), Terriere (1960), and Myers (1965). Students at The University of Texas at Austin, Baylor University, Southern Methodist University, and Texas Christian University mapped Mississippian and Pennsylvanian rocks and described fossils in central and north-central Texas as part of thesis studies. Recent contributors to Pennsylvanian geology in the region include Bretsky (1966), Brooks and Bretsky (1966), Brown (1960a, b; 1962; 1969a, b, c, and d), Brown and others (1973), Feray and Brooks (1966), Galloway and Brown (1972 and 1973), Laury (1962), Wermund (1966, 1969, and 1975), Wermund and Jenkins (1964, 1969, and 1970), Erxleben (1975), and Cleaves (1975). Regional surface mapping of Mississippian and Pennsylvanian rocks was carried out by Brown and Goodson (1972) and by Kier and others (1976 and unpub. data).

Investigation of Mississippian and Pennsylvanian strata in Trans-Pecos has not been as extensive as in central and north-central Texas. Most studies have been concentrated in the Marathon uplift (figs. 2 and 3). Earliest studies of Pennsylvanian and Mississippian strata were by Baker (*in* Udden and others, 1916) and by Baker and Bowman (1917). Later, King (1931, 1934, and 1937) published a series of reports on Trans-Pecos geology including his landmark study of the Marathon uplift. More recent studies include those by Fan and Shaw (1956), Berry and Nielson (1958), Cotera (1962 and 1969), Johnson (1962), McBride (1966 and 1970), Ross (1962, 1963, 1965, 1967, and 1969), and McBride and Thompson (1970). Additional references on the Marathon region and the Franklin and Hueco Mountains are listed in table 1.

REGIONAL SETTING

CENTRAL AND NORTH-CENTRAL TEXAS

Mississippian and Pennsylvanian rocks exposed in central and north-central Texas were deposited (1) on the Llano uplift (fig. 3), a large structural dome (Cloud and Barnes, 1948) that has a core of Precambrian igneous and metamorphic rocks; (2) on the moderately stable Concho platform during Late Mississippian and Early Pennsylvanian (Cheney, 1929; Cheney and Goss, 1952); and (3) on the eastern shelf of the west Texas basin during Middle and Late Pennsylvanian. East of the Llano uplift and the Concho platform are the Fort Worth basin and the Ouachita foldbelt (Flawn and others, 1961); south of the uplift is the Kerr basin.

During Early Mississippian, limestone, shale, and chert breccia accumulated on a pre-Carboniferous karstic erosion surface on the Llano uplift. Middle and Late Mississippian shale accumulated in a starved basin west of the Ouachita geosyncline.

Beginning in Late Mississippian and during Early Pennsylvanian, orogenic activity in the Ouachita geosyncline produced a thrust-faulted foldbelt (fig. 3). The Ouachita Mountains served as a source of sediment during the rest of the Paleozoic. During the same time, the Fort Worth basin formed as a foreland trough between the rising Ouachita Mountains and the older Concho platform. The basin was initially filled by thick terrigenous clastic wedges of mudstone and sandstone deposited by prograding fan deltas and related slope systems. Clastic wedges grade westward into starved basinal shale of the

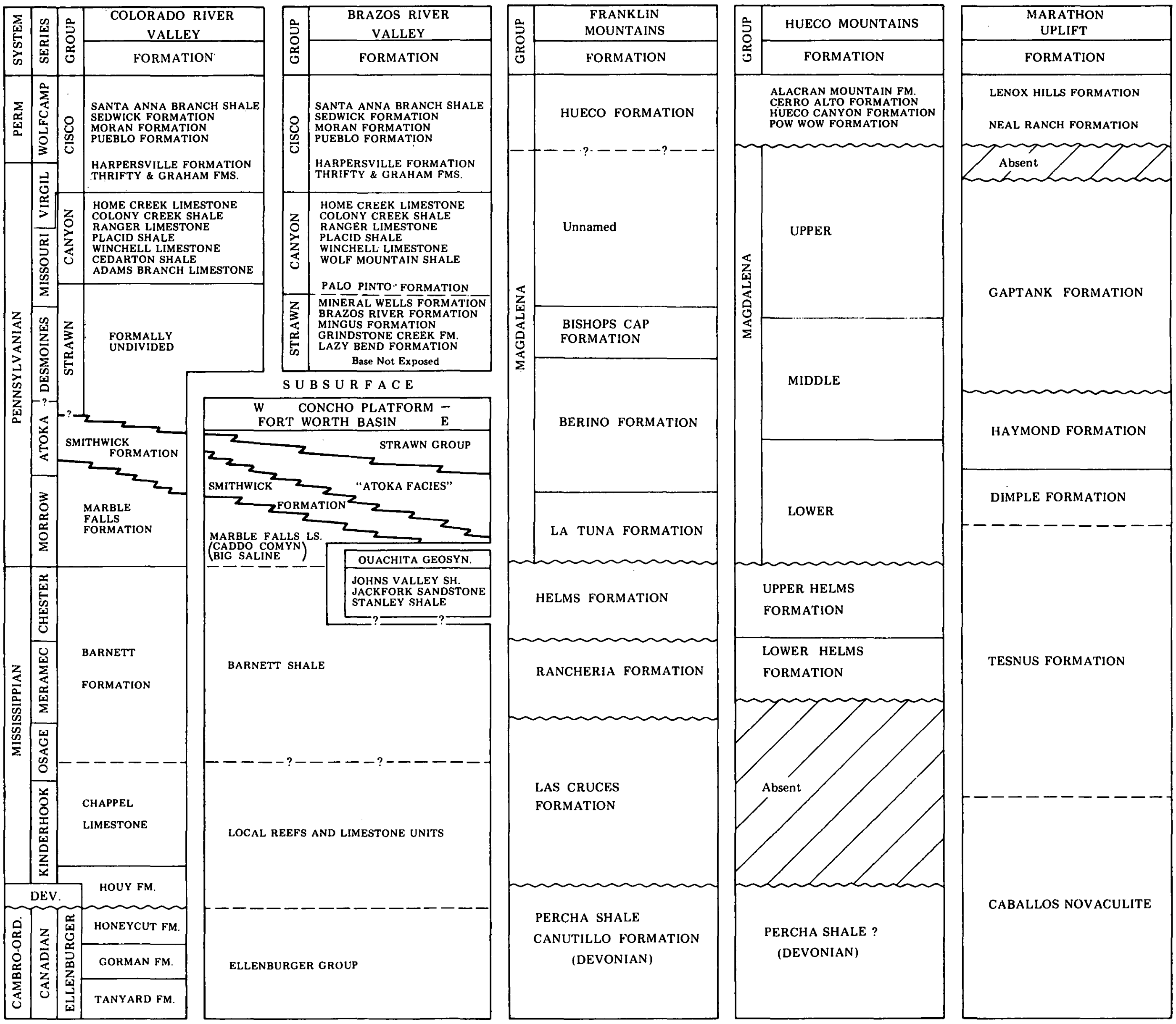

FIGURE 4.—Geologic units in the Carboniferous of central, north-central, and west Texas.

western Fort Worth basin and eastern Concho platform.

The Red River, Muenster, and Matador arches (fig. 3) are a discontinuous series of uplifted fault blocks across northwest Texas that served as foundations for major carbonate sequences and, from time to time, as a source of arkosic sediment. The Wichita, Arbuckle, and Amarillo Mountains of southern Oklahoma and the Texas Panhandle provided arkosic sediment to fluvial and fan-delta systems in north-central Texas.

The Midland basin, a moderate-sized interior basin, formed in west Texas during and after Ouachita deformation (Galloway and Brown, 1972). Fluvial and delta systems originating in the Ouachita Mountains and the uplifted eastern part of the Fort Worth basin prograded westward across the earlier Concho platform. Cyclic deposition by these clastic systems progressively constructed the eastern shelf of the Midland basin. Carbonate shelf and shelf-edge systems formed along the westward margin of the Eastern shelf as the depth of the basin increased and slope deposition was initiated.

Cyclic progradation of fluvial and delta systems westward across shelf and shelf-edge carbonate environments continued throughout the Middle and

TABLE 1.—*Contributors to stratigraphic data on the Marathon basin and the Franklin and Hueco Mountains*

Marathon basin	Franklin and Hueco Mountains
Aberdeen, 1940	King, 1934
Baker, 1963	King and others, 1945
Brooks, 1955	Laudon and Bowsher, 1949
Ellison, 1962	Nelson, 1940
Flawn, 1958	Seewald, 1968
Flawn and others, 1961	Stewart, 1958
Flores, 1972, 1977	Williams, 1963
Folk, 1973	
Goldstein and Hendricks, 1962	
King and King, 1929	
King and others, 1931	
Skinner and Wilde, 1954	
Thomson and Thomasson, 1969	
Waterschoot van der Gracht, 1931	

Late Pennsylvanian and Early Permian. Large volumes of sediment were transported across the Eastern shelf through fluvial and deltaic channels. Where deltas reached the shelf edge, shallow-marine sediments were redeposited in submarine fans on the floor of the Midland basin. Relief between the floor of the Midland basin and the Eastern shelf reached 455 m in the Late Pennsylvanian (Brown, 1973a).

As the Midland basin subsided, regional upwarping of the Ouachita foldbelt and the eastern flank of the Fort Worth basin took place. The hinge or axis of rotation between the rising Fort Worth basin and the downwarping Midland basin defines the Bend arch. Erosion of uplifted Lower Pennsylvanian sediments in the Fort Worth basin contributed considerable amounts of second-cycle sediment to Upper Pennsylvanian delta and slope environments.

Principal tectonic elements such as the Llano uplift, Ouachita foldbelt, Fort Worth basin, Concho platform, Eastern shelf, Bend arch, Red River arch trend, Oklahoma mountains, and Midland basin significantly determined the nature of depositional environments in which Mississippian and Pennsylvanian sediments accumulated. The interplay among orogenic pulses in the Ouachita and Oklahoma mountains, uplift of the eastern flank of the Fort Worth basin, subsidence of the Midland basin and westward tilting of the Concho platform affected sediment supply and water depth. Depositional processes operative in a myriad of fluvial, deltaic, embayment, shelf, platform, slope, and basin environments produced the many Mississippian and Pennsylvanian rock types.

WEST TEXAS

Pre-Permian Paleozoic rocks of the Marathon uplift in Trans-Pecos Texas were deposited in the Marathon trough, a segment of the Ouachita geosyncline that extended from Mexico to Arkansas (fig. 3). Although there are dissenting opinions (Folk, 1973; Flores, 1972, 1977), the Marathon trough apparently was a deep basin throughout pre-Permian–Paleozoic time. From Late Cambrian to Carboniferous time, the trough received about 1,000 m of slope and basinal sandstone, limestone, chert, shale, and olistostromes. This "early geosynclinal" phase of slow deposition culminated with deposition of the Caballos Novaculite.

Deposition of predominantly flysch rocks generally coincided with the beginning of the Carboniferous. During this "late or filling stage" of geosynclinal history, approximately 3,500 m of flysch deposits (Tesnus, Dimple, and Haymond Formations) and 550 m of shallower marine deposits (Gaptank Formation) accumulated in only 60 m.y. During the filling stage, most terrigenous detritus was derived from a continental mass east of the geosyncline, an element designated Llanoria by Dumble (1920) and other early workers but considered part of Africa or South America in newer plate-tectonic restorations (Rowett and Walper, 1973; Keller and Cebull, 1973). Locally, carbonate detritus and exotic blocks were derived from a positive cratonic element west of the depositional basin (McBride, 1970).

The Ouachita geosynclinal sequence underwent several pulses of deformation and mountain building from Desmoines to Middle Wolfcamp(?) time (King, 1937; Ross, 1962). The sequence was folded, faulted, and thrust northwestward at least 70 km along a major décollement surface; the deformed sequence is underlain by relatively undeformed foreland rocks. The Marathon region underwent broad domal uplift and normal faulting early in Tertiary time (King, 1937).

The Franklin Mountains and Hueco Mountains are north-trending Laramide fault blocks. About 1,700 m of Paleozoic rocks, chiefly limestone, is exposed in the Franklin Mountains. About 1,000 m of Silurian and younger Paleozoic rocks, chiefly limestone and shale, is exposed in the Hueco Mountains.

CARBONIFEROUS STRATIGRAPHY

CENTRAL AND NORTH-CENTRAL TEXAS

Major lithologic divisions of the Carboniferous of central and north-central Texas (fig. 4) are based on variations in the vertical succession of the stratigraphic sequence. These principal sequences record filling of the Fort Worth basin, cyclic sedimentation on the Eastern shelf, and partial filling of the Midland basin (fig. 3).

The Lower Mississippian (and Devonian) Houy Formation is a relict or lag deposit preserved in sinks and depressions in an erosion surface on the Llano uplift. Middle and Upper Mississippian Chappel Limestone and shale of the Barnett Formation were deposited during initial marine transgression onto the Llano uplift. In outcrop, the Lower Pennsylvanian Marble Falls Formation, predominantly limestone, represents establishment of carbonate platform and shelf environments on the Llano uplift adjacent to the Fort Worth basin. Similar Upper Mississippian and Lower to Middle Pennsylvanian shelf and shelf-edge facies were deposited west of the Fort Worth basin on the Concho platform.

The Smithwick Shale and the overlying Strawn Group record the final phase of filling of the Fort Worth basin and initial westward progradation of deltas onto the Concho platform (Cleaves, 1973, 1975). Diminished terrigenous influx and increased carbonate shelf and bank deposition distinguish the Canyon Group from the underlying Strawn and overlying Cisco Groups. Several medium to thick limestone units that were deposited in platform and open-shelf environments intertongue updip (eastward) with deltaic deposits and grade downdip into shelf-edge reef and bank deposits at the eastern margin of the Midland basin (Erxleben, 1973, 1975).

The Cisco Group records renewed deposition of terrigenous clastic materials and predominance of fluvial and deltaic environments. From 10 to 15 fluvial-deltaic progradational sequences can be recognized in the Cisco Group, each one terminated upward by transgressive sandstone and open-shelf limestone facies (Brown, 1973b). Downdip the clastic facies intertongue with extensive shelf and shelf-edge limestone deposits. Fluvial-deltaic deposits were the principal sources of sediment that ultimately was redeposited by density flows to produce thick, off-lapping wedges of deepwater deposits in the eastern part of the Midland basin (Galloway and Brown, 1972, 1973).

MISSISSIPPIAN SYSTEM

HOUY FORMATION

The Houy Formation established by Cloud and others (1957) includes strata transitional across the Devonian-Mississippian boundary. It is divided into two formal members, the Ives Breccia and the Doublehorn Shale, and several unnamed members.

The basal Ives Breccia Member is a poorly sorted, multicolored angular to subangular chert breccia with a matrix of medium to coarse, angular to subangular chert and clear quartz sand. Silicified crinoid fragments and conodonts are fairly common (Kier, 1972). Hematite, partly weathered to limonite, is very common. Maximum thickness of this member is about 1 m.

The overlying Doublehorn Shale Member is black, fissile, slightly radioactive shale that contains spores of unknown origin and silicified pieces of *Callixylon* (Cloud and others, 1957). It weathers light brown and is as much as 4.5 m thick. Other unnamed lithic units in the Houy Formation include siliceous limestone and silty calcareous shale below the Ives and phosphorite beds above the Doublehorn Shale.

The Ives Breccia Member crops out on the eastern, northern, and western sides of the Llano uplift (fig. 3). The Doublehorn Shale and other unnamed members in the Houy are found only on the northeastern side of the uplift. Poor exposure and isolation of the outcrops make interpretation difficult.

The Ives Breccia apparently is a lag deposit of locally derived chert weathered from limestone and dolomite beds of the underlying Ellenburger Group and deposited during one or more Upper Devonian and Lower Mississippian marine transgressions (Zachry, 1969; Kier, 1972). The Doublehorn Shale may be the offshore facies equivalent of the Ives (Kier, 1972). Cloud and others (1957) placed the Devonian-Mississippian boundary within the Ives, and Seddon (1970) suggested that the boundary is a disconformity and that the Doublehorn Shale is, in part, time-equivalent to the Ives. Also, at least part of the Ives Breccia may have been deposited more or less contemporaneously with the Chappel Limestone (Kier, 1972).

CHAPPEL LIMESTONE

The Chappel was named by Sellards (Sellards and others, 1932) for thin crinoidal limestone beds lying directly on the Ellenburger Group. It is predominantly a fine- to very coarse grained, poorly sorted, packed, ostracode-bearing, algal, crinoidal biomicrite and poorly washed biosparite (Kier, 1972). Most of the Chappel is light to medium greenish gray or dusky yellow. On the basis of conodonts, Hass (1959) concluded that the age of the Chappel is late Kinderhook to early Osage.

Chappel outcrops are scattered across the eastern, northern, and western sides of the Llano uplift. Commonly, the Chappel lies within or adjacent to sinks in the Ellenburger Group. Whether these sinks formed before, during, or after deposition of the Chappel Limestone is uncertain (Cloud and Barnes, 1948; Freeman, 1962; Turner, 1970). Thickness of the unit varies from about 10 cm to about 10 m.

The Chappel Limestone is apparently conformable with the underlying Doublehorn Shale Member and possibly the Ives Breccia Member of the Houy Formation. Seddon (1970) reported a continuous succession of conodonts from the Doublehorn Shale into the overlying Chappel. Locally, the Ives Breccia is adjacent to, within, or above the Chappel (Turner, 1970; Seddon, 1970; and Kier, 1972). Elsewhere, the Chappel Limestone unconformably overlies the Ellenburger Group.

Sellards, (*in* Sellards and others, 1932), Cloud and Barnes (1948), and Freeman (1962) stated that the overlying Barnett Formation was deposited unconformably on the Chappel Limestone. More recently, Zachry (1969), Turner (1970), and Kier (1972) concluded that the Barnett is probably conformable with the Chappel. Shale beds within the Chappel Limestone are similar to Doublehorn and Barnett shales (Rose, 1959; Winston, 1963; Turner, 1970; Kier, 1972). All investigators except Winston concluded that the Chappel is the shoreline or nearshore equivalent of the Barnett Formation. Deposition may have taken place during more than one marine transgression (Zachry, 1969; Kier, 1972). Whether the present distribution of Chappel outcrops approximates the original distribution of Chappel depositional environments or whether Chappel environments were much more extensive is uncertain.

BARNETT FORMATION

The Barnett Formation and subjacent Chappel Limestone lie between the Ordovician Ellenburger Group and the Pennsylvanian Marble Falls Formation and document major Mississippian marine transgressions across the Llano uplift. The Barnett was established by Plummer and Moore (1921). Previously, the shale unit was called the lower shale of the Bend Series, or simply the Lower Bend Shale (Girty *in* Paige, 1912; Udden *in* Udden and others, 1916; Moore, 1919).

Along the north side of the Llano uplift east of the town of San Saba (figs. 2 and 3), the Barnett is predominantly a black to olive-gray, very thinly laminated shale that weathers light to dark brown (Kier, 1972). Thin brown microsparite limestone beds, brachiopod and cephalopod coquinas, and large ellipsoidal microsparite concretions as much as 2.75 m in diameter are very common, especially in the upper half of the formation. Locally, the shale is very petroliferous, and freshly broken concretions yield a strong petroliferous odor.

On the northeast side of the Llano uplift, the upper 10–150 cm of the Barnett Formation is commonly a dusky to dark yellowish-brown, fine- to coarse-grained, poorly sorted, packed pelletiferous biomicrite and oomicrite. Cephalopods, brachiopods, conodonts, and ostracodes are abundant; glauconite is present in varying amounts, and most of the allochems are phosphatic. Thickness of the Barnett east of San Saba is 10.6–15.2 m.

West of the town of San Saba, the Barnett is divisible into two parts (Freeman, 1962; Turner, 1970). The lower part of the Barnett is a light-colored clay shale. Concretions are small, and phosphatic limestone is abundant at the top of the lower part of the formation. Farther west toward Brady (fig. 2), thin limestone beds are abundant in the lower part of the Barnett, and the section is progressively more phosphatic and glauconitic. The upper part of the Barnett Shale is grayish-black to yellowish-brown, fine- to coarse-grained, packed, glauconitic and phosphatic biomicrite and micrite. Micrite becomes predominant westward. Thickness of the Barnett west of San Saba is 7.6 m.

On the east side of the Llano uplift near the town of Marble Falls (fig. 2), the Barnett is light-colored shale containing concretions. The formation is capped by as much as 3.3 m of phosphatic limestone (Namy, 1969). Maximum thickness of the Barnett on the east side of the uplift is 6.4 m. On the west side of the Llano uplift near Mason (fig. 2), shale mapped as Barnett Formation (Winston, 1963) is probably Marble Falls (W.C. Bell, oral commun. 1970, *reported in* Kier, 1972).

On the north side of the Llano uplift, Gries (1970) and Schwarz (1975) recognized seven or eight species of ammonoids, four species of nautiloids, and several species of pelecypods and brachiopods. Algae, corals, bryozoans, brachiopods, conodonts, and echinoderm fragments occur in the Barnett on the east and northwest sides of the Llano uplift (Namy, 1969; Turner, 1970). On the basis of cephalopods, Schwarz (1975) assigned a late Osage to Chester age to the Barnett. Using conodonts, Hass (1953), and Defandorf (1960) assigned a late

Osage to early Morrow age to the Barnett, spanning the Mississippian-Pennsylvanian boundary.

The placement and nature of the upper contact of the Barnett Formation have been the subject of dispute. Most recent workers place the Barnett-Marble Falls contact between shale or phosphatic limestone and nonphosphatic limestone. Earlier workers did not include the phosphatic limestone in the Barnett (see Kier, 1972) and believed that the Barnett-Marble Falls contact is unconformable because (1) it generally coincides with the Mississippian-Pennsylvanian boundary; (2) in places the Barnett is absent, and Marble Falls limestone beds lie directly on Ellenburger limestone or dolomite beds; and (3) glauconite and phosphate are commonly concentrated at the contact. Zachry (1969), Turner (1970), and Kier (1972), however, noted gradational or interbedded Barnett and Marble Falls rock types at the contact. Only Namy (1969) presented good physical evidence for an unconformity between the Barnett and the Marble Falls. Nevertheless, on the basis of conodonts, Liner and others (1977, in press) inferred that the Barnett-Marble Falls contact on the northeast side of the Llano uplift represents a hiatus from middle Chester to Morrow. To the west, however, they found that the Mississippian-Pennsylvanian boundary and the inferred hiatus are within the Barnett Shale, below the sequence of phosphatic limestone.

The Barnett Formation probably accumulated in a sediment-starved basin under euxinic conditions. Evidence includes the lithic character and general absence of benthonic fossils, particularly an infauna, and the inferred length of time represented by the thin unit. The Barnett of the Llano uplift was probably deposited within an extension of the early, sediment-starved Fort Worth basin (Brown, 1973a). Maximum water depth was undoubtedly below wave base, but still relatively shallow.

Thin microsparite and coquina layers in the Barnett Formation probably reflect temporary cessation of euxinic conditions. Phosphatic beds at the top of the Barnett may record a gradual change from euxinic restricted conditions to open-marine shelf and platform environments characteristic of Marble Falls deposition (Kier, 1972). Yellowish-brown shale and limestone beds exposed on the eastern and northwestern sides of the Llano uplift were deposited under less reducing conditions at the basin margins. Where Barnett shale is thin or missing between the Marble Falls and the Ellenburger, it suggests lack of deposition rather than post-Mississippian erosion.

PENNSYLVANIAN SYSTEM

MARBLE FALLS FORMATION

The Marble Falls Formation records reestablishment of normal marine conditions and widespread limestone environments over the Llano uplift and the adjacent Concho platform. Platform, open-shelf, and shelf-edge carbonate deposition dominated the western margin of the rapidly subsiding Fort Worth basin. These carbonate environments ultimately shifted westward as they were progressively displaced by advancing Smithwick and Strawn deltaic environments.

Marble Falls was introduced by Hill (1889) for "Encrinoidal" limestone exposed along the Colorado River near the town of Marble Falls in Burnet County (fig. 2). Later, Hill (1901) concluded that the Marble Falls is correlative with limestone of the "Bend division" exposed along the north side of the Llano uplift, although the two outcrop areas are not physically connected. Various names and stratigraphic ranks have been proposed for the Marble Falls Formation at the surface and in the subsurface of central, north-central, and west-central Texas; Comyn, Big Saline (Cheney, 1940) and Sloan (Plummer, 1945, 1947a; see Kier, 1972). Most investigators have retained the name Marble Falls for surface exposures on and around the Llano uplift.

Except on the east side of the Llano uplift near the town of Marble Falls, the Marble Falls Formation can be subdivided into two outcropping units separated by an unconformity (Freeman, 1962; Namy, 1969; Zachry, 1969; Turner, 1970; Kier, 1972). The lower part of the Marble Falls is Morrow in age throughout its outcrop. The upper part of the formation, however, becomes progressively younger westward. On the east and northeast sides of the Llano uplift the upper Marble Falls is Morrow in age (Namy, 1969; Zachry, 1969; Kier, 1972); just west of the town of San Saba (fig. 2), on the north side of the uplift, the formation is Morrow and Atoka in age (Turner, 1970); and near Brady on the northwest side of the uplift, the formation is entirely of Atoka age (Freeman, 1962). The upper and lower parts of the Marble Falls accumulated under different depositional conditions. The subsurface extent of the unconformity within the Marble Falls is uncertain.

The lower part of the Marble Falls consists predominantly of light to dark cherty limestone and thin shale beds. Principal limestone types include algal biomicrite and biosparite, oosparite, spiculitic biomicrite, pelmicrite, micrite, and mixed skeletal fragment biomicrite and biosparite. Locally, coral and algal biolithite are found. Diagenetic alteration of limestones within the Marble Falls is limited generally to recrystallization of micrite to microspar or pseudospar and to inversion of aragonite to calcite.

Facies patterns in the lower part of the Marble Falls Formation are complex, and there is considerable local variation within individual exposures. Nevertheless, facies appear to have been arranged originally in a semicircular pattern around the Llano uplift. The structurally positive uplift apparently acted as the core for a major carbonate platform. Vertical accretion dominated over lateral accretion of carbonate facies, and high-energy facies tended to expand in areal extent at the expense of lower energy facies. Depositional relief was as much as 9 m and was a significant controlling factor in determining facies characteristics. Facies patterns at or near the end of deposition of the lower part of the Marble Falls demonstrate net regression. The lower Marble Falls is about 30 m thick but ranges in thickness from about 21 to 45 m. Near Mason, much of the lower Marble Falls Formation was apparently eroded prior to deposition of the upper part of the Marble Falls (W. C. Bell, oral commun., 1971).

The upper part of the Marble Falls Formation is predominantly light to dark algal biomicrite, calcarenite, siliceous spiculitic biomicrite, and shale. Although lithic types are similar to those in the lower part of the Formation, the depositional setting was distinctly different in several respects:

1. Facies patterns are oriented approximately north-south; a distinct semicircular carbonate platform on the Llano uplift cannot be recognized.
2. Shale and spiculitic limestones were deposited over the entire Llano uplift, in contrast to patterns in the lower part of the Marble Falls, where deposition of shale and spiculitic limestones was mostly restricted to off-platform environments.
3. Individual facies are thin and widespread; depositional relief was relatively low.
4. Although high- and low-energy facies were widespread, shifts in facies boundaries are common.
5. High-energy facies become more common upward within the formation, but they are not as common as they are within the lower part of the Marble Falls.
6. Facies patterns in the upper part of the Marble Falls record westward transgression.

The upper part of the Marble Falls Formation is 36–67 m thick in outcrop; average outcrop thickness of the entire Marble Falls Formation is 91 m.

Marble Falls deposition began with establishment of more open, less restrictive conditions than existed during deposition of the Barnett Shale. Incipient calcarenite shoals developed, apparently at some slight break in slope. A major carbonate platform centered about the Llano uplift rapidly formed over the shoals. Platform margins were approximately coincident with present Marble Falls outcrops. On the northeast side of the platform, depositional environments resembled the modern Bahamian platform, although platform/off-platform relief was not nearly so great (Kier and Zachry, 1973; Zachry and Kier, 1973). Ultimately the lower Marble Falls platform either built to sea level or was exposed by a drop in sea level.

The upper part of the Marble Falls Formation was deposited predominantly by algal buildups and calcarenite shoals and by shale and spiculitic biomicrite deposited in somewhat restricted depressions on the open-marine shelf marginal to the Fort Worth basin. When the lower Marble Falls carbonate platform was subaereally exposed, shale and spiculitic biomicrite continued to be deposited on the off-platform shelf between the Llano uplift and the Fort Worth basin (Namy, 1969). Subsidence of the eastern edge of the lower Marble Falls platform allowed shale and spiculitic limestone to onlap the erosion surface, followed by establishment of algal buildups and calcarenite shoals. Marine energy levels and depositional relief were less than during deposition of the lower Marble Falls.

Upper Marble Falls depocenters shifted progressively westward as the lower Marble Falls platform subsided. Lengthy erosion of the platform on the west side of the Llano uplift removed much, and locally all, of the lower Marble Falls Formation (Freeman, 1962) prior to deposition of the upper Marble Falls. Strawn deltas simultaneously prograded across the upper Marble Falls carbonate shelf from the east. The upper contact of the Marble Falls Formation with the overlying Smithwick Shale or Strawn sandstones is essentially conformable.

SMITHWICK FORMATION

The Smithwick Formation represents the initial terrigenous clastic deposits of Atoka and Strawn deltas which prograded across the Fort Worth basin and onto the Llano uplift. Regional subsurface correlations from Hill County to Brown County and from Dallas County to Stephens County (fig. 2) demonstrate the time-transgressive relationships of the Atoka and Strawn Groups and the Smithwick Formation. In outcrop, the Smithwick is the prodelta facies of Atoka and Strawn delta systems. Atoka facies are restricted to the Fort Worth basin; Strawn facies are found within the basin, but they also crop out around the Llano uplift and within the Colorado and Brazos River valleys.

The Smithwick Formation was named by Paige (1911) for shale and sandstone exposed near Old Smithwick in Burnet County. Girty (*in* Paige, 1912) inferred that the Smithwick is equivalent to the upper shale of the "Bend Series" north of the Llano uplift. Although several attempts have been made to restrict the Smithwick (Cheney and others, 1945; Plummer, 1950), application of the name Smithwick to shale overlying the Marble Falls in outcrop and in the subsurface of north and west-central Texas is generally accepted.

The Smithwick Formation consists of black, slightly calcareous, fissile clay-shale and lesser amounts of siltstone and sandstone. Minor amounts of dark limestone and conglomerate are composed of limestone, chert, and sandstone clasts. Sedimentary bed forms—ripple marks, flute casts, groove casts, and slump features—are common on sandstone bedding surfaces, particularly on the east side of the Llano uplift (McBride and Kimberly, 1963). Locally, hematitic concretions are found in the Smithwick.

Although macrofossils are rare in the Smithwick, they are concentrated in a few localities. Gries (1970) identified several species of rugose coral, particularly *Cumminsia aplata*, brachiopods, gastropods, pelecypods, and cephalopods. Turner (1970) observed spicules, arenaceous foraminifers, and plant fragments. The Smithwick is as much as 30 m thick or more on the north side of the Llano uplift (Kier, 1972) and as much as 121 m thick on the east side (McBride and Kimberly, 1963). In general, the Smithwick Formation thins westward; locally it is absent.

In outcrop, the contact between the Smithwick Formation and the Strawn Group is gradational. Upward within the Smithwick, the amount of sandstone increases, and shales become siltier (Kier, 1972). A faunal change that ostensibly represents a significant hiatus between the Smithwick Formation and the Strawn Group (Plummer, 1947a) is undocumented and may simply reflect environmental differences (Kier, 1972). Variations in thickness of Smithwick in outcrop probably reflect original depositional variations, differential compaction, and the effects of contemporaneous faulting. Differences in trends of Marble Falls-Smithwick outcrops and Strawn outcrops are due to contemporaneous structural relief on the Llano uplift, to postdepositional uplift and doming of the southern exposure of three contemporaneous yet time-transgressive facies, and perhaps to westward tilting of the Eastern shelf toward the Fort Worth basin (Kier, 1972).

Smithwick Shale exposed on the north side of the Llano uplift is the prodelta facies of Strawn deltas (Turner, 1970; Kier, 1972). McBride and Kimberly (1963) interpreted a "deep water" environment, periodically invaded by turbidity currents to explain the Smithwick east of the Llano uplift, but McBride (oral commun., 1977) no longer believes the water was particularly deep. Conglomerate in the Smithwick Formation on the east and northwestern sides of the Llano uplift is related to contemporaneous faulting and erosion of the Marble Falls Limestone (Freeman, 1962; Freeman and Wilde, 1964; McBride and Kimberly, 1963).

ATOKA GROUP

Very thick clastic strata were deposited within the Ouachita geosyncline during Late Mississippian and Early Pennsylvanian time (Stanley, Jackfork, and Johns Valley Formations). Near the end of the Morrow series, rejuvenation of structural activity in the Ouachita foldbelt provided an enormous volume of clastic sediment that was transported and deposited within the geosyncline and adjacent Fort Worth basin. The depocenter shifted westward during Atoka and Strawn deposition.

A clastic wedge nearly 2,000 m thick makes up the Atoka Group, which is restricted to the subsurface of the basin. Deposited by westward-prograding fan deltas, the group consists of thick shale and sandstone facies that probably range from shallow marine-deltaic to deeper basin facies. Westward, the Atoka is, in part, gradational with the older parts of the Smithwick Formation within the Fort Worth basin. Atokan clastic rocks may also be partly time-equivalent to upper Marble Falls facies on the Concho platform and Llano uplift. Because the group is restricted to the subsurface, its description is limited in this report.

STRAWN GROUP

The Strawn Group consists predominantly of cyclic terrigenous clastic facies deposited by Middle Pennsylvanian fluvial-deltaic systems that essentially filled and prograded westward across the Fort Worth basin, onto the Concho platform and into the incipient Midland basin. In the Fort Worth basin, the Strawn Group is gradational with the underlying Atoka Group. Only the upper part of the Strawn Group crops out, although Strawn facies on the Llano uplift may be equivalent to lower Strawn in the Fort Worth basin. Regional upwarping of the Ouachita foldbelt and the eastern margin of the Fort Worth basin continued to supply large volumes of eroded Atokan sediments to the Strawn rivers. Fluvial and deltaic deposition dominated the Concho platform and extended onto the Llano uplift. Near the end of Strawn deposition, erosion lowered the source area, and the supply of sediments diminished.

The Strawn Group crops out in both the Colorado River valley and Brazos River valley in central and north-central Texas, respectively (figs. 1, 2, 5, 6). In the Brazos River valley, Strawn rocks have been studied more extensively because of exploration for coal, oil, and gas; exposures are better, persistent limestone beds facilitate subdivision, and abundant marine faunas in the shale sections permitted time-stratigraphic interpretations. Five Strawn formations are recognized in the Brazos River valley (fig. 4) by Brown and Goodson (1972). Few formal subdivisions are recognized in the Colorado River

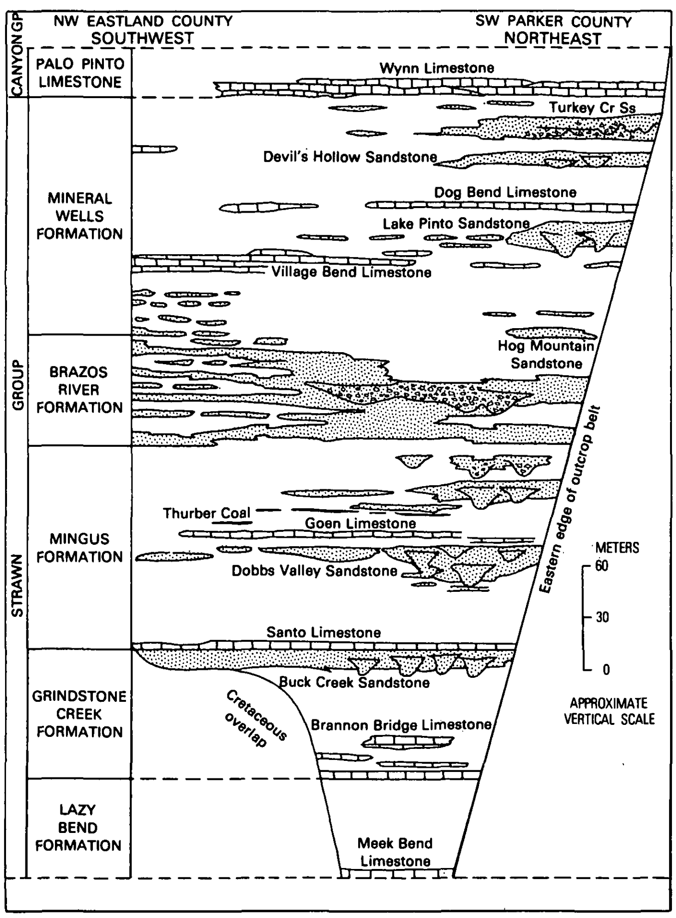

FIGURE 5.—Schematic cross section of outcropping Strawn Group in the Brazos River valley. Modified from Brown and Goodson (1972) and Cleaves (1975).

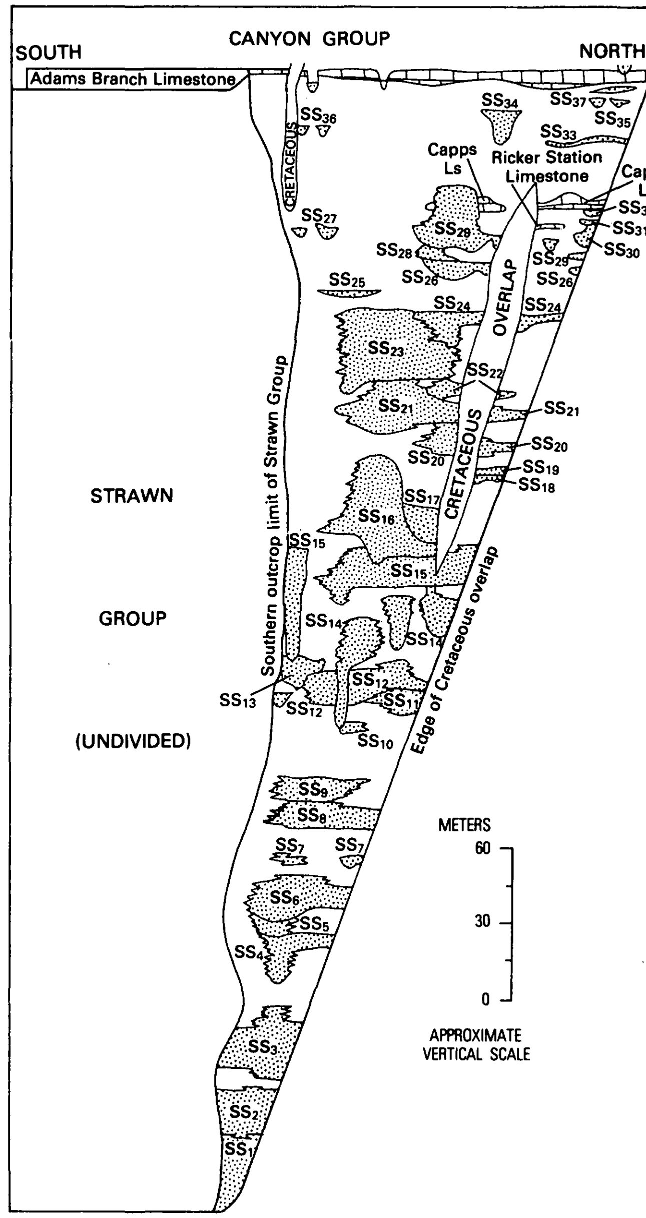

FIGURE 6.—Schematic cross section of outcropping Strawn Group in the Colorado River valley. Modified from Kier and others (1976). Unnamed sandstone units are numbered in approximate stratigraphic order from the base.

valley, and most lithic units have simply been numbered (Drake, 1893; Kier and others, 1976).

Dumble (1890) first applied the name "Strawn series" to coal- and limestone-bearing clay and shale exposed near the town of Strawn in Palo Pinto County (fig. 2). Equivalent strata in the Colorado River valley were apparently called "Milburn series," although in the same report, Tarr (1890) proposed the names "Richland Sandstone" and "Milburn Shales" for Strawn strata in the Colorado River valley. Cummins (1891) defined the "Strawn division" to include strata from the base of Canyon limestones to the base of Coal Seam #1 (Thurber Coal). Strata below Coal Seam #1 to the top of the underlying "Bend division" (Smithwick Formation) were called the "Milsap Division." The Milsap was not recognized in the Colorado River valley, however, where Cummins presumed that the middle part of the Strawn lay unconformably on the Bend. In the Colorado River valley, the top of the Strawn was placed at the base of the limestone (Capps) exposed near Brownwood (figs. 2, 4, 6).

Drake (1893) applied the name Strawn to all strata between the Bend and Canyon "divisions." In the course of his work, Drake mapped and numbered 20 Strawn "beds" cropping out in the Colorado River valley; this map was the only detailed map and subdivision of Strawn rocks in this area prior to publication of the Brownwood Sheet of the Geologic Atlas of Texas (Kier and others, 1976).

Application of the name Strawn underwent considerable evolution following Drake's (1893) work, first as a formation (Smith, 1903; Udden *in* Udden and others, 1916), then as a group (Plummer and Moore, 1921; Plummer, 1929; Scott and Armstrong, 1932; Plummer and Hornberger, 1935), and finally as a series (Cheney, 1940, 1947; Cheney and others, 1945; Spivey and Roberts, 1946; Quigley and Schweers, 1951).

In the 1950's, principal studies on the Strawn Group were by students working under S. P. Ellison at The University of Texas at Austin (Abilene Geological Society, 1954) and Leo Hendricks (1957) at Texas Christian University. The Abilene sheet of the Geologic Atlas of Texas (Brown and Goodson, 1972) used essentially the same nomenclature (figs. 4, 5, 6) as Plummer and Hornberger (1935). Cleaves (1973, 1975) studied the upper part of the Strawn and extended the lithic units mapped by Brown and Goodson (1972) westward into the subsurface. The lower part of the Strawn is not exposed in central or north-central Texas.

The Strawn Group consists predominantly of shale and sandstone and lesser amounts of limestone, coal, and conglomerate. For the group in the Brazos River valley, Cleaves (1973, 1975) made the following interpretations:

1. Shale and sandstone were deposited in delta, prodelta, and embayment environments.
2. Limestone was deposited in open-shelf environments or locally in interdeltaic-bay environments. Open-shelf carbonate rocks, mostly algal biomicrites, are regionally extensive in outcrop and in the subsurface, and they serve as marker beds for delineating formations within the Strawn Group.
3. Coal formed predominantly in marshes and swamps on delta-plain and along interdeltaic-embayment coastlines. Coal crops out principally in the Brazos River valley area, and only the Thurber coal of southern Palo Pinto County (figs. 2, 5) has been successfully mined.
4. Source areas for Strawn terrigenous clastic deposits were the Ouachita Mountains, the eastern part of the Fort Worth basin, and the Arbuckle Mountains of Oklahoma (fig. 3). Locally, the Wichita Mountains and the Criner Hills in southern Oklahoma supplied sediment to Strawn deltas and fan deltas.

Fluvial, deltaic, and related facies recognized in the Brazos River valley by Cleaves (1975) are presented in table 2.

Strawn rock types and facies exposed in the Colorado River valley differ from Strawn deposits in the Brazos River valley in several respects (figs. 5, 6). There is considerably more sandstone, locally conglomeratic, in the lower part of the outcropping Strawn Group in the Colorado Valley. Very little coal, none of it economic, is found in the Colorado Valley. Only two mappable limestone beds (Capps Limestone and Ricker Station Limestone) crop out within the Colorado Valley.

Kier (1972) summarized inferred depositional environments of the Strawn Group in the Colorado River valley and concluded that much of the outcropping Strawn is of fluvial or fluvial-deltaic origin. Source area for the terrigenous clastic sediment was to the east in the Ouachita Mountains. Conglomerate, composed of subangular to rounded pebbles, cobbles, and boulders of eroded Marble Falls Limestone, is found at or near the base of the Strawn Group in San Saba County (Kier, 1972)

TABLE 2.—*Fluvial and deltaic facies in Strawn Group, Brazos River valley*

[Data from Cleaves (1975)]

Deltaic facies	
Destructional	Distinguished on basis of stratigraphic position below open-shelf mudstones and limestone beds. Lenticular siltstone and silty sandstone interbedded with dark-gray to black bituminous mudstone; commonly burrowed, local long-crested symmetrical ripples; plant debris common; 2–30 cm thick.
Delta plain	Bituminous mudstone and siltstone; numerous root traces and tree stumps. Includes thin splay sandstone with trough crossbeds and meandering stream channel.
Distributary channel fill.	Fine- to very fine grained sandstone; sharp erosional base, abrupt gradational upper contact, lower beds commonly contoured where underlying mud is thick, no well-developed fining-upwards sequence, although base may be coarser grained than top, large-scale trough crossbeds in base grading upwards to small-scale trough beds and climbing ripple cross stratification; clay galls common near base; base of channel may contain abundant plant debris.
Interdistributary bay	Variable, distinguished on basis of stratigraphic position between channel-mouth bar and overlying channel sandstones. Commonly unlaminated brown mudstone with abundant ironstone nodules and thin muddy detrital coal zones; faunal content may include worthinid gastropods, pectinid and nuculanid bivalves, chonetid and spiriferid brachiopods, and crinoids; 0.3–3 m thick.
Channel-mouth bar	Massive beds of well-sorted fine- to very fine grained sandstone; plane beds and low-angle large-scale trough crossbeds dominant, high-angle trough-fill crossbeds common, soft-sediment deformation, particularly lower beds, growth faults occur; macrofossils rare, small plant fragments common; 6 m thick.
Proximal delta front	Thin to massive beds of well-sorted, fine- to very fine grained sandstone; oscillation ripple cross-stratification, plane beds, small- to medium-scale trough-fill crossbeds dominant, growth faults occur; macrofossils rare, small plant fragments common; 3–20 m thick (total delta front).
Marginal delta front	Massive, blocky beds of sandy, coarse siltstone and muddy very fine grained sandstone; extensively bioturbated and burrowed. Marine reworked sands transported along strike from channel-mouth bars and proximal delta front.

TABLE 2.—*Fluvial and deltaic facies in Strawn Group, Brazos River valley*—Continued

Deltaic facies—Continued	
Distal delta front	Thin beds of coarse-grained siltstone and fine sandstone interbedded with mudstone; graded beds, flow rolls, long-crested oscillation ripple cross-stratification, and small-scale trough crossbedding dominant; fossils and bioturbation rare; 1–10 m thick. Deposited in part by turbidity currents.
Prodelta	Dark-gray, brown, or black mudstone; fossils rare except in distal parts; 1.5–61 m thick.
Fluvial facies	
Confined valley fill	Fining-upward sequence; basal part composed of pebbly conglomerate with large-scale trough crossbeds grading upward to medium- to small-scale trough crossbeds, tabular crossbeds, and parallel bedding; found in shallow superimposed valleys commonly cut into delta-plain facies.
Fine-grained meanderbelt.	Fining-upward sequence; basal 1 m contains large-scale trough crossbedded chert-pebble conglomerate; overlain by small-scale trough crossbeds; upper part contains parallel-bedded silty clay partings and thin oscillation-rippled very fine grained sandstone.
Interdeltaic embayment facies	
Sheet sandstone	Thin-bedded siltstone and fine- to medium-grained sandstone; long-crested symmetrical ripples dominate, small- to medium-scale, low-angle crossbeds in lenses; burrows very common; macrofossils rare; thickness as much as 6 m. Derived from marine reworking of adjacent deltaic sediments, deposited in strand-plain and shoreface environments.
Mudstone	Massive, dark-gray to brown mudstone; thin discontinuous lenses of burrowed sandstone and siltstone; iron oxide and septarian nodules present; spiriferid brachiopods.
Coal and bituminous claystone.	Thurber coal: 0.3–1 m thick, jarosite partings; kaolinitic underclay with lycopod stigmaria and charcoal fragments; burrowed, silty sandstone below underclay and above sandstone. Bituminous claystone: platy, fissile, lenticular; contain finely divided unlaminated reedy plant debris; may or may not contain root traces.
Bayhead deltas	Similar to distributary channel-fill deposits, but small, and cut into or overlying bay facies.

and probably accumulated as fan and fan-delta deposits shed from local fault blocks.

In the Colorado River valley, the base of the Strawn Group is in contact either with the Smithwick or the Marble Falls Formation. The Strawn-Smithwick contact is conformable and gradational. Regionally, the Strawn-Marble Falls contact also is conformable (Kier, 1972). Principal evidence of conformity is the gradation from intraclastic Marble Falls limestones into Strawn sandstones or limestone conglomerates, which in turn grade laterally and vertically into Strawn sandstones. Paleontological evidence summarized by Turner (1957) also indicates a conformable contact. Observed unconformable relationships (relief on the Marble Falls (Turner, 1970) and channeling (Freeman, 1962)) are local and resulted primarily from erosion of Strawn delta distributaries into subjacent Marble Falls limestones. The Strawn Group pinches out between the Marble Falls Formation and Canyon limestones along the north side of the Llano uplift. Pinchout is apparently the result of nondeposition and perhaps local erosion associated with post-Marble Falls and pre-Canyon faulting (Cheney, 1940; Freeman, 1962; Freeman and Wilde, 1964).

CANYON GROUP

The Canyon Group is a sequence of Late Pennsylvanian (Missouri) carbonate and terrigenous clastic rocks that crop out in the Colorado, Brazos, and Trinity River valleys and are within the subsurface of west Texas. Reduction in terrigenous clastic sediment supplied from the eroded Ouachita Mountains and deposition of widespread open-marine limestone on the Eastern shelf (Concho platform) distinguish the Canyon Group from the underlying Strawn and overlying Cisco Groups. Individual limestone units as much as 26 m thick cap a series of prominent, east-facing cuestas in the Canyon outcrop belt. Steep cuesta slopes and grassy valleys are underlain by deltaic and marine shales. Lenticular deltaic sandstone bodies within the shales crop out in hummocky, post oak-covered ridges. Schematic cross sections of the Canyon Group are shown in figure 7.

The Canyon Group was named by Cummins (1891) for massive limestone and interbedded shale exposed near Canyon, a Texas and Pacific Railroad station 6.4 km west of Strawn in Palo Pinto County (fig. 2). In the Colorado River valley, Drake (1893) described and named many stratigraphic units within the Canyon Group. Plummer (1919) and Plummer and Moore (1921) named stratigraphic units in the Brazos River valley, and Scott and Armstrong (1932) named units in the Trinity River valley. Other early investigators include Dobbin

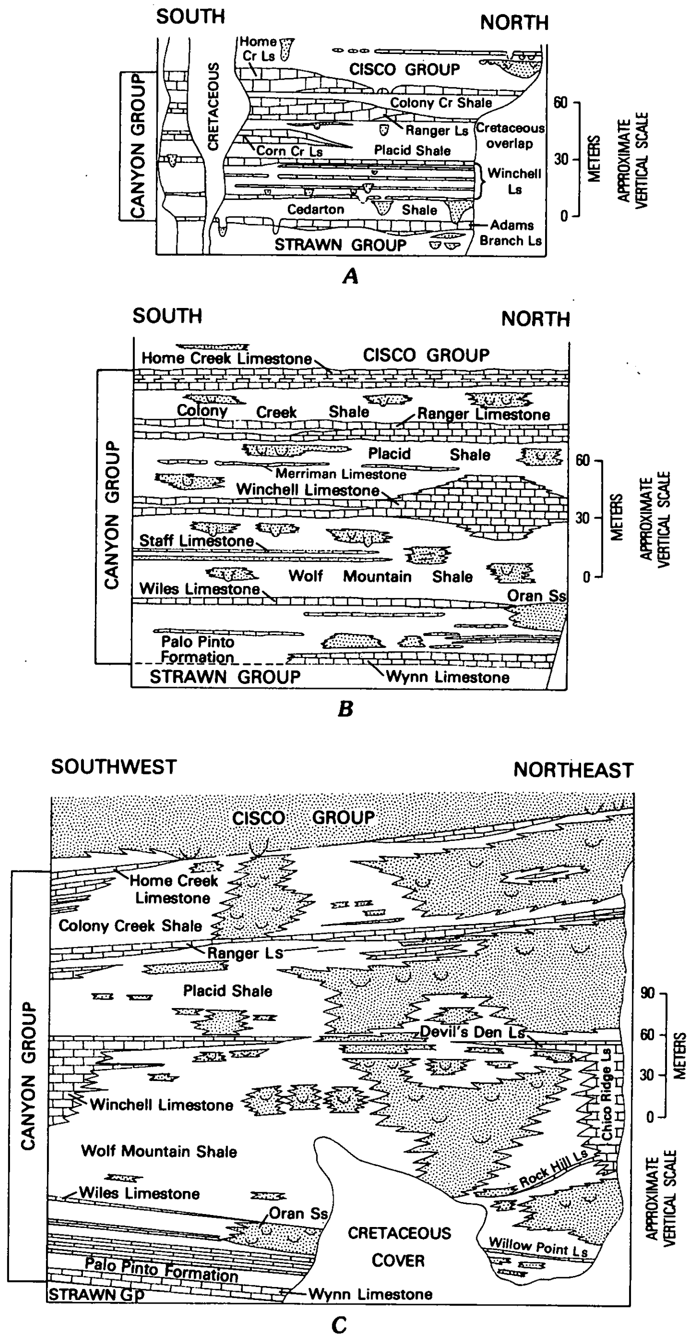

FIGURE 7.—Schematic cross sections of outcropping Canyon Group, Colorado, Brazos, and Trinity River valleys. *A*, Colorado River valley, McCulloch, Coleman, and Brown Counties (modified from Kier and others, 1976). *B*, Brazos River valley, Eastland, Stephens, and Palo Pinto Counties (modified from Brown and Goodson, 1972). *C*, Brazos and Trinity River valleys, Palo Pinto, Jack, and Wise Counties (modified from Erxleben, 1975).

(1922), Reeves (1922), Plummer and Hornberger (1935), Bradish (1937), and Lee (1938).

Cheney (1940, 1947, 1949), who redefined the stratigraphic classification of Pennsylvanian and Permian strata of Texas, correlated Canyon rocks with the Missouri Series of the midcontinent. Lithofacies studies were carried out by Wermund (1966, 1969, 1975), Wermund and Jenkins (1964, 1969, 1970), Roepke (1970), and Erxleben (1973, 1975). Brown and Goodson (1972) and Kier and others (1976) mapped Canyon outcrops in the Brazos and Colorado River valleys, respectively. Other reports on the Canyon Group include: Abilene Geological Society (1954), North Texas Geological Society (1940, 1956, 1958), West Texas Geological Society (1951), Cheney and Eargle (1951), Jenkins (1952), Feray and Jenkins (1953), Shelton (1958), Eargle (1960), Terriere (1960), Laury (1962), Perkins (1964), Raish (1964), Bretsky (1966), Brooks and Bretsky (1966), Feray and Brooks (1966), Pollard (1970), and Heuer (1973). Wermund (1966) and Erxleben (1975) summarized previous investigations of the Canyon Group.

As presently defined (Brown and Goodson, 1972; Erxleben, 1973, 1975; Kier and others, 1976), the Canyon Group comprises seven formations (figs. 4, 7). In the Brazos River valley, the contact between the outcropping Canyon Group and the underlying Strawn Group is placed at the base of the Wynn Limestone Member, lowest limestone in the Palo Pinto Formation (Brown and Goodson, 1972). The Palo Pinto Formation (and Wynn Limestone Member) pinches out southward and is absent in the Colorado River valley outcrop area. In the Colorado River valley, the base of the Adams Branch Limestone defines the base of the Canyon Group (Kier and others, 1976). Although several other Strawn-Canyon contacts have been used in the past (see Shelton, 1958; Laury, 1962; and Roepke, 1970 for summaries), the base of the Palo Pinto Formation and the base of the Adams Branch Limestone best separate predominantly marine limestone and shale deposits (Canyon) and predominantly terrigenous clastic deposits (Strawn). The Adams Branch Limestone correlates with the Staff Limestone in the Brazos River valley (Cheney, 1929). Consequently, the base of the Canyon Group in the Colorado River valley is younger than the base of the Canyon in the Brazos River valley.

Drake (1893) placed the top of the Canyon Group at the top of the "*Campophyllum* bed" (Gunsight). Since the work of Plummer (1919) and Plummer and Moore (1921), however, the top of the Canyon has been recognized at the top of the Home Creek Limestone (figs. 4, 7), the uppermost thick limestone unit in the Pennsylvanian outcrop belt.

Shale and sandstone were deposited in terrigenous clastic delta, fan-delta, and shelf environments; limestone was deposited in carbonate shelf, bank, reef, and platform environments. Major influx of Canyon terrigenous clastic sediments into north-central Texas was concentrated in a high constructive delta system that crops out at the north end of the Canyon outcrop belt (Erxleben, 1973, 1975) in Jack and Wise Counties (figs. 2, 8). Thick shale and sandstone beds in the Wolf Mountain, Placid, and Colony Creek Formations (figs. 4, 7) were deposited in lobate and elongate deltas that prograded from Ouachita Mountains westward and northwestward across the Eastern shelf. Canyon delta facies resemble those in the Strawn Group (table 2). Additional terrigenous clastic sediments were deposited in north-central Texas by a fan-delta system (subsurface only) that prograded southward from the Arbuckle-Wichita Mountains (Erxleben, 1975). Minor amounts of terrigenous clastic sediments derived from the Ouachita Mountains were deposited in central Texas (Roepke, 1970).

Thick Canyon carbonate facies crop out near Possum Kingdom Reservoir in Palo Pinto County and Lake Bridgeport in Wise County (fig. 2) where the Winchell Limestone and its equivalent, the Devils Den Limestone, are composed of bank facies. Limestone banks are predominantly phylloid algal biomicrites including the genera *Eugonophyllum* and *Archaeolithophyllum* (Wermund, 1966, 1969, 1975), which acted as sediment traps for lime mud. A variety of other organisms lived in association with phylloid algae: encrusting algae; crinoids; fenestrate and encrusting bryozoans; fusulinids; echinoids; local rugose corals of the genera *Lophophyllidium* and *Caninia*, colonial syringoporid corals; sponge genus *Heliospongia;* brachiopods including the genera *Composita, Neospirifer, Echinoconchus,* and *Juresania;* gastropods of the genera *Bellerophon* and *Straparolus;* and pelecypods, including the genera *Aviculopinna, Myalina,* and *Culunana* (Wermund, 1969; Erxleben, 1975). Biosparites are uncommon but are found near the top of the bank deposits (Wermund, 1975). Both biohermal and biostromal banks are found.

Algal-bank facies are commonly found over paleobathymetric highs caused by differential compaction of subjacent deltaic sands or limestone banks and interlobe and interbank muds, respectively. The banks may also be associated with northeast

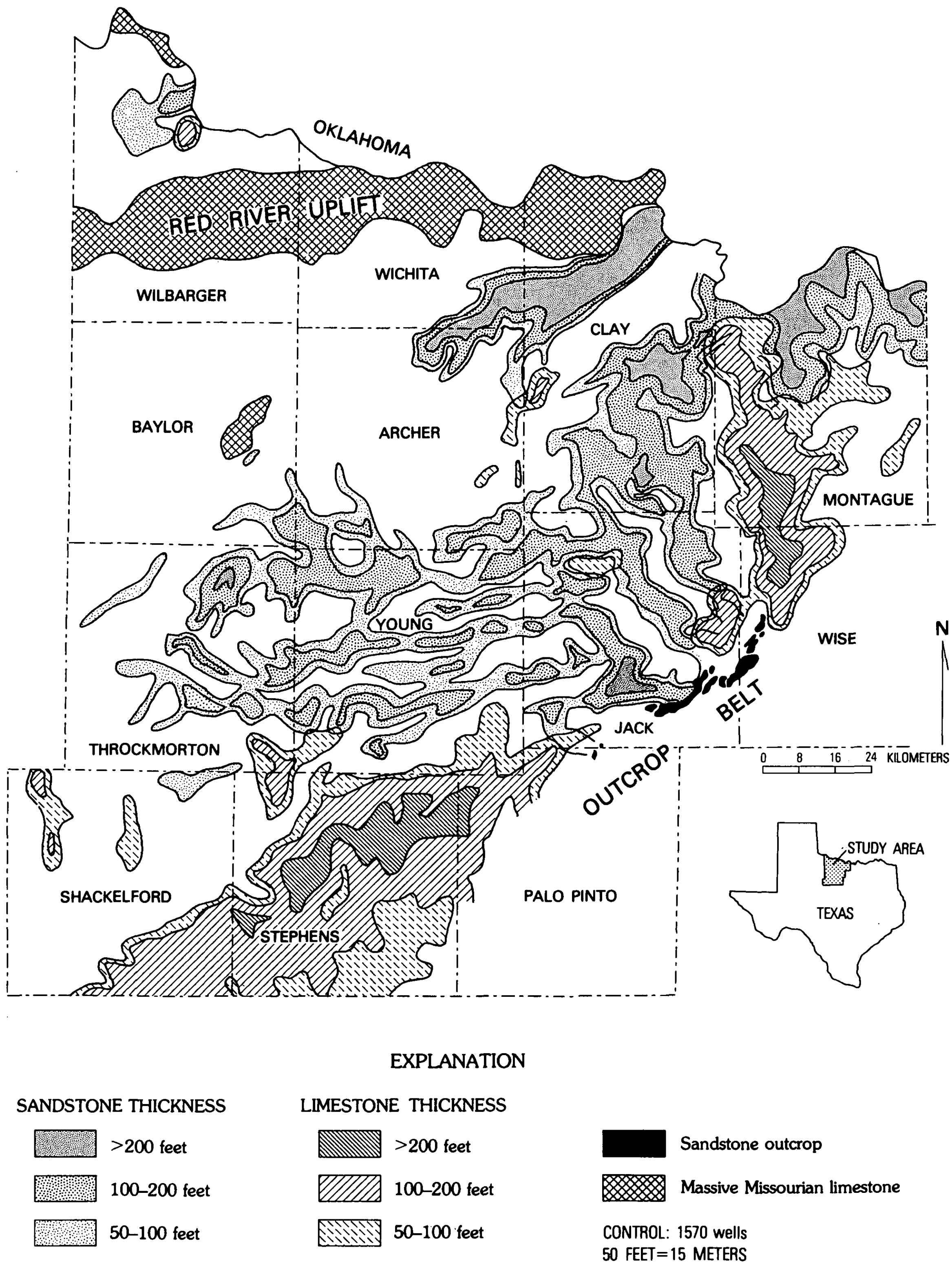

FIGURE 8.—Net thickness of sandstone and limestone of Winchell-Wolf Mountain Formations, north-central Texas. Area without pattern is shale or less than 50 feet of sandstone or limestone. Data on open-file, Texas Bureau of Economic Geology. Modified from Erxleben (1973).

structural trends (Brown, 1969c; Erxleben, 1975). Incipient paleorelief was probably a few centimeters (Wermund, 1966); maximum relief was probably 10 m (Wermund, 1975). Grain size, sorting, and crossbeds suggest limestone deposition above wave base. Analogy between Pennsylvanian phylloid algae and modern *Eudotia* algae suggests deposition in 1–3 m of water (Wermund, 1975).

Regionally extensive shelf-limestone deposits, such as the Palo Pinto, Adams Branch, Winchell, Ranger, and Home Creek Limestones (fig. 4), crop out in north-central Texas. After each major episode of delta progradation, shelf limestones onlapped (transgressed) subsiding delta lobes, providing widespread, relatively continuous marker beds that permit subdivision and correlation of Canyon strata in outcrop and subsurface. Shelf-limestone facies resemble bank facies but are irregularly bedded and contain thin marine shale beds. Individual shelf-limestone units are 1–15 m thick but may be thicker within interdeltaic embayments. An idealized Canyon depositional cycle (progradational deltaic sequence, destructional terrigenous clastic facies, and transgressive shelf limestone) is illustrated in figure 9; evolution of Canyon paleogeography is illustrated by figure 10. Platform and reef carbonates occur only in the subsurface along the Red River uplift and the eastern margin of the Midland basin, respectively (Erxleben, 1975; Wermund, 1975).

The Canyon Group is thickest in the Brazos River valley where outcrops are near the Canyon depocenter, the site of major terrigenous clastic deposition. As much as 545 m of Canyon rocks accumulated in Montague County (Erxleben, 1975; fig. 2); the Canyon Group thins southward to 273 m in Stephens County (fig. 2) and to only 120–135 m in the Colorado River valley. Although individual limestone units and interstratified terrigenous clastic facies thin southward in the Colorado River valley, limestone makes up a greater proportion of the section compared with the Canyon Group in the Brazos River valley. Near the Brady Mountains (fig. 2),

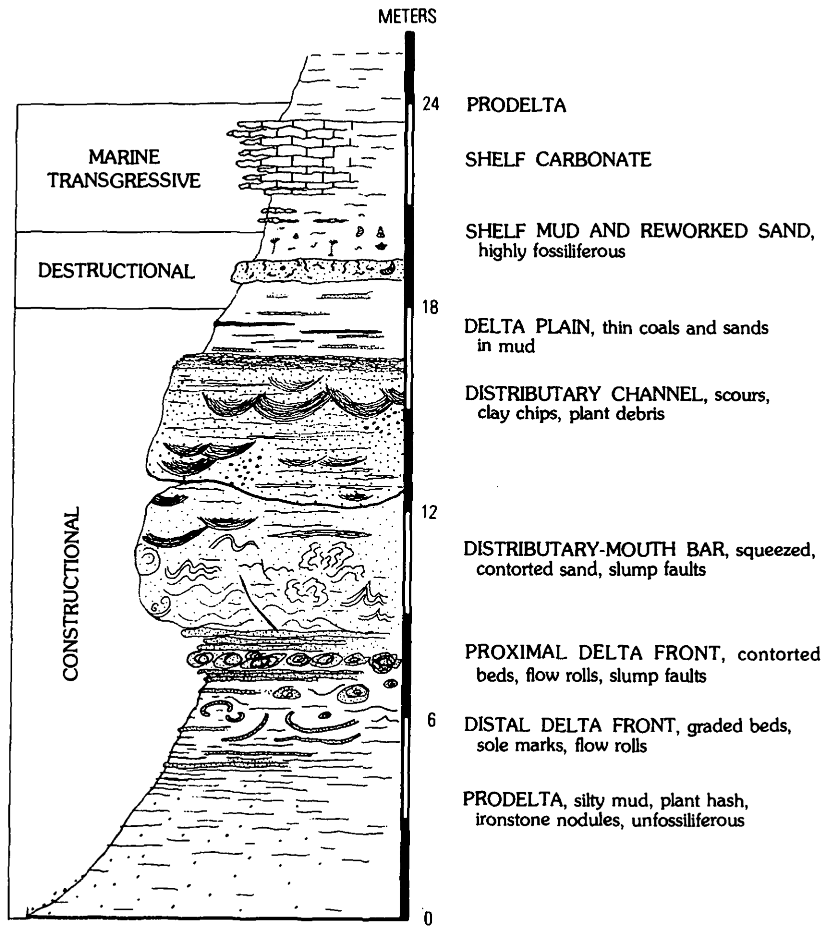

FIGURE 9.—Idealized delta sequence, Canyon Group, north-central Texas. From Erxleben (1975).

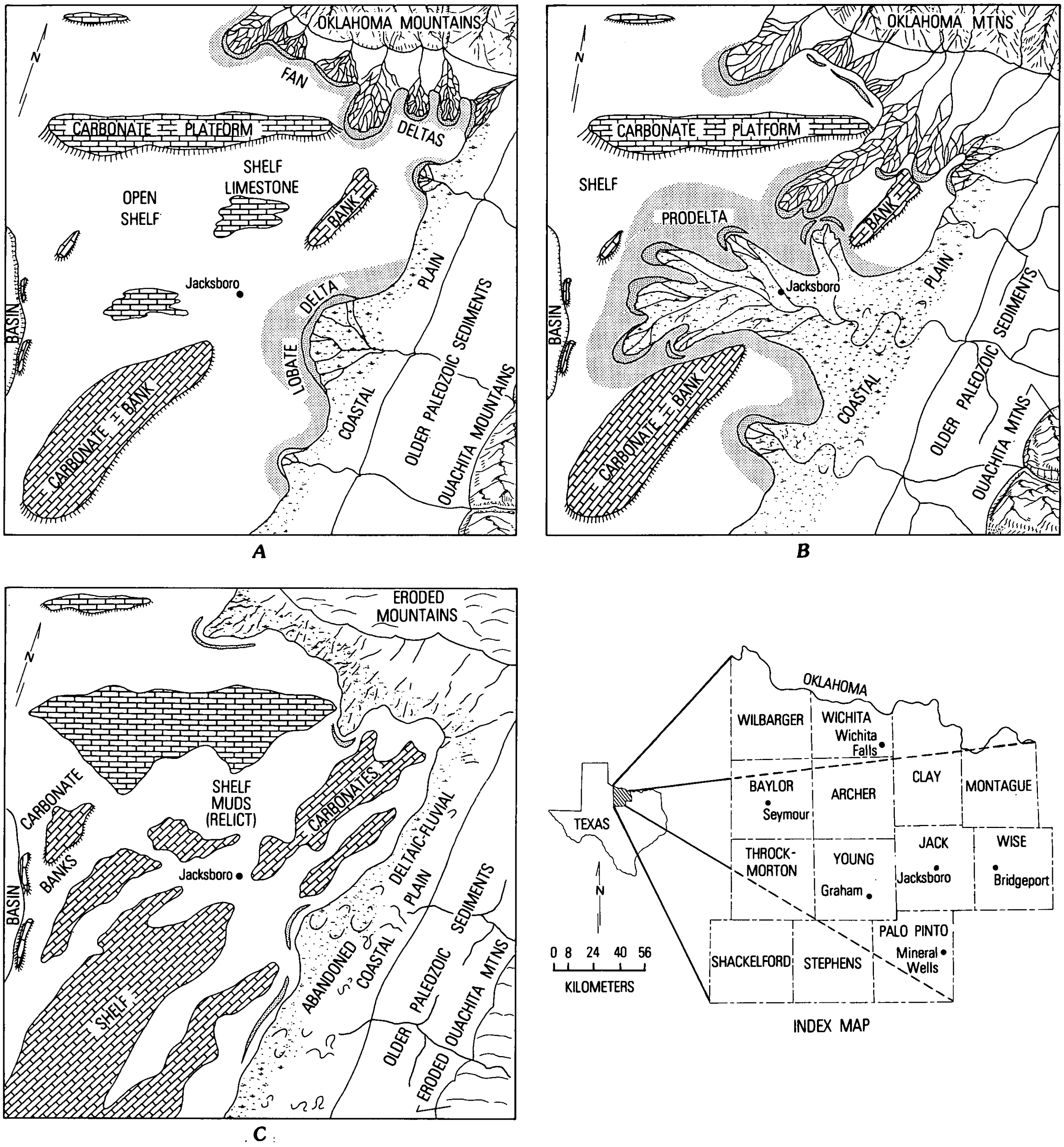

FIGURE 10.—Evolution of Canyon paleogeography in north-central Texas. *A*, Early progradation of delta systems. *B*, Maximum extent of delta development. *C*, Shelf transgression over abandoned deltaic facies. Based on three Canyon delta cycles. From Erxleben (1973).

shale and sandstone are essentially absent, and limestone makes up nearly all the Canyon Group.

CISCO GROUP

The Cisco Group, as originally defined by Cummins (1891), is a sequence of terrigenous clastic and carbonate facies that record rejuvenated uplift of the Ouachita foldbelt and Fort Worth basin. The increased sediment supply initiated extensive delta progradation across the Eastern shelf. Thick sandstone and conglomerate deposits distinguish the

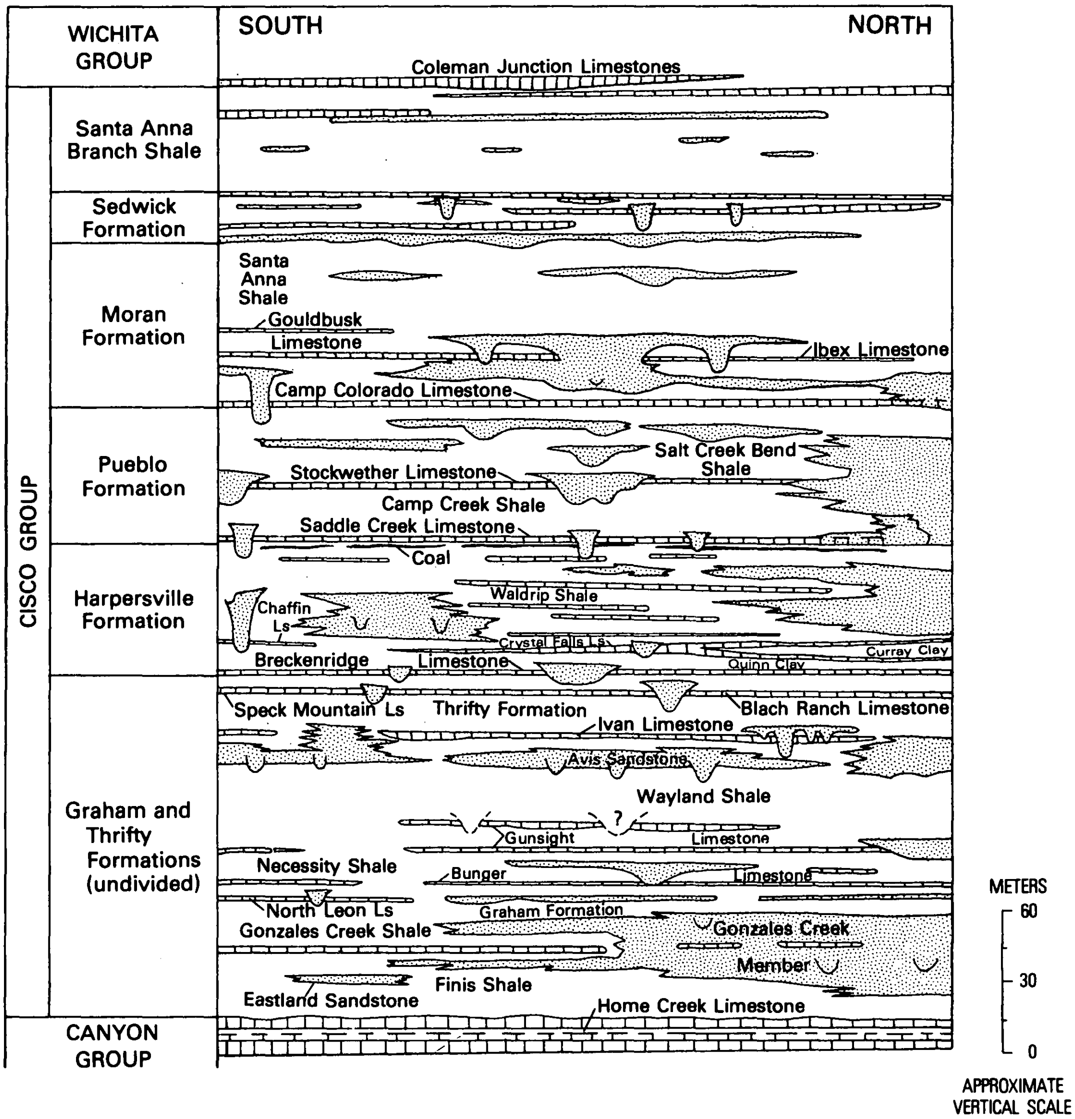

FIGURE 11.—Schematic cross section of outcropping Cisco Group in Brazos River valley. Modified from Brown and Goodson (1972).

Cisco Group from thick limestone and shale units within the underlying Canyon Group and overlying Wichita-Albany Group. Schematic cross sections of outcropping Cisco Group in the Brazos and Colorado River valleys are shown in figures 11 and 12. The Cisco Group as defined by the Bureau of Economic Geology (Brown and Goodson, 1972) is Virgil and, in part, Wolfcamp in age. Consequently, the Pennsylvanian-Permian boundary is within the group (fig. 4).

After Cummins' (1891) establishment of the Cisco, Drake (1893) divided it into "beds," many of which correspond to members in later classifications. Plummer (1919) proposed a preliminary classification of Pennsylvanian strata, including the Cisco, in which the tops of limestone beds were used as formational contacts. Lee (1938) first documented the complexity of Cisco facies and suggested some specific depositional conditions under which the group accumulated.

Cheney (1940) and Eargle (1960) redefined Cisco contacts to coincide with paleontologically inferred time boundaries, consequently elevating the Cisco Group (lithostratigraphic unit) to Cisco Series (time-stratigraphic unit). Brown (1959; 1960a, b; 1962; 1969a, b, c, d) and his students (McGowen, 1964; Seals, 1965; Waller, 1966; Galloway, 1970) studied the Cisco Group outcrop and subsurface in north-central Texas and devised a stratigraphic, sedimentologic, and structural framework for the

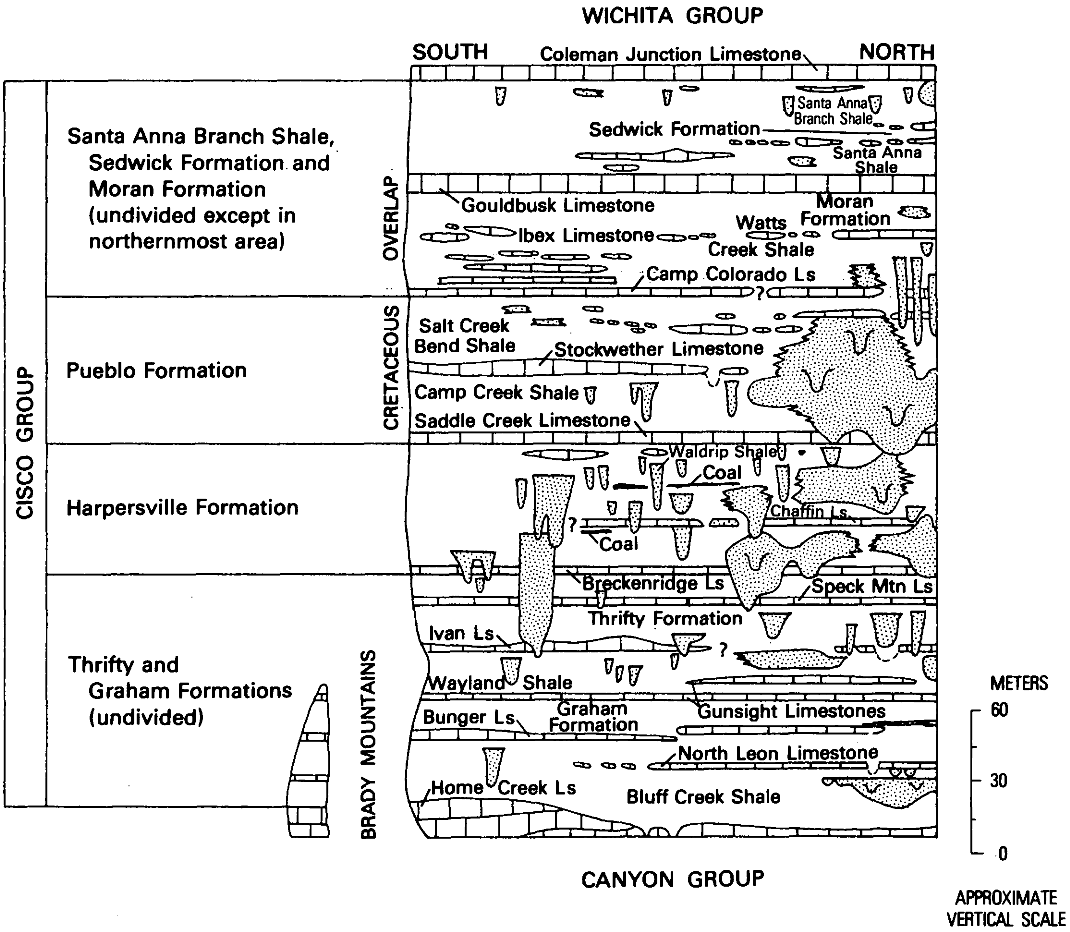

FIGURE 12.—Schematic cross section of outcropping Cisco Group in Colorado River valley. Modified from Kier and others (1976).

region. They recognized the original lithostratigraphic significance of the Cisco (Group) in surface and subsurface mapping programs aimed at lithogenetic interpretation of facies. Galloway and Brown (1972, 1973) presented an integrated depositional interpretation of middle Cisco rocks from outcrop, across the eastern shelf, into the Midland basin.

Brown and Goodson (1972) mapped outcrops of the Cisco Group throughout the Brazos River valley, and Kier and others (1976) extended this mapping into the Colorado River valley. Using the bases of widespread limestone beds as contacts, they divided the Cisco Group into six mappable formations in the Brazos River valley and four formations in the Colorado River valley (figs. 4, 11, 12). They placed the base of the Cisco Group at the top of the Home Creek Limestone (Canyon Group), and the top of the Cisco at the base of the Coleman Junction Limestone, following the original group definition of Plummer and Moore (1921).

The Pennsylvanian-Permian boundary, which is within the Cisco Group as defined by the Bureau of Economic Geology (Brown and Goodson, 1972), has been the subject of controversy for more than a century (San Angelo Geological Society, 1958). Moore (1940, fig. 4) illustrated various opinions about the placement of the time-stratigraphic boundary in Texas during the previous 60 years. The controversy is not yet resolved in North America (Wilde, 1975a, 1975b). The Pennsylvanian-Permian boundary in central and north-central Texas is difficult to determine because (1) no obvious regional physical or paleontological break provides a convenient boundary; (2) different boundaries have been selected using different faunal elements (fusulinids, brachiopods, or ammonites); (3) the Pennsylvanian-Permian boundary in the Glass Mountains of Texas, the reference area for the North American Permian, has not yet been settled (Cooper and Grant, 1972; Wilde, 1971, 1975a); and (4) few appropriate paleontological investigations have been carried out in the Cisco Group of central and north-central Texas, and none have been related to Cisco biofacies.

Nevertheless, fusulinids have been the basis for zonation in the Carboniferous and Permian throughout the world, and they have been used to recognize a Pennsylvanian-Permian boundary in North America for more than 30 years (Wilde, 1975b). The boundary in North America has been recognized by certain species of *Triticites* and the genus *Dunbarinella* in latest Pennsylvanian strata and by *Schwagerina, Pseudofusulina, Leptotriticites,* and other species of *Triticites* in earliest Permian beds (Wilde, 1975a). Roth (1931) found *Pseudofusulina* and Permian *Triticites* in Drake's (1893) Waldrip No. 2 limestone, which is within the shale (Waldrip Shale) between the Chaffin (Crystal Falls) and Saddle Creek Limestones (figs. 4, 11, 12). Cheney (1940) and Moore (1949) placed the Pennsylvanian-Permian boundary in a shale unit below Waldrip No. 2, suppressed the Harpersville Formation (Plummer and Moore, 1921), which included the boundary, and redefined and elevated the Cisco Group to Series and the Thrifty and Pueblo Formations to Groups. Their contact between the redefined Thrifty and Pueblo "Groups" was placed at the inferred time-stratigraphic boundary. Because this boundary is not mappable, Henbest (1958), Eargle (1960), Myers (1965), and others of the U.S. Geological Survey placed the Pennsylvanian-Permian boundary at the base of the Waldrip Shale (top of Crystal Falls Limestone, where present). Brown (1959) argued against suppression of the Harpersville as well as elevation of Cisco to series status. He preferred to apply the time-stratigraphic unit, the "Virgil Series" in Texas, rather than to redefine lithostratigraphy to "fit" inferred faunal-zone boundaries. Consequently, Brown and Goodson (1972) resurrected the Harpersville Formation as a regionally mappable formation and placed the highly subjective Pennsylvanian-Permian boundary in the upper one-third of the Harpersville Formation somewhere between the Chaffin (Crystal Falls) Limestone and the Saddle Creek Limestone.

The Cisco Group represents the last major episode of extensive fluvial-deltaic deposition on the Eastern shelf. After Cisco deposition, Middle and Late Permian carbonate, evaporite, and red-bed deposition dominated the shelf. Sandstone and shale are principal rock types in the Cisco Group; only thin transgressive shelf limestone beds crop out in central and north-central Texas. Downdip in the subsurface, thick shelf and shelf-edge limestone facies are common, but they commonly pinch out updip into nearshore clastic deposits. When each delta system was abandoned, it subsided, was reworked, and was transgressed by marine destructional facies (barrier and nearshore sands) and ultimately by marine shale and limestone (fig. 13). Cycles composed of regressive terrigenous clastic deposits and transgressive limestone deposits make up vertical sequences that show abrupt lateral facies changes (fig. 14). Brown (1973b) recognized 10 to 15 principal fluvial-deltaic progradational (regressive) episodes in the Cisco Group of north-central Texas.

In the Brazos River valley area during deposition of the Cisco Group, a westward or basinward shift in facies took place so that outcropping Cisco facies grade progressively upward (Brown, 1973b; fig. 14) from principally deltaic in the lower part (Virgil Series) to principally fluvial in the upper part (Wolfcamp Series). Fluvial facies recognized in outcrop are parts of braided, coarse-grained meanderbelt and fine-grained meanderbelt systems. Delta facies are components of thin high constructive elongate and lobate systems. Distribution of Cisco deltas was controlled by subjacent paleotopography induced by Canyon carbonate banks, differential sand/mud compaction, and differential rates of structural subsidence (Brown, 1969c). Interdeltaic-embayment facies include mudflats, chenierlike strandplains and coal, and brackish-bay mudstone and limestone (Galloway and Brown, 1972).

In the Colorado River valley, terrigenous clastic facies are restricted principally to the lower part of the Cisco Group (fig. 12) within the Graham, Thrifty, Harpersville, and Pueblo Formations. Scattered sand-filled fluvial channels are found in the upper part of the Cisco Group, and lenticular limestone beds are common.

Total thickness of the Cisco Group is 394 m in the Brazos River valley and 303 m in the Colorado River valley.

MARATHON UPLIFT

SYSTEMS BOUNDARIES

Neither the Devonian-Mississippian nor the Mississippian-Pennsylvanian boundary has been recognized in the Marathon uplift (fig. 4). The Mississippian-Pennsylvanian boundary clearly is within the Tesnus Formation; the Tesnus contains fossils as young as Early Pennsylvanian and as old as Late Mississippian (Ellison, 1962). The youngest (and only) fossils dated from the Caballos Novaculite, which apparently conformably underlies the Tesnus,

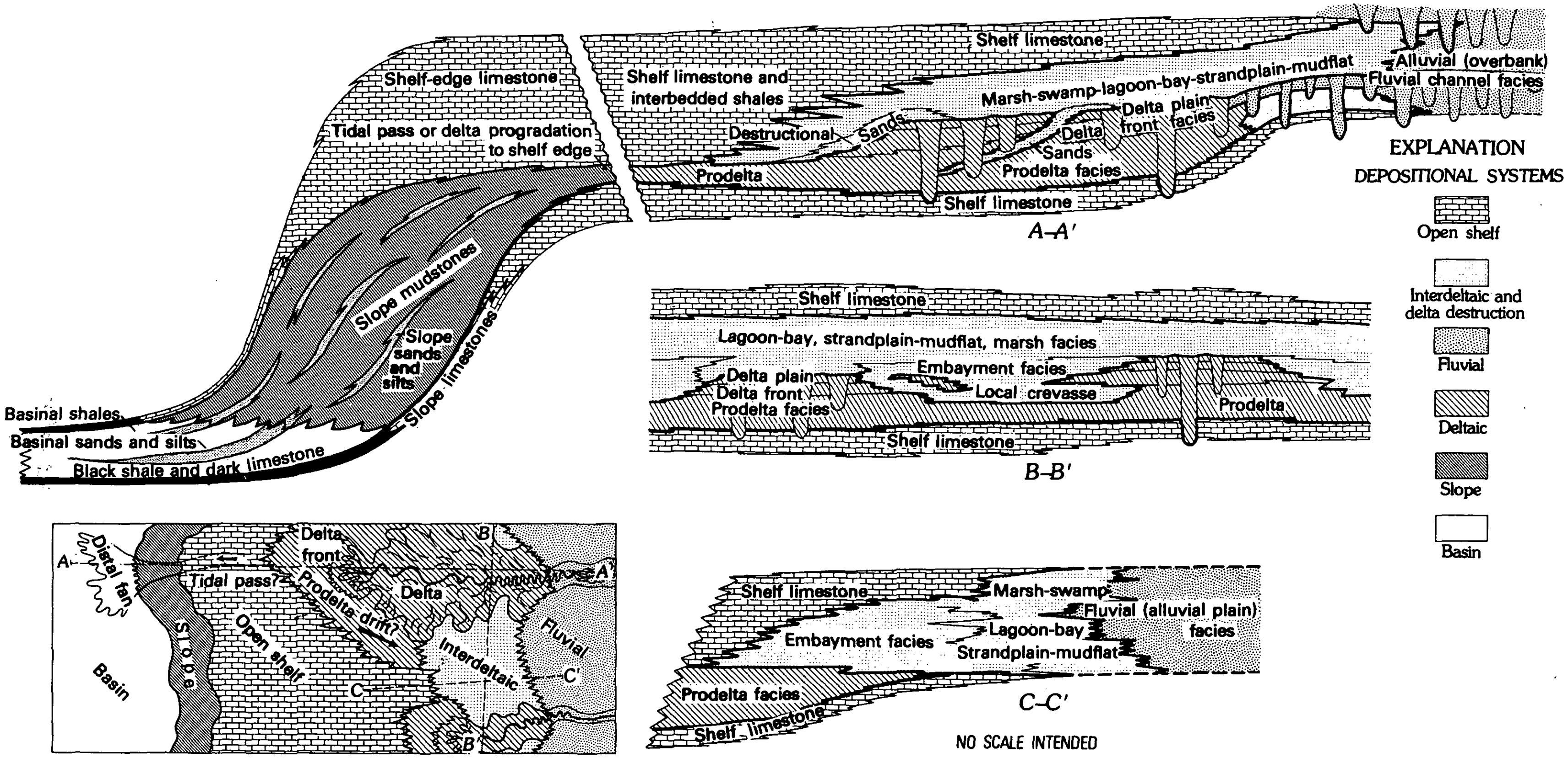

FIGURE 13.—Schematic cross section along Cisco paleoslope showing principal depositional systems. Based on 15 cross sections and 13 net-thickness maps. From Brown (1969d); reprinted with permission, Dallas Geological Society.

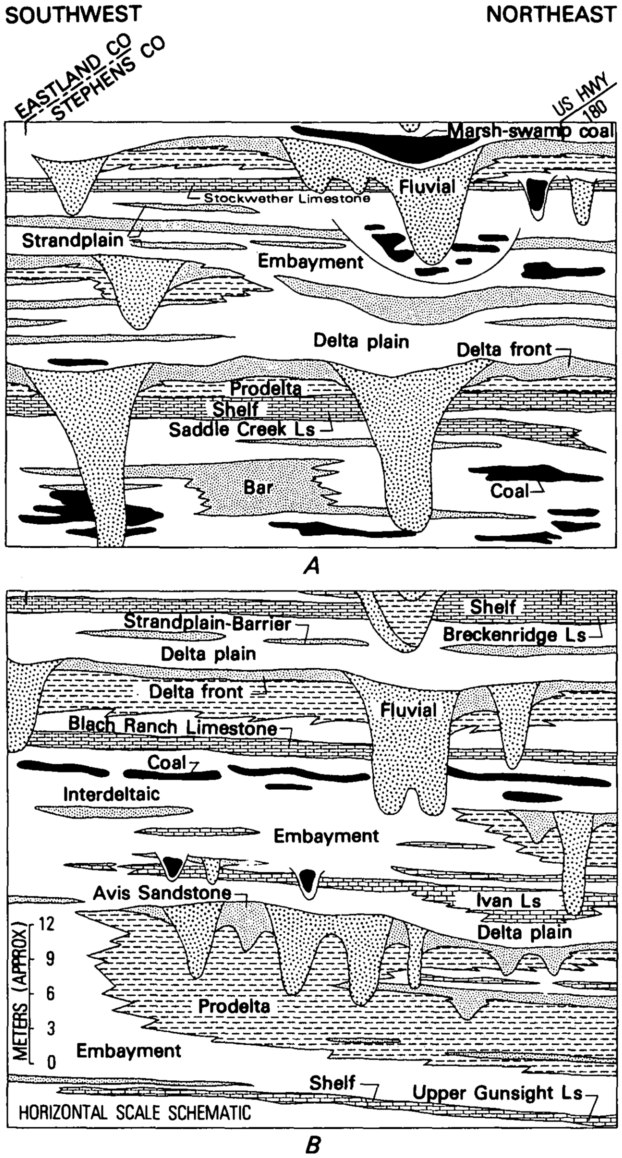

FIGURE 14.—Nature of outcropping cyclic facies, Cisco Group, Stephens County, Tex. *A*, Fluvially dominated marine-nonmarine cycles, upper part of Cisco Group. *B*, Delta-dominated marine-nonmarine cycles, lower part of Cisco Group. Based on detailed mapping and 350 measured sections. From Brown (1973).

are Late Devonian conodonts (Graves, 1952) some 50 m below the Caballos-Tesnus contact. Late Mississippian conodonts are found 5–10 m above the Caballos-Tesnus contact.

CABALLOS NOVACULITE

The Caballos Novaculite is a lens-shaped sequence 30–250 m thick containing equal amounts of massive novaculite (pure white chert) and thin-bedded green to gray radiolarite rhythmically interbedded with shale. The Devonian-Mississippian boundary probably is within the upper radiolarite. The upper boundary of the Caballos Novaculite is placed at the uppermost chert bed thicker than 5 cm. The formation was named by Udden, Baker, and Böse (1916).

TESNUS FORMATION

The Tesnus Formation (named Tesnus Shale by Baker and Bowman, 1917) is a wedge-shaped unit composed predominantly of repetitive beds of olive-drab to black shale and fine- to very fine grained sandstone. The formation is 2,000 m thick in the southeast outcrop area but thins westward to 100 m. A blanketlike shale about 100 m thick forms the base of the Tesnus in the east and composes nearly the entire formation in the west (fig. 15). An olistostrome 10 m thick containing exotic blocks of novaculite and older Paleozoic shelf carbonates crops out at the base of the Tesnus in the easternmost exposures (McBride, 1978). The olistostrome conformably overlies the Caballos Novaculite and is composed of siliceous shale in the lowermost part. Overlying beds of siltstone and sandstone gradually increase in thickness, grain size, and abundance and form a transition into the monotonous clastic Tesnus Formation. Sandstone layers range from 1 mm to 3 m thick, but most are 30–150 cm thick. Sandstone-shale ratios range from about 1:1 to 5:1, with an apparent, but undocumented, increase to the east.

Carbonaceous plant fragments and spores are locally abundant, and casts of wood as much as 30 cm long (chiefly calamarians, pteridosperms, and *Lepidodendron* (King, 1937, p. 61)) are found in shale. Sparse conodonts (Ellison, 1962), sponge spicules, radiolarians (Baker, 1963; Cotera, 1969), inarticulate brachiopods (J. Sprinkle, oral commun., 1976), and a single crustacean (Brooks, 1955) also have been found.

Most sandstones are olive-drab to light-brown, nonporous beds that contain 5–15 percent clay matrix and less than 10 percent quartz cement. Major framework components other than quartz are 5–10 percent each of feldspar and rock fragments. Several well-sorted, quartz-cemented quartzarenite beds, 10–15 m thick, are found only in the southeastern part of the basin.

Probably the most striking aspect of the Tesnus Formation in outcrop is the abundance of slump structures. In places, it is impossible to find 10 m of undisturbed (unslumped) section. Slump features include warped, folded, disrupted, and broken sandstone beds and contorted shale beds. Sandstone dikes of uneven thickness, but generally less than 5 cm

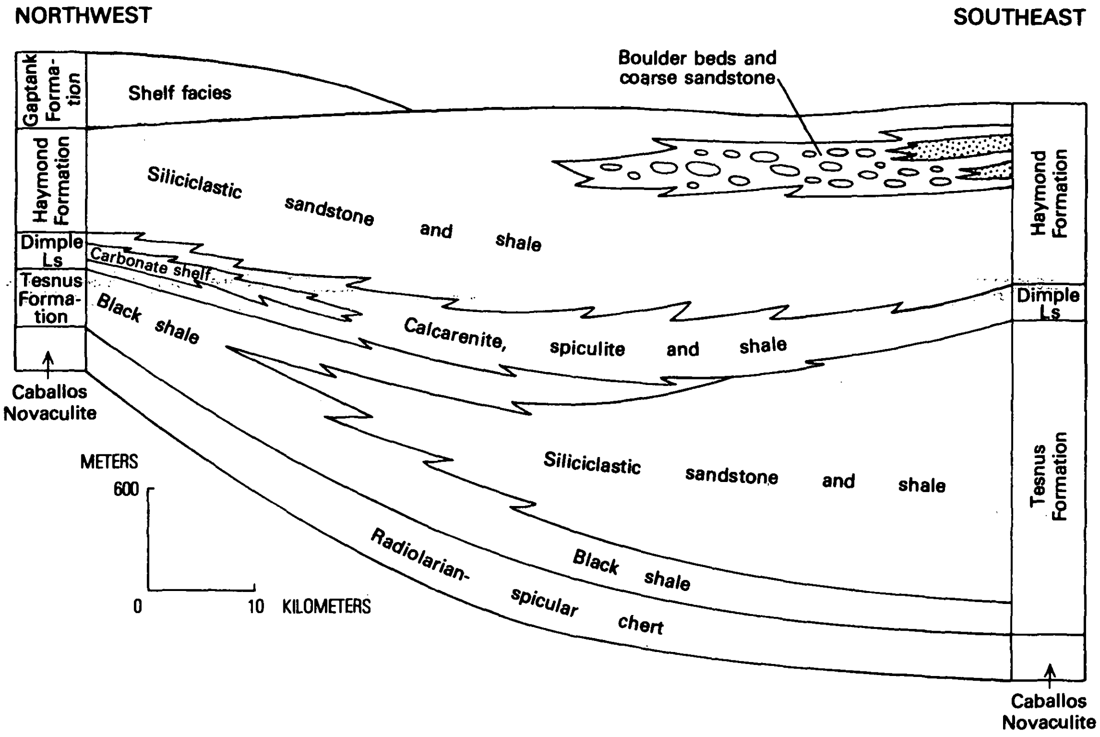

FIGURE 15.—Schematic cross section of flysch units approximately perpendicular to axis of Marathon geosyncline. From Thomson and McBride (1964).

thick, are associated features. Soft-sediment faults that transect sandstone soles are commonly regularly oriented and can be used to determine paleoslope direction (Thomson, 1973).

DIMPLE FORMATION

The Dimple Formation (named the Dimple Limestone by Udden and others, 1916, p. 46) has a gradational basal contact with the Tesnus. Over a 5-m interval, limestone beds and intercalated shale beds increase in thickness until they predominate, marking the basal Dimple contact. Maximum thickness of the formation is 300 m; the formation thins to 150 m at the margins of the Marathon uplift.

From northwest to southeast over a distance of 35–45 km, the Dimple Formation grades from rocks interpreted as "shelf" to slope and then to basin deposits (Thomson and Thomasson, 1969). The "shelf" facies are characterized by intercalated, crossbedded biosparite and oosparite (grainstones) in beds, most of which are 30–50 cm thick, lesser amounts of chert and limestone pebble conglomerate, and partings of shale that are 75 percent carbonate mud and 25 percent illite. Skeletal grains include crinoids, bryozoans, brachiopods, echinoids, fusulinids, algae, and conondonts. Trace amounts of quartz, chert, glauconite, phosphate, and heavy minerals are present. Grain size within calcarenite beds ranges from coarse to fine; sorting is good, and porosity may be as much as 10 percent.

Rocks of the basin facies of the Dimple Formation are about 70 percent calcarenite and cherty calcarenite, 20 percent mudstone, and 10 percent chert. Calcarenite and some mudstone beds are interpreted to be turbidites, whereas mudstone and spicular and radiolarian chert beds are pelagic deposits. Calcarenite beds are chiefly less than 30 cm thick, but some are 150 cm thick. Most calcarenite beds are uniform in thickness across an outcrop. Also, most beds (96 percent) are graded, but many contain other bedding types.

Rocks of the slope facies of the Dimple Formation are intermediate between shelf and basin facies. The most conspicuous feature of this facies is the presence of chert and limestone pebble conglomerate in beds 10–50 cm thick that fill erosional surfaces with as much as 30 cm relief and of calcarenite beds 60–100 cm thick that contain a single crossbed set. A 10-m-thick olistostrome containing exotic blocks crops out in easternmost exposures.

Most of the Dimple Formation is Atoka in age, but fusulinids suggest that the oldest part may be Morrow (Sanderson and King, 1964).

HAYMOND FORMATION

The Haymond Formation has a maximum preserved thickness of 1,400 m in outcrop. The basal contact with the Dimple Formation is gradational over a thickness of 5 m. The placement of the upper contact with the Gaptank Formation is disputed.

About three-fourths of the Haymond is composed of thin rhythmic beds of shale and fine-grained quartzose sandstones, which comprise more than 15,000 shale/sandstone couplets. The formation contains an olistostrome and associated lenticular coarse sandstone bodies that are exposed in strike sections along the eastern margin of the uplift. The olistostrome, or wildflysch unit, is 330 m thick and contains indigenous and exotic debris from pebbles to blocks 40 m long.

Sandstone beds have 3–10 percent clay matrix and less than 10 percent calcite or quartz cement. Cementation and compaction have destroyed all porosity. Framework composition averages 71 percent quartz, 15 percent feldspar, and 14 percent rock fragments. In the rhythmic sequences, sandstone beds are graded, laminated, and current rippled; many show convolute lamination. Current-formed sole marks are abundant.

The age of the Haymond is uncertain. Although an abundant Pennsylvanian fauna has been collected from exotic blocks in the boulder beds (King, 1937, p. 72), the only abundant indigenous fossils in the Haymond are plant remains and trace fossils. Fragmental remains of fusulinids, echinoids, brachiopods, and mollusks are found in the area of Dugout Mountain in beds transitional between Haymond and the overlying Gaptank Formation. In addition, fusulinids and echinoid fragments are found in two calcarenite beds in the lower part of the formation. An Atoka age (King, 1937, p. 72; Skinner and Wilde, 1954; Wilde, oral commun., 1962) was assigned to the fusulinids. These beds are turbidites of "Dimple aspect" that are several hundred feet above the base of the Haymond Formation.

DEPOSITIONAL ENVIRONMENTS OF THE CABALLOS NOVACULITE AND OVERLYING FLYSCH

Environments and processes of deposition of the flysch units and the underlying Caballos Novaculite remain controversial. Most early workers interpreted all formations to be of shallow-water origin (Waterschoot van der Gracht, 1931; King, 1937), whereas most later workers (King *in* Flawn and others, 1961, p. 84; Johnson, 1962; Cotera, 1969; McBride and Thomson, 1970; McBride, 1970) infer a deepwater origin. McBride (1970) interprets the Tesnus to be chiefly submarine fan deposits, and the Haymond to include submarine fan, basin-plain, and large-scale submarine slide deposits (wildflysch unit). Folk (1973) and Flores (1972, 1977), however, have argued for shallow-water origin of facies in the Caballos, Haymond, and Tesnus.

GAPTANK FORMATION

The Gaptank Formation conformably overlies the Haymond Formation. Although exposures are relatively small, stratigraphic subdivision of the Gaptank Formation has been the subject of controversy. Figure 16 charts the history of classification and shows various contacts within the Gaptank Formation exposed in the northern part of the Marathon uplift. A Chaetetes-bearing limestone 15 m thick is interbedded with limestone and sandstone of early Desmoinesian age. An overlying lenticular conglomerate member as much as 200 m thick is of early Missouri age (Ross, 1963, p. 17); five resedimented debris beds grading from limestone conglomerate to calcarenite characterize this unit. A superposed sandstone and shale member 200 m thick is poorly exposed. The uppermost ledge-forming limestone member about 150 m thick is composed of limestone bodies that rise southward in the section. Ross (1967) recognized shallow-shelf, shelf-edge, and deepwater facies in the uppermost limestone member. Conspicuous grainstones, packstones, and wackestones are present, and local bioherms are rich in dasycladacean algae.

According to Ross (1967) the outcrop belt of the Gaptank trends at a low angle to the northeast to north-northeast depositional strike of the formation. He inferred that the Gaptank is composed of several cyclic carbonate facies, which include (upward) slope, shelf-edge, and shelf deposits. Gaptank carbonate facies separate shallow-water clastic deposits to the southeast from deep-water clastic deposits to the northwest.

FRANKLIN AND HUECO MOUNTAINS

In the Hueco Mountains, the Helms Formation is composed of interbedded limestone and shale 150 m thick and is divided into a lower cherty member and chert-poor upper member. King and others (1945) reported that the Helms is sparingly fossiliferous but listed no fossils. The Magdalena Group, which rests unconformably on the Helms, was divided by King and others (1945) into three informal members. The lower member consists of 150 m of dark-gray, thick-bedded biomicrite; the

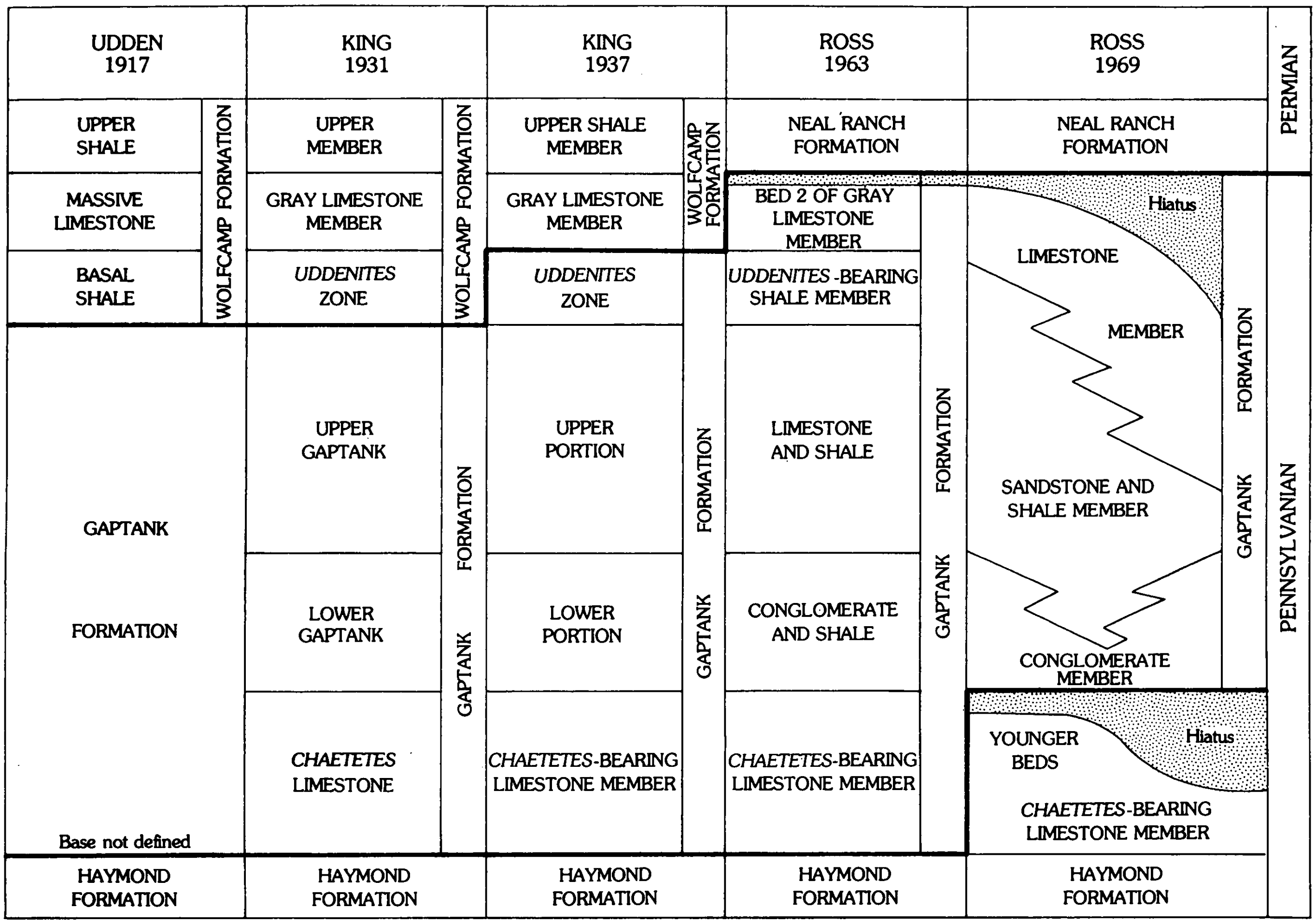

FIGURE 16.—Stratigraphic nomenclature of the Gaptank and adjacent formations. From Ross (1969).

middle member is composed of 100 m of marl, shale, and limestone; and the upper member is composed of 150 m of light-gray coral and algal biomicrite, biosparite, and conglomerate (Seewald, 1968).

In the Franklin Mountains, Mississippian strata include (upward) the Las Cruces Limestone, an even-bedded (beds 40–60 cm thick) gray micrite 20 m thick; the Rancheria Formation, a black argillaceous and cherty limestone 80 m thick; and the Helms Shale, a gray and green shale 50 m thick containing interbedded sandstone and limestone (Laudon and Bowsher, 1949). Unconformities bound the Las Cruces, Rancheria, and Helms Formations. The overlying Magdalena Group consists of more than 500 m of thin-bedded, dark-gray micrite and biomicrite containing some minor shale beds. The Berino Member is a shaly unit about 155 m thick, which separates the overlying Bishops Cap Member from the basal La Tuna Member. The uppermost, and unnamed, unit is poorly exposed and presumably is composed mostly of shale. The Magdalena-Helms contact is inferred to be unconformable (King and others, 1945). Environmental interpretations have not yet been reported for the Franklin Mountain sequence.

BOUNDING UNITS

CENTRAL AND NORTH-CENTRAL TEXAS

LOWER BOUNDARY

In outcrop, basal Mississippian rocks (fig. 4) generally unconformably overlie the lower part of the Ordovician Ellenburger Group, a thick group of carbonate strata that crop out around the Llano uplift (fig. 3) and extend throughout the subsurface of west Texas. The pre-Mississippian erosion surface appears to have little or no topographic relief. Regionally, however, the pre-Mississippian unconformity is angular (Barnes and Cloud, 1972).

The Ellenburger Group is composed predominantly of limestone and dolomite that accumulated in warm, shallow-marine environments "sedimentologically and ecologically" similar to the Bahamian Banks (Barnes and Cloud, 1972, p. 32). Numerous sinkholes formed in the Ellenburger during one or more periods of exposure, and post-Ellenburger strata commonly collapsed into or were originally deposited in them. Very locally, Upper Ordovician, Silurian, and Devonian limestone, shale and chert breccia (assigned to the Burnam, Starke, Piller Bluff, Stribling, Bear Spring, Zeson, and Houy Formations and several unnamed stratigraphic units) are preserved in collapse structures, cracks, and fissures in the top of the Ellenburger (Barnes and Cloud, 1972; Barnes and others, 1946, 1947, 1953, 1966; Cloud and others, 1957; Seddon, 1970). Aggregate thickness of the sinkhole deposits is less than 5 m.

Stratigraphic relationships among post-Ellenburger and pre-Mississippian deposits, as well as relationships with overlying Mississippian strata, are poorly understood. Stratigraphic and paleontological information about the remnant formations is limited. The Houy Formation apparently is, in part, Mississippian (Cloud and others, 1957; Seddon, 1970) and may record continuous deposition across the Devonian-Mississippian boundary. (See Seddon, 1970; Barnes and Cloud, 1972; and Kier, 1972.) Distribution of post-Ellenburger and pre-Mississippian rocks is so limited on the Llano uplift that the basal Mississippian boundary is essentially unconformable.

UPPER BOUNDARY

Permian rocks conformably overlie Pennsylvanian rocks in central and north-central Texas and record continuous sedimentation across the Pennsylvanian-Permian boundary. The boundary, as defined by fusilinids, is in the upper part of the Harpersville Formation (Plummer and Moore, 1921; Brown and Goodson, 1972) in the middle of the Cisco Group. No regionally significant hiatus has been recognized. Deltaic sedimentation, which dominated Cisco deposition, continued uninterrupted from Late Pennsylvanian into Early Permian time.

Lower Permian (Wolfcamp) deposition was marked by a gradual reduction in the influx of Cisco terrigenous clastic materials. During this time, shelf carbonates and marine muds onlapped the foundering Cisco deltas, and shelf environments dominated the Eastern shelf. Middle Permian carbonate and marine clastic shelf environments expanded, especially in central Texas, and were replaced upward by Middle and Upper Permian tidal-flat and sabkha environments (Smith, 1974).

WEST TEXAS

In the Marathon region of west Texas, the lower part of the Tesnus Formation, which is of Mississippian age, gradationally overlies the Caballos Novaculite, which is inferred to be principally Devonian in age. Late Devonian conodonts are found about 50 m below the top of the Caballos (Graves, 1952), and Late Mississippian conodonts are found 5–10 m above the base of the Tesnus (Ellison, 1962).

An angular unconformity separates the youngest Carboniferous unit, the Pennsylvanian Gaptank Formation, from Permian strata, either the Neal Ranch Formation or the Lenox Hills Formation of early and late Wolfcamp age, respectively (Ross, 1963). The Neal Ranch Formation is composed of shale, limestone, and siltstone cyclothems, and the Lenox Hills Formation consists of chert and limestone conglomerate, shale, and siltstone. Both formations were deposited in shallow water on the northern flank of the Marathon orogenic belt (Ross, 1963).

In the Franklin Mountains, limestone of Kinderhook age (Las Cruces Formation) rests unconformably on Upper Devonian shale (Percha Shale) according to Nelson (1940). In the Hueco Mountains, limestone beds of Meramec age (Helms Formation) rest on shale inferred on lithic character to be of Late Devonian age (King and others, 1945).

At the outcrop in the Franklin Mountains the Pennsylvanian-Permian boundary is inferred to be in the covered interval between limestones in the upper part of the Magdalena Group (Missouri age) and the overlying Hueco Formation (Wolfcamp age). In the Hueco Mountains, the boundary is placed within the Hueco Group (Seewald, 1968, p 47).

SUBSURFACE GEOLOGY

The subsurface geology of Carboniferous strata on the eastern shelf of north- and west-central Texas and within the Midland basin of west Texas is as well known as any sequence of rocks in the world (fig. 3). Tens of thousands of oil wells have penetrated fluvial and deltaic, shelf and shelf-edge, carbonate platform, and slope and basin facies throughout the region. Less intensively drilled but still reasonably understood are Carboniferous rocks in the Palo Duro and Dalhart basins of the Texas Panhandle and the Fort Worth, Kerr, and Val Verde Foreland basins that separate the Ouachita geosyncline on the east and southeast from the Texas cratonic region on the west and northwest.

Carboniferous strata in the Delaware basin and beneath thick Mesozoic and Cenozoic rocks along the Rio Grande in Trans-Pecos Texas are poorly understood because of structural complications and limited well information. Carboniferous rocks of the Marathon region are allochthonous, and little is known about their subsurface extent. Subsurface geology of Carboniferous strata within the basins and on the shelves of the west Texas oil province necessarily must be generalized in this report.

MISSISSIPPIAN ROCKS

Mississippian strata are of two principal types within the subsurface of Texas: (1) thick shale beds and interbedded sandstone within the Ouachita geosyncline and (2) limestone and marine shale facies on positive elements of the Texas craton (Texas peninsula, Adams, 1954; Concho arch, Cheney and Goss, 1952) and within shallow flanking basins, respectively. Within the frontal zones of the buried Ouachita foldbelt (geosyncline) are thick sequences of Mississippian fan-delta and slope shale and interbedded sandstone (Stanley, Jackfork, lower Johns Valley, and equivalent formations; Flawn and others, 1961). Upper Mississippian and Lower Pennsylvanian terrigenous clastic facies were deposited within the rapidly subsiding geosyncline and were highly folded and faulted during Early Pennsylvanian time in central and north-central Texas. Farther southwest, in the Val Verde basin (Young, 1960) and in the Marathon region, orogenic deformation took place as late as Early Permian.

Mississippian cratonic facies crop out in small areas of the Llano uplift and in some faulted mountains of southwest Texas and southeastern New Mexico. These shallow-marine facies grade into or intertongue with shale deposited in shallow flanking and intracratonic Mississippian basins (Barnett Formation). Pre-Pennsylvanian erosion resulted in the removal of Mississippian strata over significant areas of the Texas craton. Locally, as in Eastland, Stephens, Young, and Jack Counties (fig. 5), Mississippian biohermal reefs(?) formed along the eastern flank of the Concho arch or platform. Geosynclinal Mississippian deposits are restricted to the subsurface in central Texas, although the sequences crop out in southern Oklahoma and in the Marathon area of southwest Texas.

PENNSYLVANIAN ROCKS

In central and north-central Texas, Lower Pennsylvanian subsurface strata are inferred to have been deposited by two depositional systems (fig. 17). Fan-delta and slope clastic deposition (Atoka Group and part of Smithwick Formation) dominated in the Forth Worth foreland basin, supplied by the orogenically active Ouachita foldbelt. Contemporaneous deposition of limestone and marine shale (Marble Falls Formation and associated units) took place on the Concho platform and eastern platform margin. In west Texas and the Texas Panhandle, Lower Pennsylvanian platform carbonate facies grade laterally into arkosic fan-delta and slope facies near positive structural elements and grade basinally into deeper marine limestone and shale within the incipient Midland basin.

Middle Pennsylvanian subsurface strata (Strawn and Canyon Groups) are composed dominantly of proximal fluvial-delta facies from a diminishing sediment supply eroded from the Ouachita Mountains and from older Atoka foreland deposits. The deltas repeatedly prograded westward across the Concho platform into the deepening Midland basin (figs. 17, 18, 19). Extensive, contemporaneous, carbonate shelf-edge bank systems intertongue updip (sourceward) with nearshore clastic deposits and basinward with thin basinal shale and limestone beds. Middle Pennsylvanian carbonate shelves define the eastern and western flanks of the Midland and Palo Duro basins. As the Midland basin deepened during Middle Pennsylvanian time, the carbonate shelves were successively superposed or, in many places, they retreated landward in a recessive manner. Many isolated carbonate platforms formed at this time, such as the Scurry County "Horseshoe Atoll" and many en echelon banks formed along the edge of the relict Eastern shelf (fig. 3). Carbonate deposition continued on the Red River fault blocks.

Upper Pennsylvanian subsurface strata in north-central and west Texas reflect rejuvenation of the eastern source area and accelerated subsidence of the Midland basin. Extensive, cyclic fluvial-deltaic sandstone and shale sequences of the Cisco Group are deposited on the eastern shelf of the Midland basin (figs. 20, 21). The delta deposits grade westward into shelf and shelf-edge limestone facies, which in turn grade basinward into deep-water shale and sandstone. Most of the Midland basin contains only very thin Cisco shale and siliceous limestone beds ("starved basin" deposits (Adams and others, 1951)). Basinward progradation of delta, shelf, and slope deposits filled the Palo Duro and northern Midland basins by Early to Middle Permian (fig. 3). Along the margins of the Amarillo uplift and other similar structural elements in

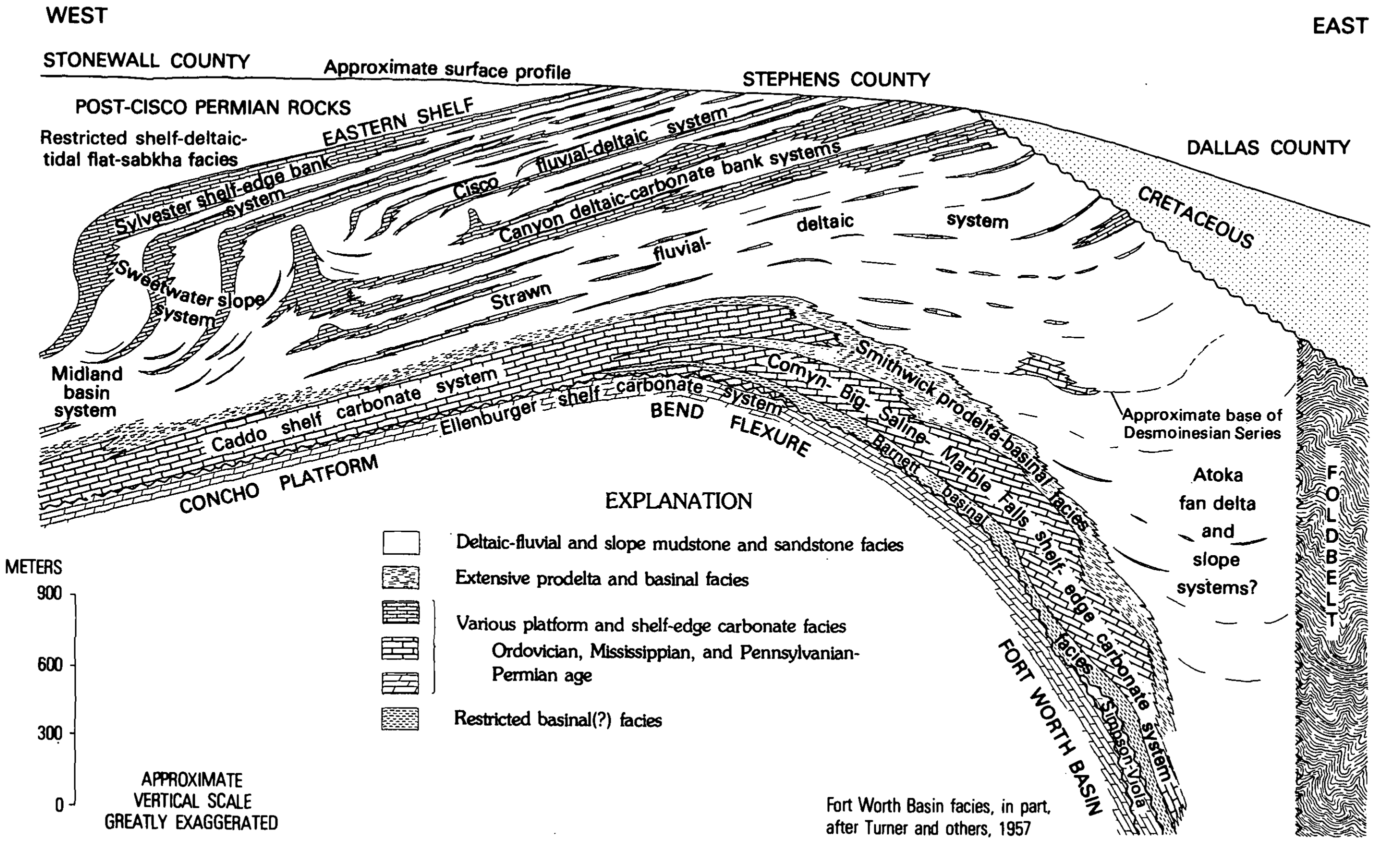

FIGURE 17.—Evolution of depositional systems, north-central Texas: Fort Worth basin, Concho platform, and Eastern shelf.

southern Oklahoma and eastern New Mexico are thick arkosic fan-delta deposits of Late Pennsylvanian age. Platform carbonates were deposited on the central basin platform.

Strata on the northwestern shelf of the Midland basin in eastern New Mexico and the northern shelf of the Anadarko basin in Oklahoma generally resemble sequences that compose the eastern shelf of the Midland basin. The southern flank of the Anadarko basin in Texas is filled by thick arkosic fan-delta deposits.

GEOLOGIC HISTORY

CARBONIFEROUS EVENTS

CENTRAL AND NORTH-CENTRAL TEXAS

The geologic history of central and north-central Texas is closely tied to the tectonic development of the Fort Worth (foreland) basin, the eastern shelf of the Midland basin, and the Red River uplift and southern Oklahoma mountains (fig. 3). Structural evolution of these basins and associated tectonic elements determined to a great extent the nature and distribution of the principal basin-filling depositional systems.

Beginning with Late Mississippian and Early Pennsylvanian structural activity in the Ouachita geosyncline, the Fort Worth foreland basin became well defined (figs. 3, 17). Platform and shelf-edge carbonate environments (Marble Falls, Big Saline, Comyn, and Caddo) contemporaneously dominated the Concho platform. Late Mississippian and Early Pennsylvanian shelf edges faced generally eastward toward the rapidly subsiding, but not necessarily deep, Fort Worth basin. Generally equivalent westward-prograding terrigenous clastic wedges (Atoka Group) entered the basin along a high-gradient paleoslope from the Ouachita foldbelt to the east. Thousands of feet of Atoka mudstone and sandstone of probable fan-delta and related slope origin graded westward and basinward into the thin, relatively starved basinal Smithwick facies. Basinal Smithwick shale and siltstone intertongued westward with shelf-edge carbonates of the Concho platform.

As Atoka clastic wedges built westward under gradually diminished but westward-shifting basinal subsidence, the shelf edges of the Concho platform carbonates retreated westward in a series of

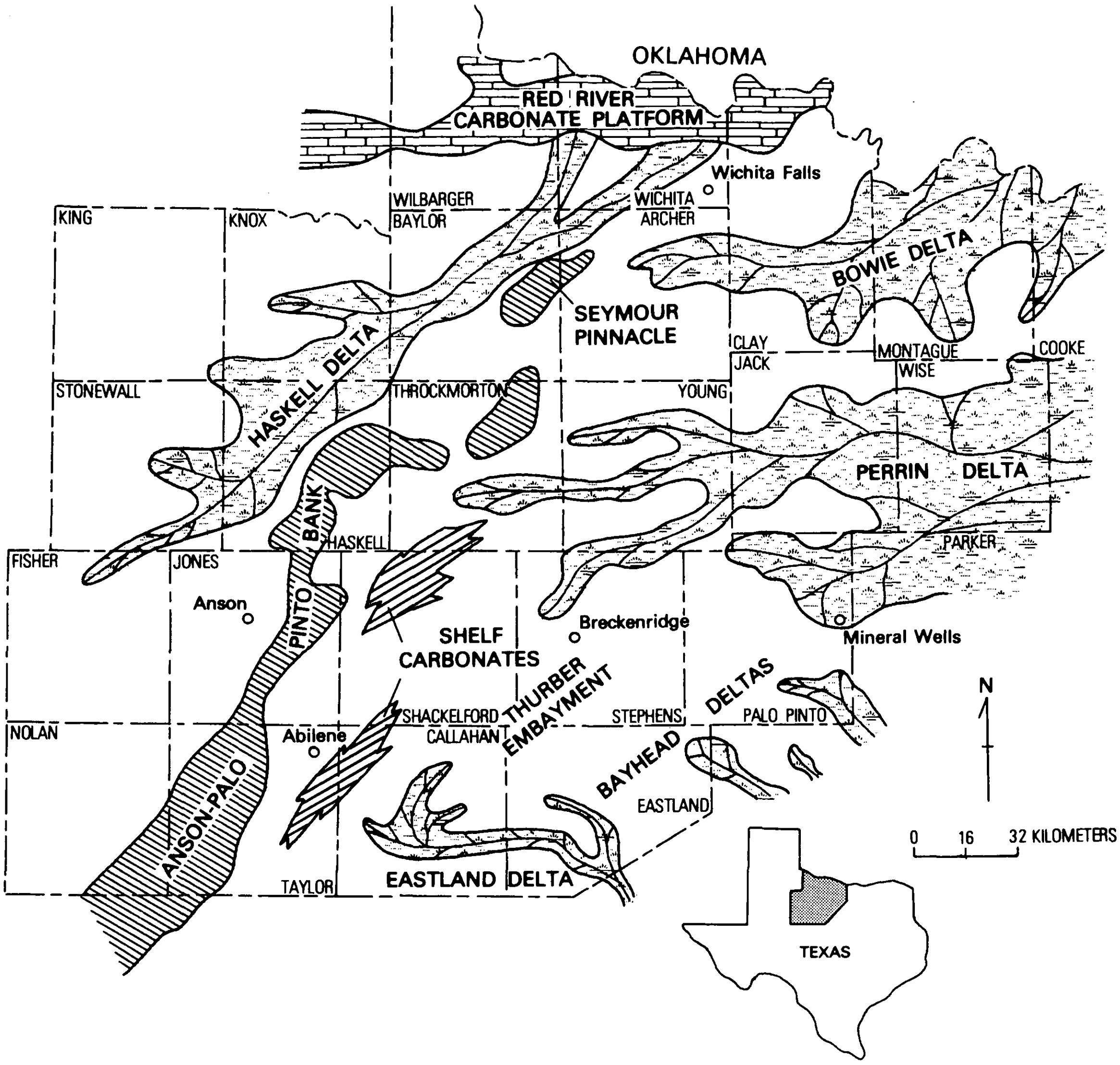

FIGURE 18.—Distribution of Strawn depositional systems, north-central Texas. From Cleaves (1975).

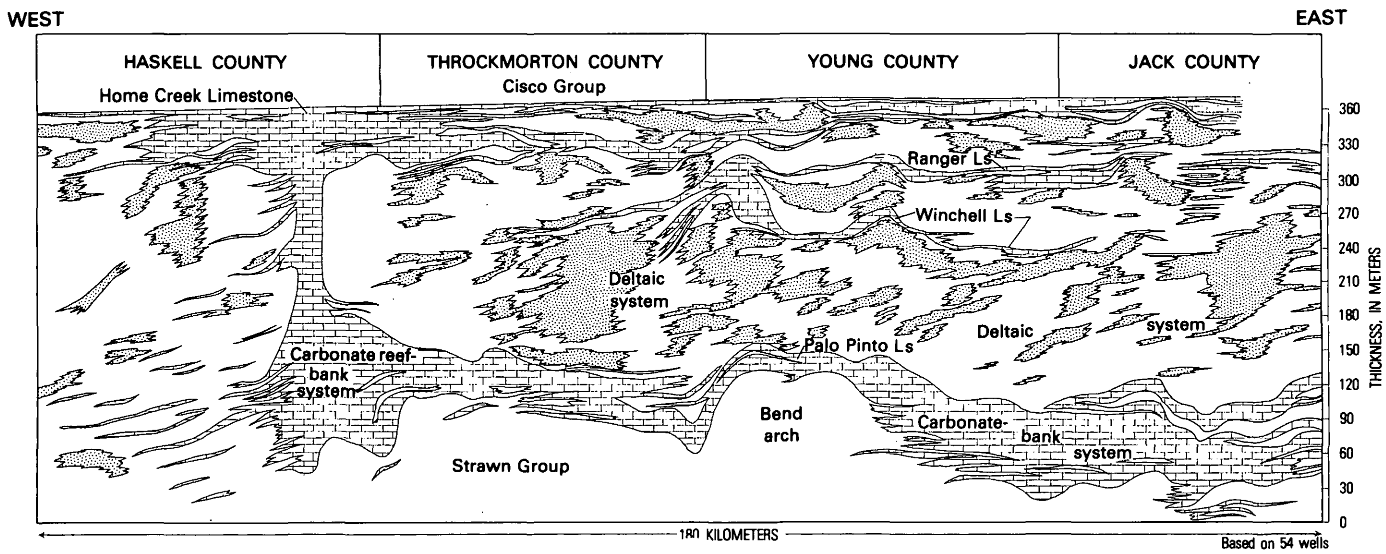

FIGURE 19.—Subsurface cross section of Canyon Group from Jack County to western Haskell County, Tex., showing deltaic and carbonate facies; based on 54 electric and sample logs. From Erxleben (1975).

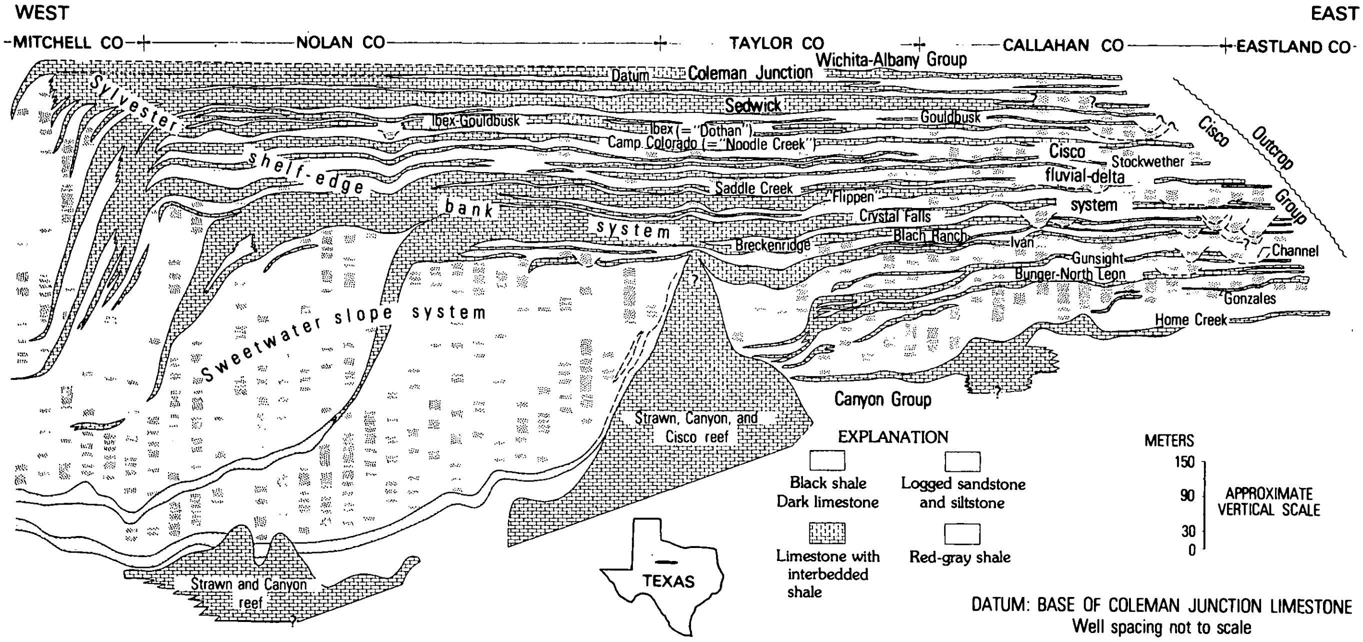

FIGURE 20.—Subsurface cross section showing facies of the Cisco Group, from Mitchell County to Eastland County, Tex. Based on 60 wells. From Brown (1969d). Data on open-file, Texas Bureau of Economic Geology.

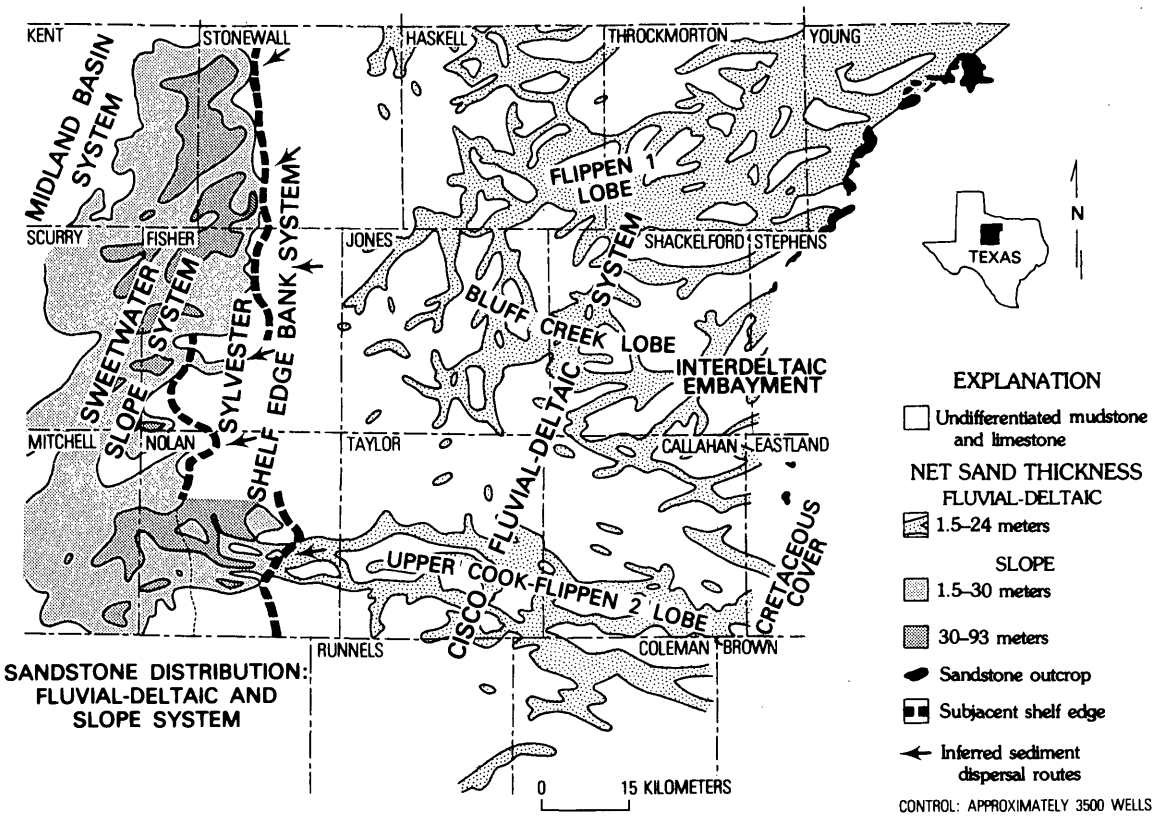

FIGURE 21.—Net-thickness map, Upper Cook-Flippen Sandstone, Cisco Group, north-central Texas. Based on 3,500 wells; data on open-file, Texas Bureau of Economic Geology. From Galloway and Brown (1972, 1973); reprinted with permission, American Association of Petroleum Geologists.

progressive "back steps," overlapped by the advancing Smithwick facies (fig. 17). Coarse clastic facies of Atoka fan deltas ("Bend Conglomerate") reached the western flank of the Fort Worth basin late in the waning stages of Atoka deposition. Facies within the Fort Worth basin, both terrigenous clastic and carbonate, indicate an uncommonly high degree of time transgression. Depositional environments shifted westward as a result of a shift in basin axis. Variable rates of subsidence and sediment supply also affected this shift.

EASTERN SHELF AND MIDLAND BASIN

Decreased subsidence in the Fort Worth basin and diminishing Atoka clastic input marked deceleration of Ouachita orogenic activity. During early Strawn (Desmoines Series) deposition, terrigenous clastic deposits gradually assumed a deltaic character, having lower paleogradients and a very shallow basin. During middle Strawn deposition, fluvial-deltaic systems overlapped the shelf-edge carbonate facies (Marble Falls, Caddo) and began several cycles of extensive progradation westward across the Concho platform (figs. 17, 18). Youngest Smithwick prodelta-basinal facies were deposited in the path of the delta systems. During late Strawn deposition, delta-fluvial sedimentation continued as the Concho platform underwent a gradual westward tilting and increased subsidence in response to accelerated subsiding of the Midland basin. Even though the stability of the Concho platform decreased near the end of Strawn deposition, this structurally positive element provided support for many upper Strawn and Canyon reefs and limestone banks (figs. 17, 19). Deposition of these carbonate systems initiated the high-relief shelf edges that later characterized the Eastern shelf during deposition of the Cisco Group.

Regional upwarping in the Ouachita foldbelt and the eastern flank of the Fort Worth basin was coincident with Midland basin subsidence and provided a significant supply of second-cycle sediments to Strawn deltas. The hinge or axis of rotation between the subsiding Midland basin and the gradually rising eastern Fort Worth basin defines the present Bend flexure or arch (fig. 3).

As the source areas were lowered by erosion and as paleogradients diminished, less terrigenous sediment reached the Pennsylvanian coastline. Extensive and long-lived carbonate-bank and reef systems began to form on the stable platforms provided by abandoned Strawn deltaic clastic materials (fig. 19). As the Midland basin continued to subside, many reefs or banks grew vertically to maintain necessary water depth. Trends of atoll-like limestone bodies that parallel the basin margin grew throughout much of the uppermost Pennsylvanian, but they are most common in the Canyon Group. Some of the carbonate banks, growing on structurally positive trends of the Eastern shelf, extended upslope to the present outcrop area where they intertongued with Canyon deltaic systems. Although Canyon deltas prograded extensively during three principal deltaic cycles, terrigenous clastic deposition was about equally balanced with limestone deposition.

Near the end of Canyon (Missouri Series) deposition, rejuvenation in the Ouachita foldbelt and eastern Fort Worth basin slightly increased paleogradients and significantly increased sediment supply, much of which was second-cycle detritus from earlier Atoka fan-delta facies and easternmost Strawn fluvial facies (fig. 17). As the supply of terrigenous clastic materials increased, extensive lower Cisco (Virgil Series) delta-fluvial systems began building westward across the eastern shelf, overlapping Canyon carbonate facies (fig. 20). Cisco delta systems prograded 10 to 15 times across the relatively stable eastern shelf of the Midland basin. Accelerated subsidence of the Midland basin provided as much as 455 m of relief between Cisco shelf edges and the floor of the Midland basin. Fluvial-delta systems built across the shelf and supplied sediment to thick, basinward-prograding slope-fan facies (figs. 17, 20). During deposition of the upper part of the Cisco Group (Permian, Wolfcamp Series), sediment supplied from the east again diminished, and thick, low-relief limestone shelf-edge banks became increasingly prominent.

After Cisco deposition, extensive carbonate-shelf and shelf-edge facies gradually restricted circulation on the landward parts of the eastern shelf. Minor deltaic and fluvial systems supplied fine-grained sediment that prograded the coastline locally and provided sediment to extensive tidal-flat systems. These tidal-flat systems accreted basinward and were overlapped by broad supratidal-flat (sabkha) evaporite systems.

RED RIVER ARCH AND OKLAHOMA MOUNTAINS

The complex history of the Wichita, Arbuckle, and Red River structural elements (Tomlinson and McBee, 1959) is recorded in thick clastic wedges extending southward and southwestward into the northernmost part of north-central Texas. These arkosic or "granite wash" deposits represent fluvial and fan-delta deposition along steep paleoslopes adjacent to fault blocks near the Red River in north Texas and southern Oklahoma. The fan deltas prograded basinward as a braided complex; prodelta facies and reworked fan-fringe deposits are of marine origin. Fan-delta deposition was contemporaneous with limestone deposition on adjacent, structurally positive blocks.

WEST TEXAS

Reconnaissance work in the Hueco Mountains indicates that the Magdalena Group records deposition on a shallow carbonate platform and adjacent slope. Other stratigraphic units have not been interpreted, and no environmental interpretation has yet been published for the Franklin Mountains.

The Marathon region was the site of slope and deep basinal sedimentation for most of the Paleozoic, including most of the Carboniferous. Radiolarian chert and shale of the upper part of the Caballos Novaculite probably were deposited in water depths greater than 1,000 m. Initial black mud of the Tesnus was followed by deposition of thin distal turbidites and shale that reflect progradation of a clastic wedge from east to west. Siliciclastic detritus was derived almost entirely from Llanoria (Africa and South America). Which deltaic environments are represented in addition to distal prodelta environments remains controversial.

Uplift of the western margin of the geosyncline initiated an episode of calcareous flysch deposition recorded in the Dimple Formation. Sediments derived from carbonate banks on the shelf and from uplifted older rocks were transported down a slope and into the basin as slides, debris flows, and turbidity currents. Pelagic calcilutite, shale, and spiculites are minor deposits. Unlike the Tesnus, the Dimple is thickest in the center of the geosyncline, rather than toward its source.

Renewed uplift of Llanoria brought a return to siliciclastic flysch deposition of the Haymond Formation. Initial deposits of the Haymond, like those of the Tesnus, consisted of black mud, which was followed by alternating turbidites and pelagites that make up the bulk of the formation. Turbidites increased slightly in thickness prior to the deposition of the slide, debris-flow slump, and turbidite beds of the chaotic boulder-bed member, but the wildflysch deposition apparently began abruptly because of the sharp base of the boulder-bed unit. Turbidite-pelagic deposition continued after formation of the wildflysch unit, but the sandstone beds locally are burrowed internally, suggesting that the geosyncline was getting shallower. At the northernmost outcrops, the Haymond passes upward into shelf and slope deposits of the Gaptank Formation.

POST-CARBONIFEROUS EVENTS

CENTRAL AND NORTH-CENTRAL TEXAS

Progressively restricted depositional environments existed along the margins of the Midland and Palo Duro basins (fig. 3) throughout the Permian. Tidal-flat, sabkha, and evaporite deposits predominated. Locally, fluvial and deltaic systems prograded across the marginal-marine deposits to supply fine-grained terrigenous clastic materials to the evaporite basin. As much as 2,350 m of Permian strata accumulated in the Midland basin; 1,200 m crops out on the eastern side of the basin.

Triassic rocks of the Dockum Group are inferred to overlie unconformably Permian rocks exposed in the vicinity of Palo Duro Canyon in the Texas High Plains, but to the south, the Triassic strata apparently are conformable with the Permian (J. H. McGowen, oral commun., 1977). The Dockum is inferred to be Late Triassic on the basis of vertebrate remains, but underlying strata assumed to be Permian are unfossiliferous, and no lithologic break has been observed in outcrop. Dockum deposition was principally lacustrine, centered within the relict Midland and Palo Duro basins. Locally, fluvial and deltaic systems prograded into the Triassic lakes. As much as 667 m of Triassic sediments was deposited; 330 m of Triassic rocks is exposed in outcrop.

After deposition of the Triassic Dockum Group, all of central, north-central, and west Texas underwent Jurassic and Early Cretaceous erosion that coincided with subsidence and seaward tilting of the Gulf Coast basin. Geomorphic patterns—shale valleys, sandstone hills, and limestone cuestas—and erosional relief that formed in central and north-central Texas at that time were similar to modern topography in the region. Headward eroding, eastward-flowing rivers supplied sediment derived from the exposed Triassic and upper Paleozoic rocks to delta and barrier-island systems along a transgressive Cretaceous shoreline.

Basal Cretaceous deposits in central and north-central Texas are quartzose and calcareous sand and gravel, principally of fluvial and shoreline origin. Continued westward onlap of Cretaceous marine environments out of the east Texas basin resulted in deposition of calcareous shale and limestone.

Eventually, all of central, north-central, and west Texas was submergent, and a major reef-lagoon system formed on the structurally high Llano uplift and Concho arch (fig. 3). Shallow-marine environments existed east and north of the reef tract.

Gradual uplift and erosion followed Cretaceous deposition in central and north-central Texas. No major Tertiary tectonic events took place in the region. Drainage similar to present-day patterns was established initially about Eocene time (O. T. Hayward, oral commun., 1977).

Major alluviation in central, north-central, and west Texas during Late Miocene and Pliocene time resulted in deposition of the Ogallala Formation. Aggrading fluvial systems originating in the Rocky Mountains deposited sand and gravel in a thick coalescing alluvial plain that ultimately stretched from Texas to South Dakota. In Texas, the alluvial deposits extended eastward over Triassic and upper Paleozoic rocks to about 144 km east of the present caprock escarpment (O. T. Hayward, oral commun., 1977). At the same time, valley-fill fluvial systems (Uvalde gravels) extended westward across lower Tertiary coastal deposits to upper Miocene and Pliocene Gulf shorelines.

During the Pleistocene, much of the distal part of the Ogallala alluvial plain was eroded, producing the Caprock escarpment. Headward erosion of the Pecos River system isolated the Ogallala alluvial plain in Texas from its source area in New Mexico (Thomas, 1972). Several episodes of Pleistocene alluviation followed, during which the Seymour Formation and other high gravels, derived predominantly from the Ogallala Formation and nearby Cretaceous exposures, were deposited by gulfward-flowing streams. Subsequent regional uplift, erosion, and stream piracy have isolated the Seymour and the other gravel deposits along drainage divides that are as much as 52 m higher than present-day drainage (Epps, 1973).

WEST TEXAS

In the Marathon region, the Paleozoic sequence underwent an episode of mountain building by compressional deformation during Late Pennsylvanian and early Wolfcamp (Ross, 1962) that caused the formation of folds, thrust faults, and strike-slip faults. A thick sequence of late Wolfcampian and Leonardian clastic and carbonate strata and Guadalupian and Ochoan carbonate strata, which were deposited over parts of the area, were subsequently tilted northeastward and eroded. Shelf carbonate rocks of Cretaceous age but of unknown thickness blanketed the area, and they, in turn, were arched and partly eroded during and following domal uplift in Tertiary time.

The Franklin Mountains and Hueco Mountains apparently underwent faulting during Laramide tectonism, but the main uplift of the Franklin Mountains began in late Cenozoic during formation of the Rio Grande rift. Some faults are still active. Quaternary alluvial fans flank both the Franklin and Hueco Mountains today.

ECONOMIC PRODUCTS

Carboniferous rocks in central, north-central, and west-central Texas have contributed significantly to the economic development of Texas and of the Nation. Oil and gas production has been predominant but industrial and ceramic clay, coal, and constructional limestone historically have been locally and periodically important. Uranium may occur within Carboniferous rocks in sufficient quantities to warrant further intensive exploration. Ground-water potential from Carboniferous rocks is poor, and ground water has not been significantly exploited. No resources of economic value have been recognized in the Marathon basin or in the Franklin and Hueco Mountains.

OIL AND GAS

Carboniferous rocks in Texas have undergone 60 years of intensive exploration for and production of petroleum. Earliest discoveries and principal production from Carboniferous rocks took place in north- and west-central Texas on the eastern shelf of the Midland basin, along the Red River, Muenster-Electra uplifts, and in the "Horseshoe Atoll" centered in Scurry County (fig. 3). Between 1916 and 1921, earliest discoveries were in sandstone beds of the Strawn Group and in the Marble Falls Limestone (and associated facies) in fields such as Ranger, Desdemona, and Breckenridge. Farther west, Cisco sandstone reservoirs were soon discovered (fig. 22).

Carboniferous sandstone reservoirs are generally shallow (less than 1,000 m) fluvial and deltaic facies located on subtle structural closures and in stratigraphic traps. Shelf-edge Marble Falls limestone facies provide slightly deeper targets directly beneath the shallow clastic reservoir rocks. Arkosic and feldspathic reservoirs (granite wash) of fan-delta origin are productive along the uplifted fault blocks that constitute the Red River, Wichita, and

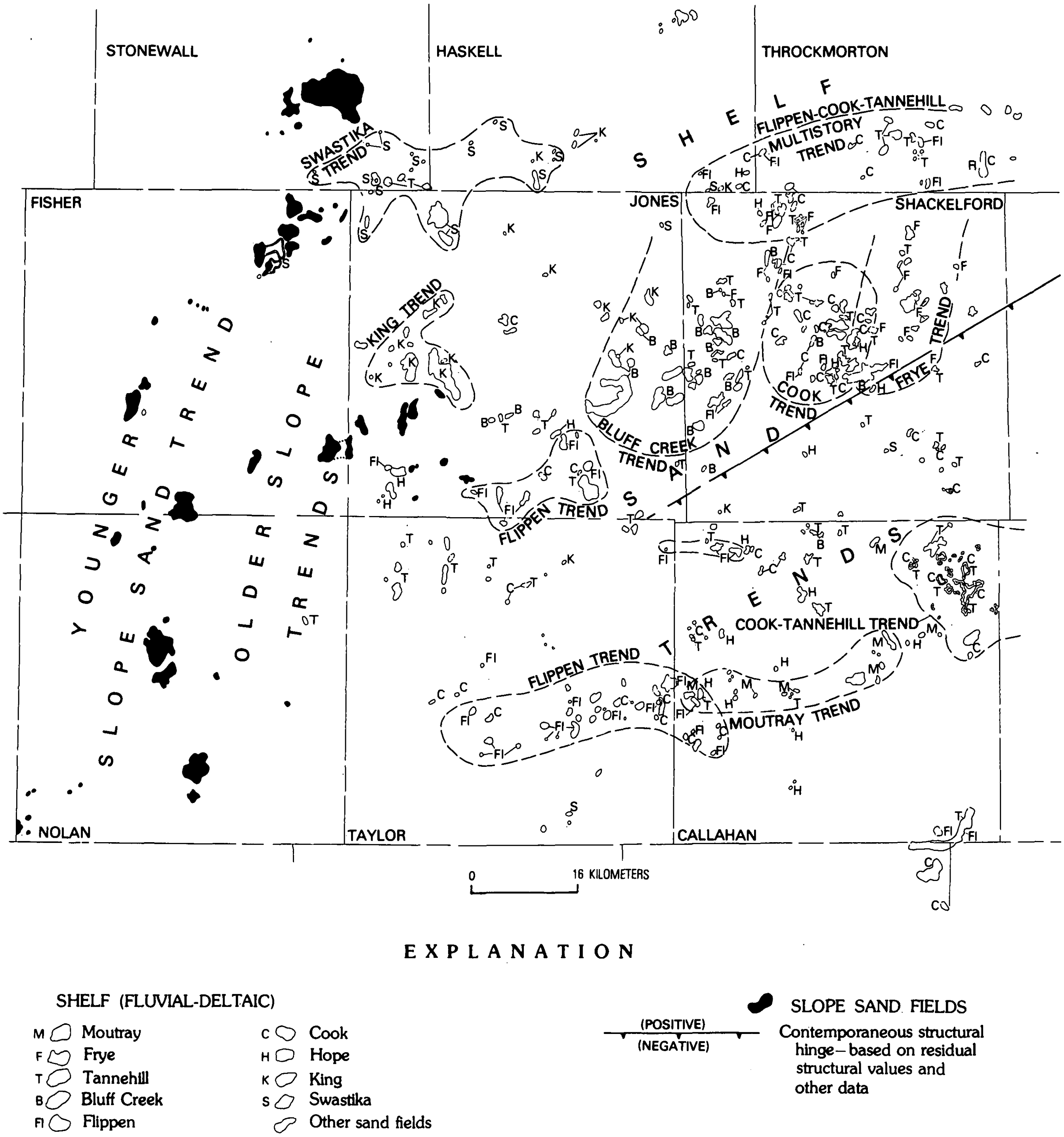

FIGURE 22.—Distribution of oil production from Cisco Group on central part of the eastern shelf, north-central Texas. Based on Abilene Geological Society data, by L. F. Brown, Jr., and W. E. Galloway. Modified from Abilene Geological Society (1960).

Amarillo uplifts (fig. 3). Most shallow structures were located by surface mapping. Shallow structures supported more than 40 years of intensive wildcatting, and today the oil province is again the site of active exploration for marginal oil and gas left behind in the small lenticular reservoirs.

After World War II, the single greatest discovery in Texas was the Horseshoe Atoll, an extensive complex carbonate system west of the Eastern shelf within the Midland basin (fig. 3). This discovery of Strawn, Canyon, and Cisco limestone reservoirs triggered extensive exploration of Pennsylvanian and Permian shelf edges during the 1950's. Many similar Strawn and Canyon limestone bank (or reef?) reservoirs were discovered trending approximately northeast along the eastern margin of the Midland basin. Some sandstone reservoirs of deep-water, submarine-fan origin were also discovered within the Midland basin adjacent to the edge of the relict Upper Pennsylvanian eastern shelf.

Principally gas has been discovered in the thick Lower to Middle Pennsylvanian fan-delta and deltaic sandstones and conglomerates in the Fort Worth and Sherman basins (fig. 3). Intensive exploration is currently underway in the Kerr and Val Verde basins (fig. 3), which are southwestern extensions of the foreland basin system separating the Texas craton and the Ouachita foldbelt.

Production from Pennsylvanian carbonate rocks takes place on the central basin platform. Deltaic sandstone beds of Early Pennsylvanian age were discovered recently on the northwest shelf in nearby eastern New Mexico. Localized oil fields produce from small isolated Mississippian reefs(?) found along the eastern flank of the broad Concho platform, the precursor of the later Eastern shelf.

COAL

In the last two decades of the 19th century, coal production from cyclic Pennsylvanian fluvial and deltaic facies in north-central Texas was a regional industry. Coal was used principally within the region for heating, although much of the Strawn coal was used until about 1920 as boiler fuel on locomotives of the Texas and Pacific Railroad. Mining in the region terminated in 1943.

Pennsylvanian coal in north-central Texas (fig. 23) is principally in the Mingus Formation, Strawn Group, and in the Harpersville Formation, Cisco Group (Mapel, 1967; Evans, 1974). Coal is found also in the Canyon Group, but it is very limited in distribution. Atoka coal deposits are only in the subsurface near the Red River arch.

The Thurber coal within the Strawn Group was the most economic deposit in the region, although several other coals were mined. The Thurber coal crops out beneath thick Brazos River Sandstone for about 19–24 km in southwestern Palo Pinto and Erath Counties (fig. 2). Shallow mining was concentrated on dip slopes near Thurber and Mingus along the Erath-Palo Pinto County line. The coal is about 1 m thick and was deposited within marshes and swamps bounding the Thurber embayment (Cleaves, 1975). Strawn coals rank as high-volatile bituminous with 2–8 percent moisture, 10–25 percent ash, 1.5–4 percent sulfur, and Btu (dry basis) from 10,390 to 13,755 per pound.

Several coal beds in the Harpersville Formation, Cisco Group, are found near the Pennsylvanian-Permian boundary (on the basis of fusulinids). Several lenticular coal beds as much as 1 m thick were deposited along shorelines bounding large interdeltaic embayments (Brown, 1973b). Harpersville coal production was concentrated near Newcastle in Young County, near Crystal Falls in Stephens County, near Cisco in Eastland County, and in McCulloch and Coleman Counties (fig. 2). Various local names have been applied; the Newcastle coal was the most productive in the Cisco Group. Cisco coals are variable and have a moisture content of 8–18 percent, ash content about 15 percent, sulfur from 1.1 to 8.9 percent, and heating value (dry basis) from 5,669 to 7,054 Kcal per kilogram.

Limited possibilities exist for future stripping and for possible in situ gasification of Texas subbituminous coals. Metal byproducts and special uses in cement offer other possibilities for use of Pennsylvanian coal resources.

CLAY PRODUCTS

Common industrial clays mined from Middle and Upper Pennsylvanian strata are used extensively in the brick, tile, clay pipe, and expanded aggregate industry in north-central Texas (Brown, 1958). The industry is concentrated west of the Dallas-Fort Worth metroplex near Mineral Wells, Palo Pinto County. Five plants produce industrial clay products from prodelta facies that are associated with several Strawn delta systems. Principally illitic in composition, the clays are not of ceramic quality.

Locally near Cisco, Eastland County, and Cross Cut, Brown County (fig. 2), Cisco deposits such as the Quinn Clay (Plummer and Bradley, 1949) contain sufficient kaolinite to be used for cast items and glazed tile. Large volumes of expanded aggregate are produced from Canyon and Cisco prodelta-marine clays near Ranger and Eastland in Eastland County, respectively (fig. 2).

CONSTRUCTIONAL LIMESTONE

Limestone deposits of the Canyon Group are extensively quarried for use as aggregate in concrete and as base material for highways and airport runways. The largest operation is at Bridgeport in Wise

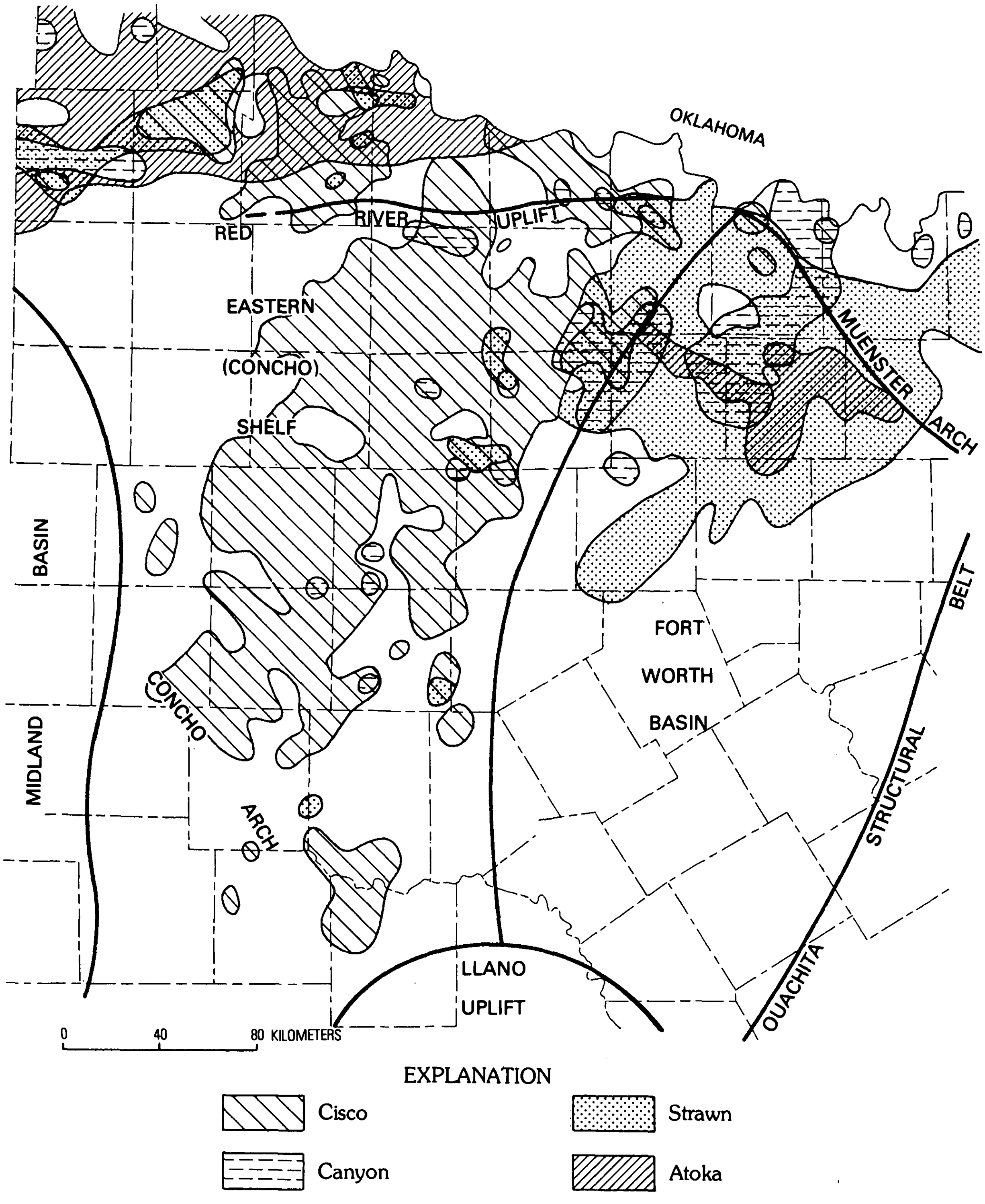

FIGURE 23.—Distribution of Pennsylvanian coal deposits, north-central Texas. From Mapel (1967).

County (fig. 2), where 200 feet of Canyon limestone bank facies, called the "Chico ridge or bank," supplies most industrial limestone for the Dallas-Fort Worth region. Other large quarries in both Strawn and Canyon limestones operate at Brownwood in Brown County, along Interstate Highway 20 in Parker, Palo Pinto, and Eastland Counties, and along other Texas highways.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— South Dakota

By ROBERT A. SCHOON

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1110-T

Prepared in cooperation with the
South Dakota Geological Survey

Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in South Dakota

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1979

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—SOUTH DAKOTA

By ROBERT A. SCHOON[1]

ABSTRACT

Carboniferous rocks are widely distributed in the subsurface of South Dakota, but surface exposures are limited to approximately 3,900 km² (1,500 mi²) in the Black Hills area. In the northern Black Hills, Tertiary intrusive rocks cut and mineralized the Mississippian Madison Formation. These mineralized zones were responsible for the gold rush of 1876 to the northern Black Hills. Economic development of the Black Hills predated classic studies of the area (1899-1907) and South Dakota's admission to the Union (1889). As a result, production records were poorly preserved or never existed.

At present, commercial use is made of the following resources: oil (Pennsylvanian rocks), water (Mississippian rocks), scenery that attracts tourism (caves in Mississippian rocks), and road metal and ballast (Mississippian rocks). Mississippian rocks appear to have the greatest potential resources for future economic development; the resources are water, oil in deeper parts of the Williston basin, and geothermal energy in the south-central part of the State.

INTRODUCTION

In the Black Hills region of South Dakota, Carboniferous rocks comprise the upper part of the Devonian-Mississippian Englewood Limestone, the Mississippian Madison (Pahasapa) Limestone, and the Pennsylvanian part of the Minnelusa Formation. Basinward, to the northeast and to the east, these Carboniferous units increase in thickness and are readily divided into eight formations. These formations are, in ascending order: Mississippian – Lodgepole, Mission Canyon, and Charles Formations (collectively known as the Madison Group); Pennsylvanian – Fairbank, Reclamation, Roundtop, Hayden, and Wendover-Meek Formations as defined by Condra, Reed, and Scherer, (1940) and McCauley, (1956). North of the Black Hills area, the Lower Pennsylvanian sedimentary rocks are not as easily divided. Here it is a common, but not recognized, practice to use two terms, the Amsden (Lower Pennsylvanian) and the Minnelusa Formations to include rocks of Pennsylvanian and Permian age.

The Carboniferous rocks crop out on the periphery of the Black Hills (fig. 1). On the geologic map (fig. 2), the outcrop pattern forms a band encircling the Black Hills. This band varies in width from an average of 4.8 km (3 mi) on the east side of the Black Hills to 26 km (16 mi) on the west side. Carboniferous sedimentary rocks crop out in a small area near Bear Butte (secs. 17, 18, 19, and 20, Tps. 6 N., R. 6 E.). No other surface exposures of Carboniferous rocks are found in the State; however, Carboniferous rocks are widely distributed in the subsurface (fig. 3).

During the Carboniferous, no known igneous activity took place in South Dakota.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the South Dakota Geological Survey.

CENOZOIC EVENTS AND SUBSEQUENT WEATHERING

According to Gries and Tullis (1955, p. 35), the close of the Mesozoic Era showed no evidence of impending Cenozoic uplift of the Black Hills area. They also stated; "White River (Middle Oligocene) beds rest on rocks as old as pre-Cambrian in the central Hills, so uplift, concurrent dissection and removal of more than 6,500 feet of sediments had been accomplished by that time." Plumley (1948, p. 536) suggested smaller scale uplifts during Miocene or Pliocene and at the beginning of the Pleistocene. Geologic history of the area is imperfectly known, but sufficient data exist to show that Carboniferous rocks were subjected to subaerial erosion throughout most of the Cenozoic Era.

In the northern Black Hills, Tertiary igneous intrusions (fig. 2) accompanied the early uplift of the area. These intrusive rocks cut Paleozoic rock and, according to Gries and Tullis (1955, p. 35), in some places extend to Upper Cretaceous rocks. Important metallic mineral deposition in Carboniferous rocks resulted from this intrusive activity.

DRAINAGE AND AREAS OF BEST EXPOSURES

The best Carboniferous exposures generally are found

[1] South Dakota Geological Survey, Vermillion, S. Dak. 57069.

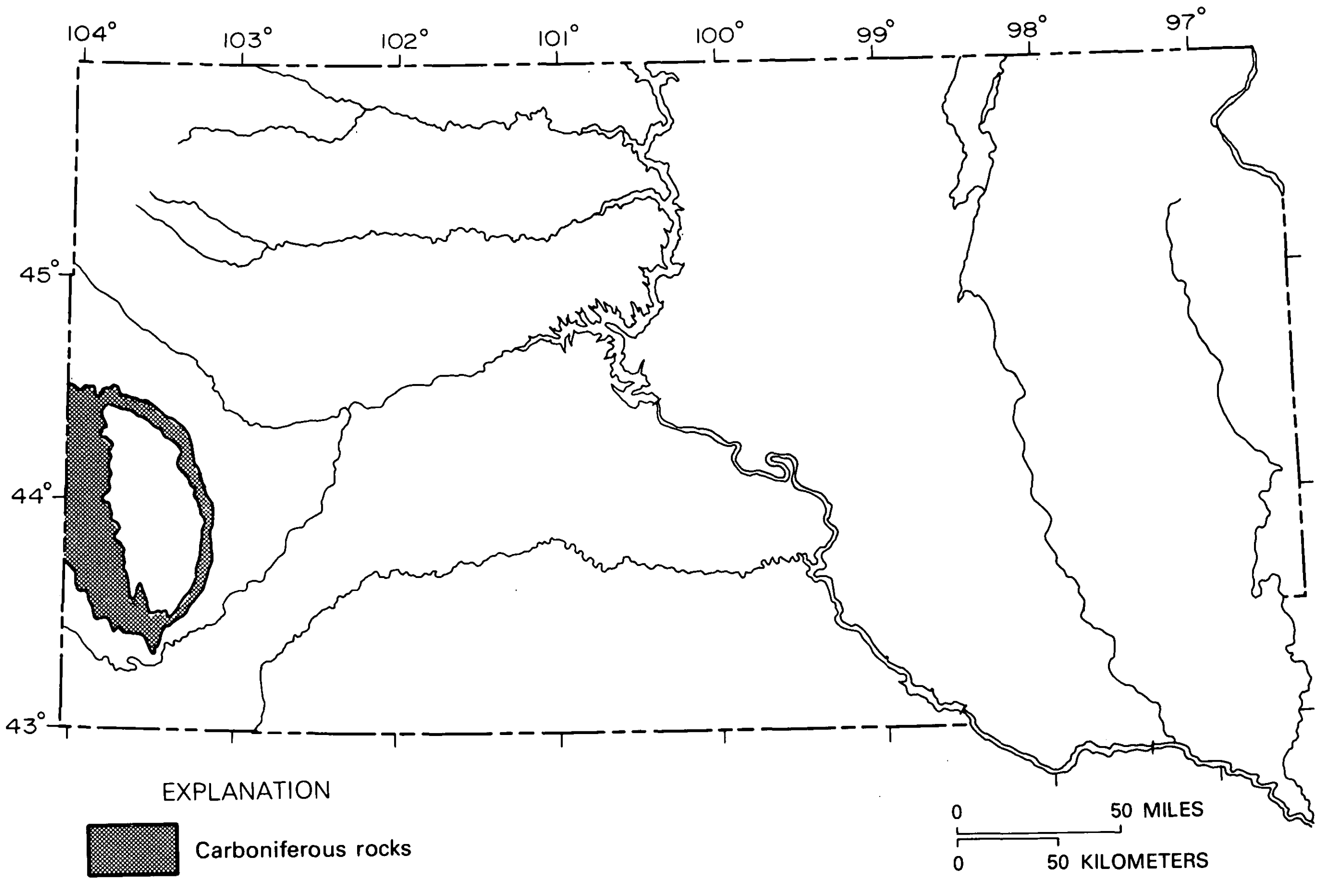

FIGURE 1.—Carboniferous exposures in South Dakota.

adjacent to the well-developed radial drainage courses of the Black Hills area and on the west side of the Black Hills where Mississippian rocks form a broad high plateau. Many caves and sinkholes in the Carboniferous rocks (Tullis and Gries, 1938, pp. 233–271, and Brown, 1944, p. 8) partly circumvent normal drainage routes. This diversion was demonstrated by a dye test conducted by Rahn and Gries (1973, p. 12), who stated, "It shows that water that disappears into sinkholes in one drainage basin may reappear as springs . . . in a completely different drainage area." Thus, measurable amounts of surface-water runoff do not follow surface drainage patterns in the area of Carboniferous exposures.

HISTORY

The Englewood Limestone named by Darton (1901, p. 509) for exposures along the Chicago, Burlington and Quincy Railroad cut 2 miles south of Englewood, Lawrence County, S. Dak., has since been shown to be equivalent to the lowermost Lodgepole Formation (Andrichuk, 1955). Klapper and Furnish (1962, p. 2071) examined conodonts from the Englewood and found an upper Devonian fauna and a lower Mississippian fauna in the middle part of the formation.

In 1901, Darton (p. 509) used the Pahasapa to define a massive gray limestone in the Black Hills region that ranged from 76 to 152 m (250 to 500 ft) in thickness. Earlier, Peale (1893, pp. 33–39) used the term Madison to define a 381-m (1,250-ft) thickness of limestone in the Madison Valley in southwest Montana. Thus, the term Madison is preferred. Subsequently, Collier and Cathcart (1922, p. 173), working in the Little Rocky Mountain region of Montana, divided the Madison into two formations, the Lodgepole Limestone (lower) and the Mission Canyon Limestone (upper). Seager (1942, p. 1420) used the term Charles Formation, and in 1952 Sloss (p. 67) stated, "Dolomitic and brecciated zones high in the Pahasapa section of the Black Hills are probably Charles equivalents."

In 1875, Winchell (p. 38) used the term Minnelusa to define approximately 23 m (75 ft) of "* * * nearly white, crystalline, subsaccharoidal sandstone * * *." Jagger (1901, pp. 178–181) expanded the term to include 183 m (600 ft) of sedimentary rocks in the northern Black Hills. Darton (1901, p. 500) used the term in the same manner as Jagger and considered the Minnelusa to be of Pennsylvanian age. In 1956, V. T. McCauley showed that the Minnelusa of the Black Hills is equivalent to the Hart-

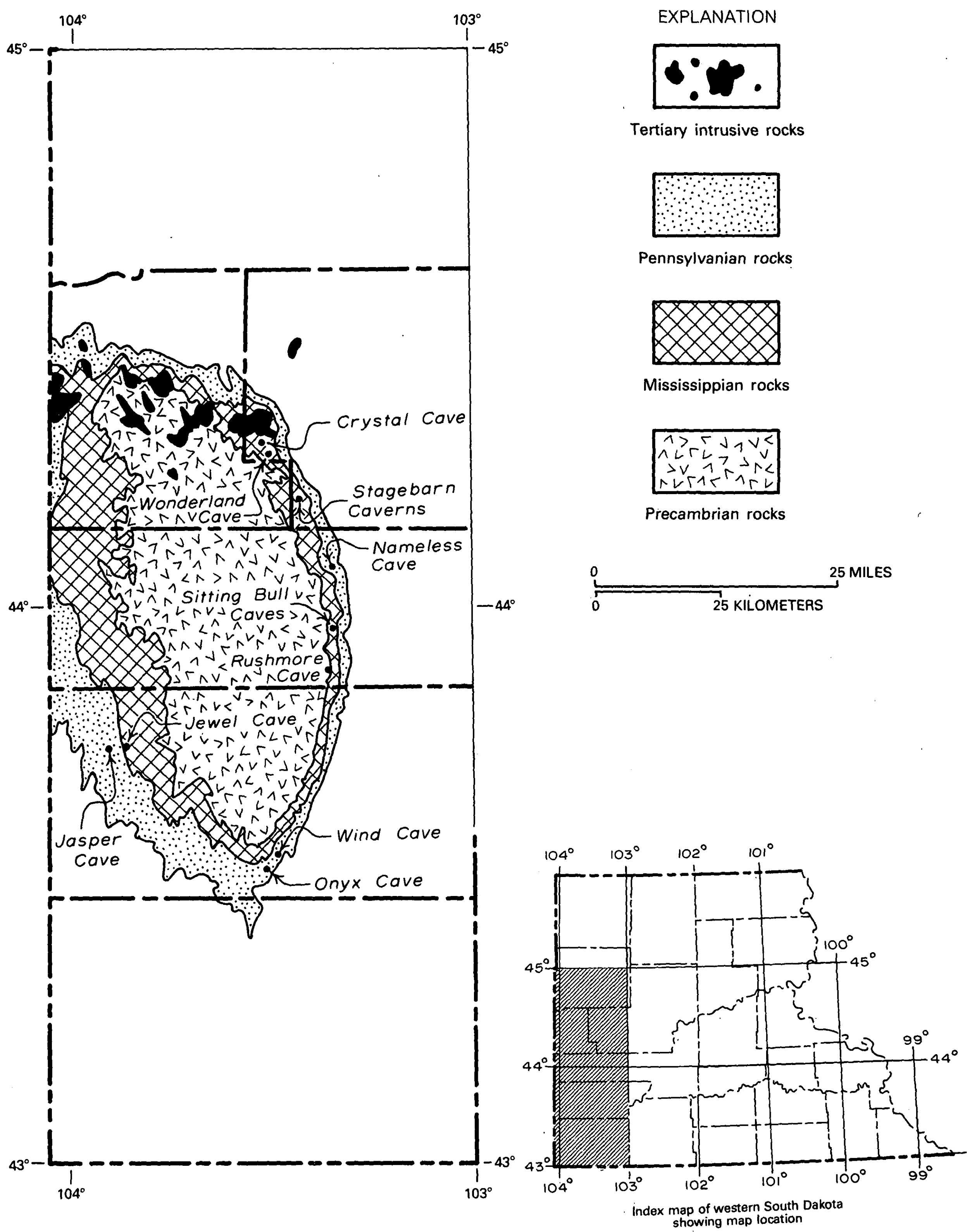

FIGURE 2.—General surface exposures of Carboniferous rocks and location of caves.

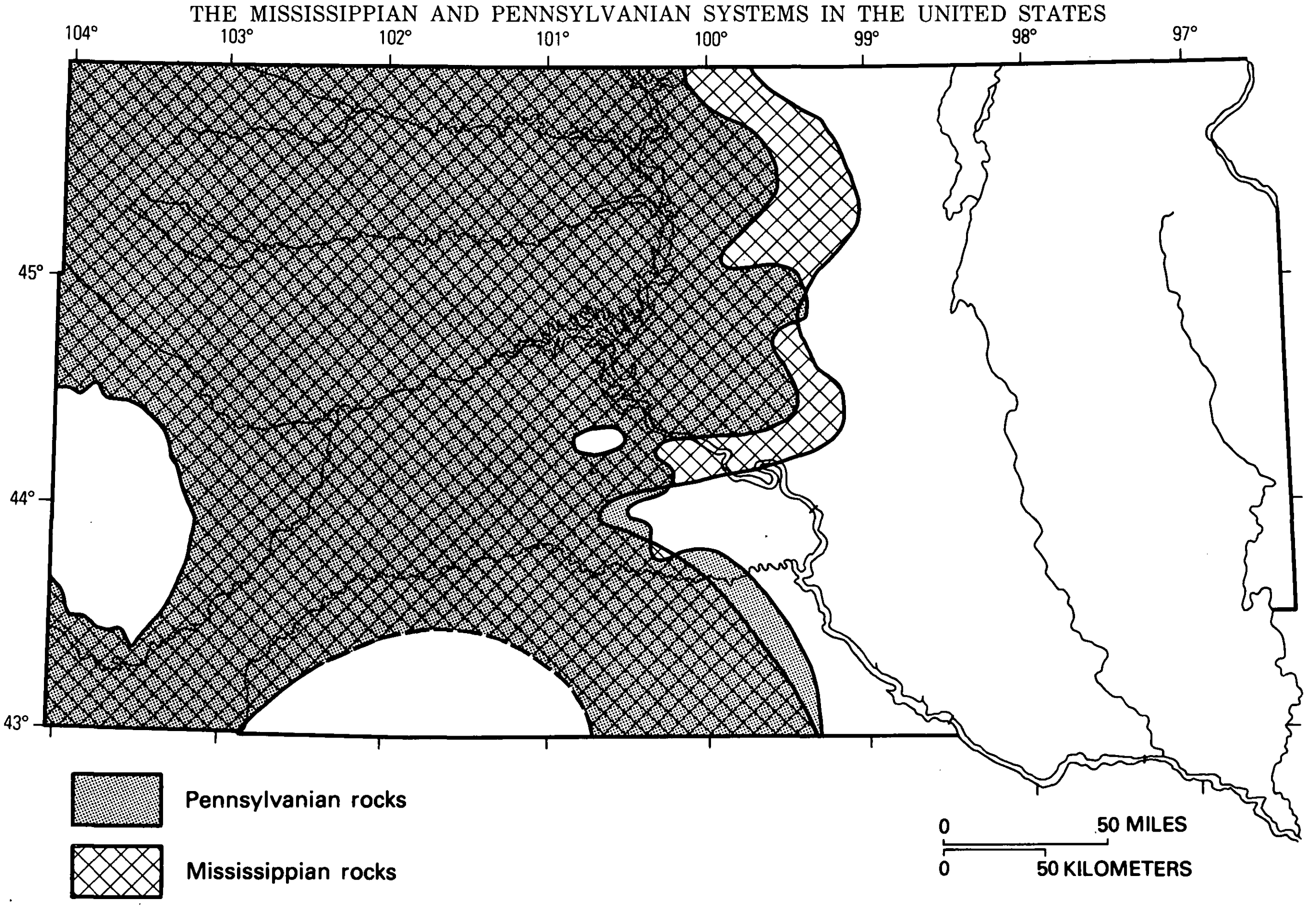

FIGURE 3.—Subsurface distribution of Carboniferous rocks in South Dakota.

ville Group of eastern Wyoming and that sedimentary rocks above the Wendover-Meek are of Permian age. The South Dakota Geological Survey recognizes the equivalency of the terms Minnelusa Formation and the Hartville Formation. Because the term Minnelusa has precedence, it is retained, and the subdivisions of the Hartville are used.

GEOLOGIC SETTING

The Englewood Formation in the northern Black Hills unconformably overlies the Ordovician Whitewood Limestone. Southward, the rocks underlying the Englewood are progressively older. The Devonian-Mississippian contact within the Englewood is transitional (Jack Kume, oral commun., 1978). South of the Black Hills, the Amerada No. 1 Moody oil test (NW¼ NE¼ sec. 12, T. 12 S., R. 6 E.) penetrated Carboniferous rocks at a depth of 1,308 m (4,290 ft) overlying Precambrian rocks at 1,515 m (4,970 ft).

Agatson (1954) and Jennings (1959, p. 994) suggested that the red shale ("red marker") in the upper part of the Minnelusa marked the base of the Permian in the Black Hills area. McCauley (1956, p. 157) also stated that the red shale "marker" was at the top of the Wendover-Meek and demonstrated that the Permian-Pennsylvanian contact is unconformable.

STRUCTURAL EVENTS DURING THE CARBONIFEROUS

Structural events during the Mississippian and Pennsylvanian Periods are not well known. Sedimentary rocks present in the extreme northwestern part of the State (Big Snowy Group) and an undetermined thickness of the Charles Formation in deeper parts of the Williston basin are absent in the Black Hills. This suggests sufficient time for considerable erosion of the Madison in the Black Hills area. Gries and Tullis (1955, p. 33) stated that, although isopach studies show a wide local variation in thickness of Mississippian rocks, field evidence indicates an erosion surface of very little relief. Thus, at or near the close of the Mississippian Period, the Black Hills area was only slightly uplifted.

The beginning of the Pennsylvanian was accompanied by downwarping in the Black Hills area. This downwarping was most pronounced in the extreme southwestern part of the State, where Pennsylvanian sedimentary rocks are more than 213 m (700 ft) thick. North of the Black Hills, the same interval is 61 m (200 ft) thick

(Amerada No. 1 State oil test, sec. 4, T. 14 N., R. 4 E.). Isopach maps show general thinning of Pennsylvanian sediments north of the Black Hills region. These two localities may be on opposite sides of the "* * * two nearly flat-topped domes or uplifted blocks" described by Noble (1952, p. 31). According to Noble, the eastern north-trending block has been more highly elevated than the western northwest-trending block; this position suggests that the two blocks composing the Black Hills uplift probably moved independently. However, isopach maps indicate that only minor structural events took place during the Permian-Cretaceous Periods.

LITHOSTRATIGRAPHY

NATURE OF MISSISSIPPIAN AND PENNSYLVANIAN CONTACT

In South Dakota, the Carboniferous is divided into eight units that define subsurface rocks. The correlation of these subsurface units with their equivalents in the Pennsylvanian outcrop areas of the Black Hills is difficult. Bates (1955, pp. 1991–1995) disclosed the reason for this difficulty in his argument that the term Minnelusa Sandstone (of Black Hills usage) is a misnomer. He also stated that the Minnelusa in the type area (Rapid Creek Canyon) contains only about 58 percent sandstone and that subsurface sections contain an even lower percentage.

According to Gries and Mickelson (1964, p. 111), the upper surface of the Madison has a pre-Minnelusa karst topography and contains sinkhole breccias, cavern filling, and residual clays. Charles Baker (unpub. date, 1952) stated "Cuts along Highway 16 in the vicinity of Jewel Cave National Monument show numerous examples of the collapse and downfolding of the basal Minnelusa red laterite into solution sinkholes and caverns of the underlying Madison Limestone." If the term "collapse" is correctly used, then at least some of the sinkholes formed after the basal Minnelusa sediments were deposited on top of the Madison. This may also explain the uncontrollable lost circulation zones encountered in drilling for oil in deeper parts of the Williston basin. A few of these zones are in areas where the timespan was relatively short between Mississippian and Pennsylvanian deposition; this suggests some solution activity after the Pennsylvanian Period began.

PRINCIPAL LITHOLOGIES

The Upper Devonian-Lower Mississippian Englewood Limestone exposed in the Williston basin (Sandberg 1962, fig. 6). However, Sandberg (p. 56) recognized an angular unconformity between the Englewood and Madison Limestones in the Black Hills area. Kume (1963, pp. 23–25) divided the formation in the Deadwood junkyard section into three units, a lower unit 7.6 m (25 ft) thick of gray silty shale, a middle unit 7.3 m (24 ft) thick of grayish-red, purple, argillaceous, and partly shaly limestone, and an upper unit 1.5 m (5 ft) thick of yellowish-gray dolomitic limestone. Elsewhere, in the northern Black Hills near the city of Deadwood, Andrichuk (1955, p. 2176) described the Englewood as dark-gray to dark-purple-gray shale containing Mississippian graptolites.

The Madison (Pahasapa) Limestone (formerly called Gray Limestone) crops out on the periphery of the Black Hills, and, according to Gries (1952, p. 71), it consists of 91.5 to 192 m (300 to 630 ft) of medium-grained crystalline, light-gray to buff limestone or dolomitic limestone. A short distance southeast of the Black Hills, the Madison thins to a feather edge on the Precambrian surface. (See fig. 3.) The term Pahasapa is the Sioux Indian name for the Black Hills; it was used by Darton (1901, p. 509) for exposures in the Black Hills. The type of locality was not designated.

The term Minnelusa was used by Winchell (1875, p. 38) to define 22.9 m (75 ft) of " * * * nearly white, crystalline, subsaccharoidal sandstone, coarsely granular when weathered and hard; has somewhat the aspect of crinoidal limestone but without crinoid stems." Succeeding authors expanded the term Minnelusa Sandstone to include approximately 395 m (1,300 ft) of sedimentary rocks in the southwestern part of the State that consists of rocks of Permian and Pennsylvanian age. Reed (1955, p. 46), McCauley (1956, p. 157), and Jennings (1959, p. 994) stated that the top of the Pennsylvanian rocks was marked by the "Red Shale" unit of the uppermost Wendover-Meek Formation.

The principal lithololgies of Pennsylvanian sedimentary rocks appear to be better known in subsurface sections than in outcrops in the Black Hills. In the Sun Lance No. 1 Nelson oil test (NE¼ SE¼ sec. 21, T. 7 S., R. 1 E.), the basal Fairbank Sandstone is composed of 10.7 m (35 ft) of silty and sandy, plastic, variegated red, gray, green, and purple shale at depths of 900.7 to 911.4 m (2,955 to 2,990 ft). From 875.4 to 900.7 m (2,872 to 2,955 ft) is the Reclamation Formation, which is light-tan, fine crystalline to sublithographic limestone containing interbedded green and red plastic shale. The Roundtop Formation, from 804.7 to 875.4 m (2,640 to 2,872 ft), is composed of white to gray fine- to medium-grained sandstone, interbedded with light-colored sublithographic limestone and red and green shale. The overlying Hayden Formation, from 743.7 to 804.4 m (2,440 to 2,640 ft) has 12.2 m (40 ft) of white, fine- to medium-grained well-sorted, subrounded, friable sandstone at the top. Below this sandstone, the formation is predominantly tan to gray fine crystalline dolomite interbedded with splintery, black, highly radioactive shale. From 691.9 to 743.7 m (2,270 to 2,440 ft) is the

Wendover-Meek Formation. The top 6 m (20 ft) is soft, plastic to flaky, brick-red shale. This shale unit is known as the "red marker" bed that is commonly accepted as the top of the Pennsylvanian sedimentary rocks. Underlying the "red marker" is white, light- to-dark-gray, fine to microcystalline dolomite and intervals of dolomitic and anhydritic, fine- to medium-grained sandstone and shale.

Except for the term "red marker," subsurface terminology has not been used in outcrop studies in the Black Hills area. The reason for this may be the preponderance of sandstone in surface exposures, whereas, in subsurface sections near the Black Hills area, carbonate and evaporite deposits are more plentiful than sandstone. If these observations are correct, significant changes occur between the outcrop and subsurface sections.

EVENTS AFFECTING DEPOSITIONAL ENVIRONMENT

Gries and Tullis (1955, p. 33) suggested that,

> Sagging of the shelf in early Mississippian time permitted a readvance of the sea from the north. Locally dark gray shales were the first to be deposited, but over most of the area, carbonates, contaminated with iron-rich material from the Deadwood and Pre-Cambrian surfaces, formed the Englewood formation. As submergence continued these old surfaces were buried, and deposition of very pure carbonates followed. The shoreline probably retreated to a line north of the Black Hills before Charles time.

This relatively low-lying landmass existed in the vicinity of the northern Black Hills until nearly Middle Pennsylvanian time because the lower two formations (Fairbank and Reclamation) are absent at the Weller No. 1 Wiesman oil test (SW ¼SE¼ sec. 30, T. 7 N., R. 4 E.) owing to nondeposition. Here, only 82.3 m (270 ft) of Pennsylvanian sedimentary rocks are present. Thus, the statement by Jennings (1959, p. 987) that " * * * based on new faunal evidence * * * the greater part of the Minnelusa in the northern Black Hills is Permian in age" is corroborated by subsurface correlations, and lower Pennsylvanian rocks are not uniformly present in that area.

Farther south, between the Chadron arch (fig. 4) and the southern Black Hills, correlations of the Fairbank, Reclamation, and Roundtop Formations are difficult; however, from the extreme southwestern part of the State to the southern part of the Williston basin, correlations are easily made. This suggests that during Early Pennsylvanian time the eastern block of the Black Hills (Noble, 1952, p. 31) was elevated in relation to the western block. In addition, the northern part of the eastern block was apparently elevated more than the southern part. The early Pennsylvanian sea encroached from the north and surrounded, perhaps covered, the area of the Black Hills. The depositional environment ranged from nonmarine to nearshore marine through

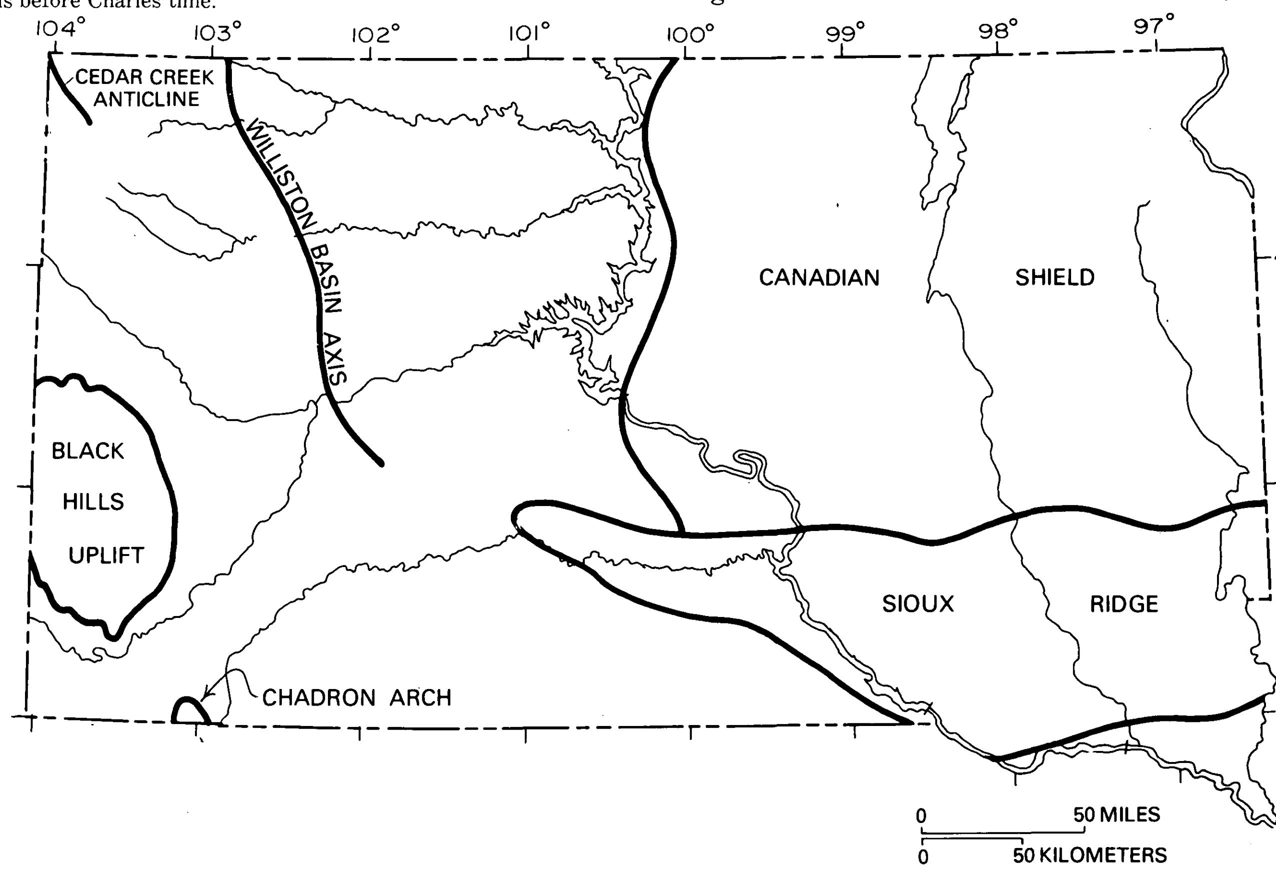

FIGURE 4.—Structural elements in South Dakota.

most of the Early Pennsylvanian Period in the area of the Black Hills. This may have prompted Jennings' conclusion (1959, p. 991), "This evidence at hand indicates that the northern and southern Hills deposits of the Minnelusa are not even grossly correlative on interval along. In addition, individual fossiliferous horizons are not consistent between the two ares [sic]."

BIOSTRATIGRAPHY

AGE OF ROCKS

A complete record of Carboniferous floral and faunal fossils is not present in the Black Hills area. The hiatus between Mississippian and Pennsylvanian rocks in this area represents a timespan equal to that required to deposit the upper part of the Mission Canyon, Charles, Kibbey, Otter, and Heath Formations of the Mississippian age and rocks of Early Pennsylvanian age that are present in deeper parts of the Williston basin. In addition, McCauley (1956, p. 153) recognized an unconformity on the Pennsylvanian surface in shallower parts of the basin. Because no sedimentary rocks are found at the top of the Pennsylvanian and the Lower Pennsylvanian sedimentary rocks are unevenly distributed over the Hills area, many individual fossiliferous beds are not consistent from the northern to the southern Black Hills areas. Further studies of the caliber of Jennings' work may solve many of the present problems in correlating Pennsylvanian rocks in the Hills area.

In 1930, Dillé collected fossils from the base of the Minnelusa in the southern Hills and concluded that the basal beds of the formation were the equivalent of the Des Moines stage of the Carboniferous. In the same year, Roth described a section at Loring Siding (also in the southern Hills) 30.5 m (100 ft) above the base of the Minnelusa. In this section, Roth recognized a fusulinid that resembled *Fusulinella euthysepta* Henbest and suggested that the lower part of the Minnelusa was deposited during the Des Moines stage. On the basis of fossils from four fossil-bearing beds near Beulah, Wyo., in the northwestern Black Hills, Brady (1931, p. 188) suggested that these forms were typical of Missourian beds of the midcontinent Pennsylvanian and that the upper part of the Minnelusa in the area was "Upper Coal Measures" in age.

TYPES OF FOSSILS

Fossils from the Pennsylvanian are generally poorly preserved. Most researchers have centered their attention on fusulinids that are present throughout the Minnelusa outcrop. Principal exposures are along the west flank of the Black Hills, along the northern flank in Bear Butte and Sand Creek Canyons, on the southern flank in Beaver Creek and Hot Brook Canyons, and in the vicinity of Loring Siding. Additional exposures are in Spearfish, Little Elk Creek, Rapid, Redbird, and Hell Canyons (Jennings, 1959, p. 986).

Fossils are common in the Madison (Pahasapa) Limestone and, according to Gries (1952, p. 71), indicate a Kinderhook-Osage age for the formation. The Englewood (lowermost Madison) contains fossil corals and brachiopods, and exposures above the Deadwood junkyard yielded the second known occurrence of Mississippian graptolites in North America (Ruedemann and Lockman, 1942). Mississippian rocks crop out around the entire Black Hills, and collecting areas exist in widespread areas.

ECONOMIC PRODUCTS

COAL

To date, no commercially important seams of coal have been discovered in Carboniferous rocks in South Dakota. The Hayden Formation appears to be the most promising prospect in the entire Carboniferous section because very carbonaceous shale is present in subsurface sections. Charles Baker (unpub. data, 1952) mentioned a ⅙-m (½-ft) coal seam in the Minnelusa exposure along Highway 16 near Jewel Cave and ½-m (1½ ft) of coal or oil shale in he Beaver Creek section.

URANIUM

The Hayden Formation is viewed with some interest by energy companies. This interest has centered on the highly radioactive black shale that occurs wherever the Hayden is present in the State. At present, no uranium is produced from rocks of Carboniferous age.

PETROLEUM

The five-well Barker Dome oil field is in sec. 34, T. 6 S., R. 1 E. and produces from the "Leo Sand" (informal term) approximately 10 m (30 ft) below the top of the Hayden Formation. This field was discovered in Custer County in 1955, and, at present, the total cumulative production is 209,326 barrels. Numerous shows have been reported from Pennsylvanian rocks in oil tests in deeper parts of the Williston basis, but the Barker Dome field is the only field in the State that produces from Pennsylvanian rocks.

Mississippian rocks do not yield oil in commercial quantities anywhere in the State. Many shows of free oil from drill-stem tests of unsuccessful oil wells in the extreme northern part of the State have been reported. Nearly all these shows were from the top of the Mission Canyon Formation just below the lowermost anhydrite bed of the Charles Formation. These reported shows appear to follow a trend, which is not accurately defined owing to the scarcity of subsurface information. It is believed that the possibilities are good for the discovery of commercially productive oil fields in this area in the near future.

METALLIC ORES

The Black Hills of South Dakota have a long history as a famous gold-producing area. Perhaps the first page of this history was inscribed on a stone by an early gold seeker. According to Dockery (1952, p. 9), Lewis and Ivan Thoen found a slab of native sandstone at the foot of Lookout Mountain near Spearfish. Inscribed on this slab was, "Came to these hills in 1833 seven of us, DeLacompt, Ezra Kind, G. W. Wood, T. Brown, R. Kent, Wm. King, Indian Crow. All ded but me Ezra Kind. Killed by Ind beyond the high hill got our gold June 1834. Got all the gold we could carry our ponys all got by the Indians. I have lost my gun and nothing to eat and Indians hunting me." The Theon stone is in the Adams Memorial Museum in Deadwood, S. Dak.

Horatio Ross, attached to Custer's expedition of 1874, panned gold in paying quantities near the present town of Custer S. Dak. The story broke in Chicago on August 24, 1874, and initiated the gold rush of 1875. According to Shapiro and Gries (1970, p. 1), the area of the Black Hills has been mined nearly continuously from 1877 to the present; total value of production is more than $75 million in gold and silver.

The Carboniferous rocks in the Black Hills area appear to have many potential host zones throughout the section, but Shapiro and Gries (1970, p. 18) stated, "No Tertiary ore deposits are known any higher in the section than the top of the Pahasapa." These Tertiary ore deposits in the Madison (Pahasapa) Limestone are concentrated in the northern Black Hills (fig. 2). Shapiro and Gries (1970, p. 40) indicated that the most important gold-silver ore-bearing Madison rocks are found in the Ragged Top mining district. Of lesser importance is a small prospect in the Madison near Galena that was reported by Irving (1904, p. 172). In the Ragged Top district, production was from replacement deposits along vertical fissures cutting the Madison Formation and from irregular masses of brecciated carbonate rock at several localities. In some places, these deposits yielded ores assaying as much as 100 oz of gold per ton.

The Madison Formation is host to lead-silver ores in the Carbonate Mining district. Shapiro and Gries (1970, p. 41) described two distinct types of ore. The first type of ore partly filled a N. 85° E. vertical fissure in the area and consisted of soft pinkish-red, ferruginous gangue reportedly containing galena, lead carbonate, and cerargyrite. This ore-bearing fissure trended east over a distance of 823 m (460 ft) and varied in width from 0.6 to 7.6 m (2 to 25 ft).

The second type of ore consisted of irregular shoots filling old solution cavities within the Madison Limestone and was intimately associated with porphyry dikes. These ores often carried very high values of silver and lesser amounts of gold, lead, and manganese.

Most of the ore deposits in the Madison Formation have been discovered by surface investigation. Future discoveries may be made by geophysical surveys, by prospect drilling, and by the study of the solution-cavity network in the Madison and (or) Minnelusa Formations.

Most of the metallic deposits in the Madison Limestone were mined out in the early 1900's. Production records have been lost or poorly recorded; thus, accurate overall figures are not available. Those that do exist are found in Shapiro and Gries (1970) or in U.S. Bureau of Mines (1954, 1955).

Metallic ores have not been reported from Carboniferous rocks in the southern Black Hills where potential host rocks are present. Post-Cambrian igneous activity is not evident in that area; therefore, mineralization of Carboniferous rocks has not taken place.

NONMETALLIC MINERALS

Pennsylvanian rocks in the Black Hills area are not used for commercial purposes. Gries (1964, p. 99) stated that quarry operations in the purer carbonate rocks of the Madison Limestone generally have been along railroad routes, and many of these quarries have produced road ballast for the railroads. At Loring Siding, two quarries supplied burned lime and limestone for use in the sugar-beet industry. At Dumont (south of Lead), the Madison Limestone was quarried for use in the Deadwood smelters, and at Pringle, the formation is used for the production of lime and various rock products.

POTENTIAL GEOTHERMAL RESOURCES

Geothermal development is in its infancy in South Dakota, but high energy costs will spur its growth. For some years, the City of Midland has used water from the Madison to heat the grade-school and high-school facilities. The hospital management at the city of Pierre is investigating the possibilities of using hot water from the Madison aquifer in its heating and sterilization systems in the hospital complex. The City of Phillip uses the Madison aquifer for a water supply, but the heated water (73°C, 160°F) is allowed to cool in Lake Waggoner prior to use.

In this south-central part of South Dakota, a geothermal gradient more than twice the worldwide average exists over an area of 36,260 km² (14,000 mi²) (Schoon and McGregor, 1971, fig. 1). The Madison Limestone is present in most of this area and yields copious amounts of fresh water. The long-range aim of the South Dakota Geological Survey is to determine the heating mechanism that is causing the abnormal gradient. When this aim is realized, water from the Madison will almost certainly be used in geothermal development.

GROUND WATER

The Madison Limestone is present in the subsurface of the western two-thirds of the State. This formation yields large amounts of water to wells. This fact attracted the attention of engineers, who have formulated plans to use large-scale withdrawals of water for industrial use. However, many cities and ranchers in western South Dakota use the Madison aquifer as a water source. In addition, the Madison aquifer recharges the Dakota artesian system in the southeastern part of the State, and thousands of wells drilled to the Dakota produce water that originates from the Madison (Schoon, 1971, fig. 19). Therefore, an accurate estimate of withdrawal or water loss from the Madison cannot be made. Any combination of withdrawal and water loss over recharge will result in depletion of water and in serious economic loss to the thousands who use the Madison aquifer and the Dakota artesian system.

Recharge to the Madison aquifer is unique in that surface-water runoff can be observed recharging the formation in the Black Hills area. This was demonstrated by Rahn and Gries (1973, fig. 6). Brown (1944, p. 19) reported that Doty Spring had ceased flowing during an extended period of drought. At this time, a man entered the spring opening and crawled several hundred feet into the fissure. The man reported that the fissure opened out into a room 30 feet high. Crooks (1968, p. 1955) concluded that, rather than a network of small solution cavities, conduitlike systems are available to transport runoff water, and that water in these conduits moves faster than the surface water. Rahn and Gries (1973, p. 41) also reported that some caves are below streambeds, and in such caves, it is possible to crawl to a point where water still flows above.

The Mississippian rocks in the greater part of the Williston basin, as in the Black Hills area, were exposed to weathering before the deposition of Pennsylvanian sediments. That these areas are also characterized by water-filled caverns at depth appears to be confirmed by a few drillers, who reported that tool strings drop 6 feet or more during the drilling in the Madison Limestone.

Owen (1898, p. 164), in describing Onyx Cave, mentioned the existence of a "cave river" of considerable volume. Crooks (1968, p. 55) is probably well within the bounds of reason in suggesting that "conduit-like networks" exist in the formation and serve to transport surface runoff to ground-water storage.

CAVERNS AND ABANDONED MINES

Speleologists and the curious tourist can find adventure in the natural caves and the abandoned mines in the Madison Limestone of the Black Hills area. In addition to the current value of these subterranean passages as tourist attractions, these caverns and abandoned mines also have potential value as storage areas and shelters in the event of nuclear attack.

All these caves have formed in the Madison Limestone. Figure 2 shows the general location of caves of the Black Hills. Highway directions are as follows: Jewel Cave is 21 km (13 mi) west of Custer on U.S. Highway 16 to Newcastle, Wyo.; Jasper Cave is 1.6 km (1 mi) west of Jewel Cave; Wind Cave is 16 km (10 mi) north of Hot Springs, S. Dak.; Onyx Cave is 4.8 km (3 mi) southwest of Wind Cave; Rushmore Cave is 6.4 km (4 mi) east of Keystone, in Hayward, S. Dak.; Sitting Bull Cave is about 19 km (12 mi) southwest of Rapid City on U.S. Highway 16; Nameless Cave is 8 km (5 mi) west of Rapid City; Wildcat Cave is a short distance west of Nameless Cave; Stage Barn Caverns are 14.8 km (3 mi) south of Piedmont, S. Dak.; Wonderland Cave is on Elk Ridge and is reached by leaving U.S. Highway 14 north of Piedmont; Crystal Cave is on the opposite side of Elk Canyon. Ice Cave is near Galena, S. Dak.; and Davenport Cave is near Sturgis, S. Dak. Descriptions of these caves have been provided by Tullis and Gries (1938).

The many abandoned mines in the Black Hills may be quite hazardous. For an accurate account of the location and general condition of these mines, see U.S. Bureau of Mines (1954, 1955).

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Wyoming

By DAVID R. LAGESON, EDWIN K. MAUGHAN, *and* WILLIAM J. SANDO

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Wyoming

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—WYOMING

By DAVID R. LAGESON[1], EDWIN K. MAUGHAN, and WILLIAM J. SANDO

ABSTRACT

Lower Carboniferous strata of Wyoming (Mississippian to early Middle Pennsylvanian) represent two marine depositional sequences separated by a period of epeirogenic uplift and erosion. The Mississippian part of the succession is a wedge that thickens northwestward from a zero edge in southeastern Wyoming; this wedge is disconformably overlain by predominantly thin-bedded terrigenous strata of the upper part of the succession.

In western Wyoming, the lower depositional sequence is represented by the Madison Group, which includes the Mission Canyon and Lodgepole Limestones. The Lodgepole includes, in ascending order, a basal dark shale (Kinderhookian), which represents the upper tongue of the Cottonwood Canyon Member; the Paine Member (Kinderhookian); and the Woodhurst Member (early Osagean). The overlying Mission Canyon Limestone is of middle Osagean to early Meramecian age. Throughout most of central Wyoming, the lower depositional sequence is represented by the Madison Limestone, which includes six members, ranging from Kinderhookian to early Meramecian. These are, in ascending order, the Cottonwood Canyon Member (conodont-bearing shale, siltstone, and silty dolomite), lower dolomite member (thick-bedded crinoidal dolomite), Woodhurst Member (cross-bedded crinoidal dolomite and limestone), cherty dolomite member (thin-bedded dolomite and dolomitic limestone), cliffy limestone member (cherty bioclastic crinoidal limestone), and the Bull Ridge Member (a lower red and yellow shale and siltstone and an upper stromatolitic limestone and dolomite). In the Hartville uplift area of eastern Wyoming, the lower depositional sequence is represented by the Guernsey Formation, which consists of limestone and dolomitic limestone unconformably lying on Precambrian, Cambrian, and Ordovician rocks. Strata equivalent to the Madison Limestone in the Black Hills include the lower Mississippian Englewood Formation (red and purple dolomite, limestone, and shale) and the Kinderhookian-to-Osagean Pahasapa Limestone.

The upper depositional sequence of lower Carboniferous strata in Wyoming consists of three (four in western Wyoming) transgressive facies, which are members of the Amsden Formation. The Darwin Sandstone Member (late Meramecian to late Chesterian), the basal transgressive unit, rests disconformably on the Bull Ridge Member of the Madison. The Horseshoe Shale Member conformably overlies the Darwin and consists of a late Chesterian to Early Pennsylvanian succession of transgressive red beds. The Moffat Trail Limestone of western Wyoming is a transgressive wedge-shaped body of bioclastic limestone between the Horseshoe Shale and Ranchester Limestone Members of the Amsden Formation. In an alternative to the view expressed sequence is the Ranchester Limestone (late Chesterian-to-Morrowan in western Wyoming and Morrowan-to-Atokan in central Wyoming), which consists of interbedded dolomite, limestone, sandstone, and shale.

Two points of view are presented concerning the stratigraphic relations and paleontologic interpretation of the Amsden Formation. In an alternative to the view expressed above, the Pennsylvanian rocks in Wyoming comprise two stratigraphic sequences bounded by regional unconformities. Red mudstone and limestone characterize strata in the lower sequence, and sandstone and dolomite characterize strata in the upper sequence. Initial deposits of the lower sequence are red mudstone probably no older than about middle Morrowan, except in the subsurface of southwestern Wyoming, where early Morrowan strata may be present. The basal red mudstone lies unconformably upon Mississippian strata, except in the vicinity of the ancestral Rocky Mountain Front Range, where it lies upon Precambrian rocks. The red beds grade upward to interbedded limestone and mudstone of late Morrowan and early Atokan age, but the limestone strata are absent at many places because of beveling on a Middle Pennsylvanian erosional surface.

Initial deposits of the upper sequence are heterogeneous and locally may be thinly bedded red, green, or varicolored mudstone, sandstone, or dolomite of late Atokan to early Des Moinesian age. Basal strata of the upper sequence lie unconformably upon lower Pennsylvanian strata at most places in Wyoming; however, locally, where the lower Pennsylvanian strata are absent because of erosion or possibly because of nondeposition, the upper sequence lies upon Mississippian rocks. The upper sequence in central and northwestern Wyoming is mostly sandstone no younger than Des Moinesian. The upper sequence in eastern Wyoming consists of sandstone and dolomite about equally and also includes Upper Pennsylvanian strata as young as Virgilian. Upper Pennsylvanian strata are also part of the upper sequence in the thrust belt in western Wyoming. Pennsylvanian strata terminate upward at an erosional unconformity beneath Permian rocks everywhere in Wyoming.

The principal economic products from Carboniferous rocks in Wyoming are petroleum and high-calcium industrial limestone. Other commodities of minor importance include building stone and aggregate.

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INTRODUCTION

By David R. Lageson

Carboniferous strata in Wyoming (Mississippian and Pennslyvanian) include the Madison, Amsden, and Tensleep Formations and their equivalent units. This report is divided into two main parts dealing with stratigraphy and a third shorter part on economic products. The first part on stratigraphy deals with strata equivalent to and including the Mississippian Madison Limestone; the other discusses the Pennsylvanian Tensleep Formation and equivalent strata. Two points of view are presented concerning the stratigraphy and age relations of the Amsden Formation .We hope that this two-part discussion of stratigraphy will clearly define the age and stratigraphic relationships of the units involved.

The Madison Limestone thickens towards western Wyoming, where it attains group status and includes the Lodgepole and Mission Canyon Limestones. Throughout central Wyoming, the Madison has been divided into six members; these are, in ascending order, the Cottonwood Canyon Member, lower dolomite member, Woodhurst Member, cherty dolomite member, cliffy limestone member, and Bull Ridge Member. In the Hartville uplift area and the Black Hills of eastern Wyoming and western South Dakota, the Madison is represented by the Guernsey, Englewood, and Pahasapa Formations. The Amsden and Tensleep Formations of central Wyoming (Bighorn Mountains westward to the Gros Ventre and Teton Ranges) grade eastward into the Minnelusa, Hartville, Casper, and Fountain Formations, and southward into the Round Valley Limestone, Morgan Formation, and Weber Sandstone. Pennsylvanian rocks in the thrust belt of western Wyoming are included in the Wells Formation.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Geological Survey of Wyoming.

Carboniferous strata are typically exposed along the flanks of large asymmetric Precambrian-cored anticlines that formed during the Laramide orogeny (early Tertiary). These strata have been exposed as a result of late Tertiary regional uplift, erosion, and exhumation of Laramide structures. In western Wyoming, allochthonous strata are exposed on the hanging walls of west-dipping thrust plates of the Salt River, Snake River, Wyoming, and Hoback Ranges.

Carboniferous strata in Wyoming were first recognized by geologists of the U.S. Geological and Geographical Survey of the Territories during the 1860's and 1870's. Formal naming and recognition was done subsequently by the U.S. Geological Survey (Darton, 1904; Peale, 1893). The Geological Survey of Wyoming, since its inception in 1933, has published (most notably, Thomas and others, 1953) on various aspects of Wyoming's Carboniferous rocks. In recent years, increased petroleum production in the Rocky Mountains from Carboniferous reservoirs has maintained industry and academic interest in these strata.

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LOWER PART OF THE CARBONIFEROUS

By William J. Sando

INTRODUCTION

The lower part of the Carboniferous (Mississippian and lower part of the Pennsylvanian) in Wyoming consists of a marine sequence of predominantly carbonate rocks, which attains a maximum thickness of approximately 660 m. These rocks crop out on the flanks of most of the mountain uplifts of the State and are present in the subsurface of the intermontane basins (fig. 1). Throughout most of Wyoming, the original geographic positions of the Carboniferous rocks were not appreciably affected by subsequent tectonic movements, but in the western part of the State (Salt River Range, Snake River Range, Wyoming Range, Hoback Range), Laramide overthrusting brought rocks deposited farther west into juxtaposition with an autochthonous sequence.

The rock sequence is represented by 12 different formations or parts of formations in different parts of the State (fig. 2). In a large area that occupies most of the central part of Wyoming, beds of Late Devonian, Early Mississippian, and early Late Mississippian (late Famennian-early Viséan) age are included in the Madison Limestone, which is overlain by the Amsden Formation of Late Mississippian to early Middle Pennsylvanian (late Viséan–early Westphalian) age. In the westernmost part of the State, the Madison is used as a group term that

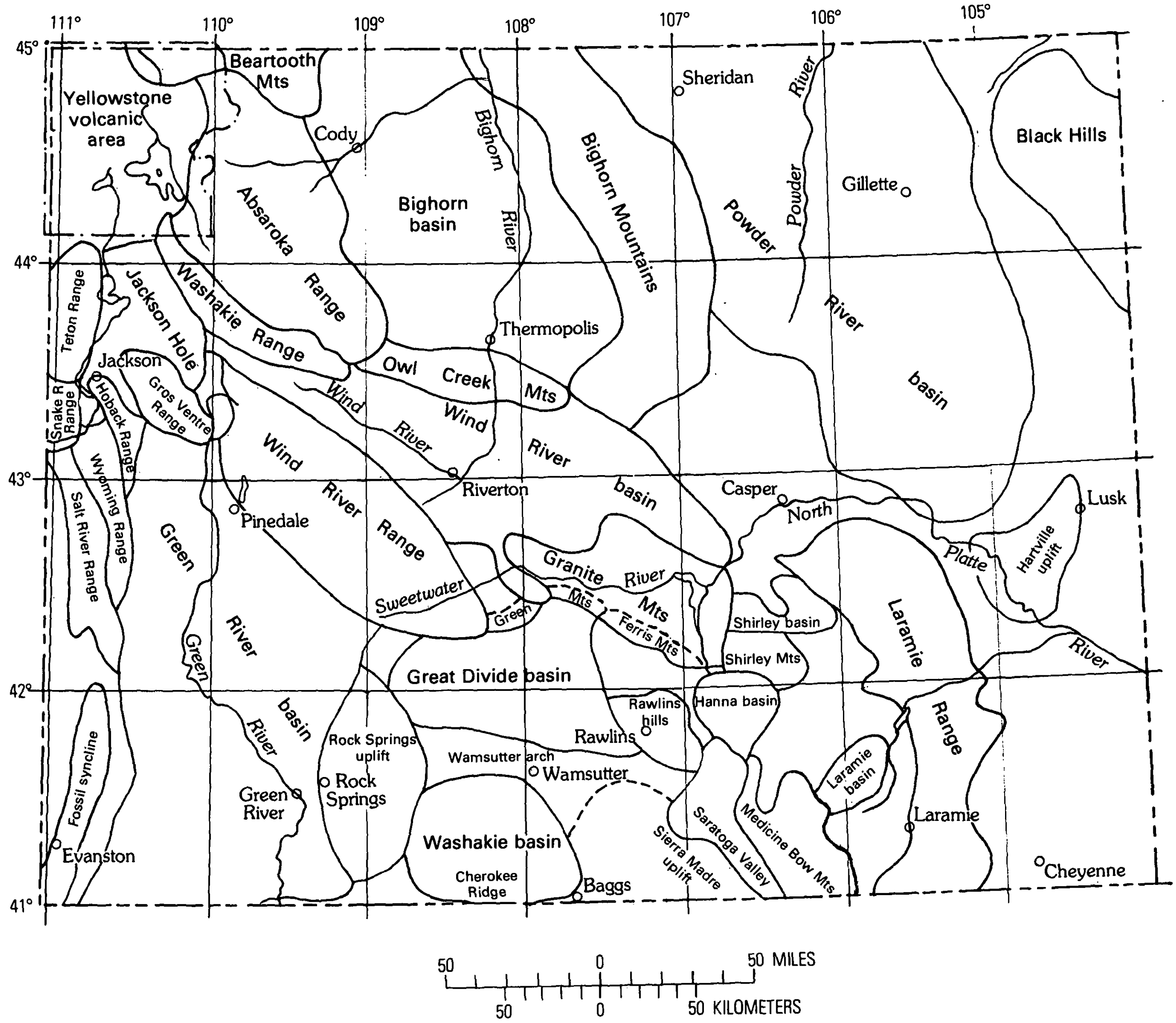

FIGURE 1.—Map of Wyoming showing present structural features (from Welder and McGreevy, 1966). Unnamed areas cannot be classified as either major uplifts or as basins.

includes two formations, the Lodgepole Limestone and Mission Canyon Limestone; the group is overlain by the Amsden Formation. In most of southeastern Wyoming, the lower part of the sequence consists of an unnamed sandstone unit of probable Late Devonian and Early Mississippian age overlain by the Madison Limestone (Early and early Late Mississippian), which is in turn overlain by the Casper Formation (Pennsylvanian and Permian) or the Fountain Formation (Pennsylvanian). In the Hartville uplift and adjacent basinal areas, the lower part of the sequence is the Guernsey Formation (Late Devonian and Early Mississippian), which is overlain by the Hartville Formation (Pennsylvanian and Permian). Terminology is variable in the subsurface of the Powder River basin in northeast Wyoming, but the Mississippian part of the sequence is most commonly called Madison Limestone and the Pennsylvanian part, Minnelusa Formation. In the northeastern corner of the State, on the flank of the Black Hills uplift, the Englewood Formation (Late Devonian and Early Mississippian) is overlain by the Pahasapa Limestone (Early Mississippian), which is in turn overlain by the Minnelusa Formation (Pennsylvanian and Permian). Nomenclature is uncertain because of the scarcity of information where the Carboniferous rocks are deeply buried in south-

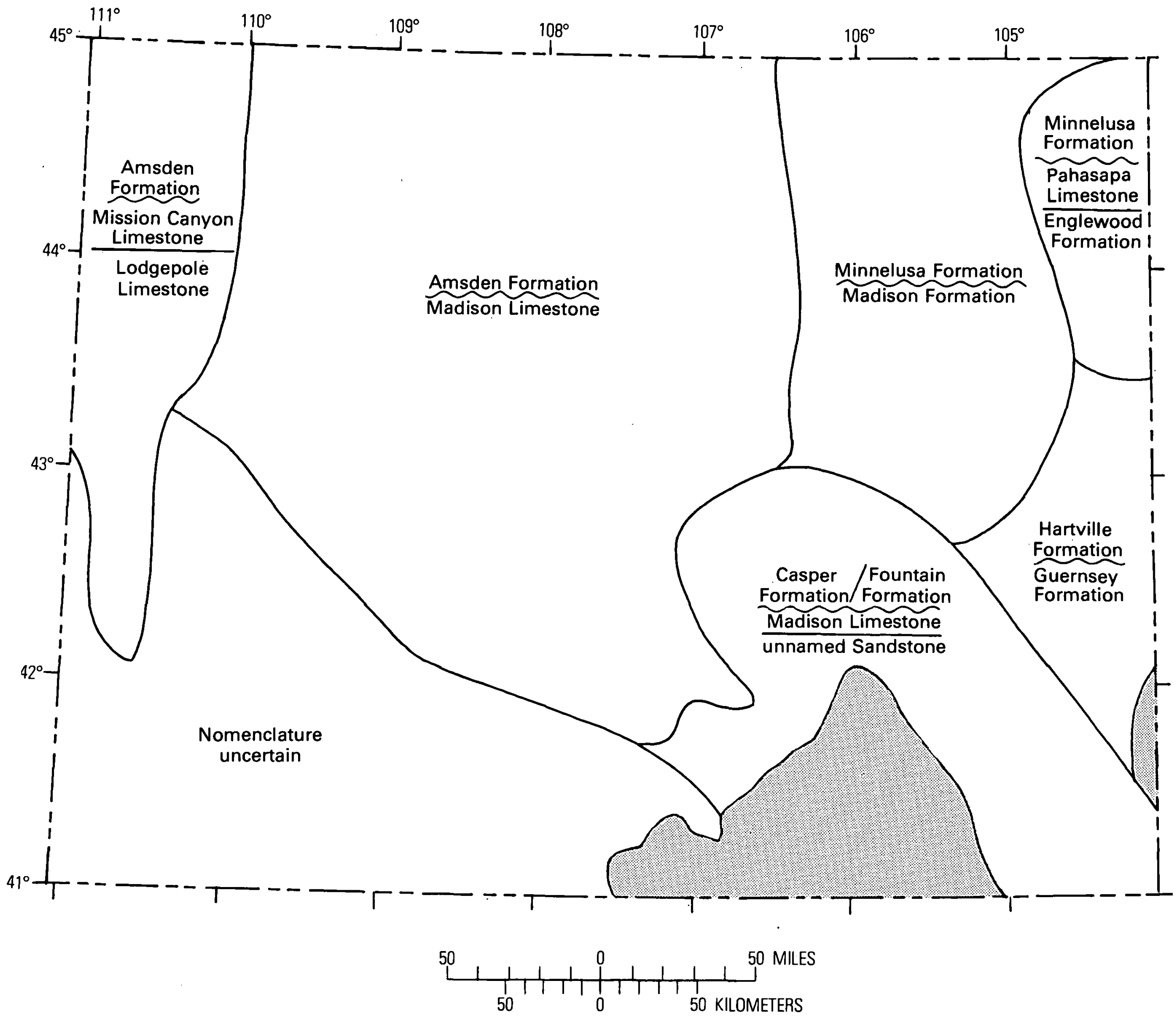

FIGURE 2.—Map of Wyoming showing formational nomenclature of Mississippian rocks and lower part of Pennsylvanian rocks. Mississippian rocks absent in shaded areas. Wavy line between formation names denotes disconformity; straight line indicates conformal contact. Precambrian rocks of the transcontinental arch shaded.

western Wyoming, which includes the Green River basin, Rock Springs uplift, and Washakie basin. Boundaries between areas of different nomenclature (fig. 2) are largely artificial and reflect different historical centers of classification of rocks of similar age and lithology.

This discussion is based mainly on regional syntheses by Mallory (1967), Sandberg and Mapel (1967), Sando (1976a), and Sando and others (1969, 1975). Other important regional summaries of the stratigraphy are Agatston (1954), Andrichuk (1955), Craig (1972), Mallory (1972), Rose (1976), Sandberg (1967), Sandberg and Klapper (1967), and Sando (1967b). Important studies of smaller areas in Wyoming are Bates (1955), Jenkins and McCoy (1958), Love and others (1953), Maughan (1963), McCaleb and Wayhan (1969), Sando (1967a, 1968, 1972, 1974, 1975, 1976b, 1977), Sando and Mamet (1974), and Todd (1964).

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I am deeply indebted to B. L. Mamet for foraminifer zone determinations, many of which are as yet unpublished. C. A. Sandberg kindly determined conodont zones in many sections and contributed stratigraphic data on the lower part of the Madison.

C. A. Sandberg and J. T. Dutro, Jr., reviewed the paper and provided useful suggestions.

REGIONAL STRATIGRAPHY

The Mississippian-early Middle Pennsylvanian succession in Wyoming represents two principal cycles of marine deposition separated by a period of epeirogenic uplift and erosion (fig. 3). The lower depositional sequence or cycle consists of the Madison Limestone and its correlatives; the upper depositional sequence or cycle consists of the Amsden Formation and its correlatives. Throughout most of Wyoming, the Mississippian-Pennsylvanian boundary is within the upper depositional sequence; sedimentation was continuous across the systemic boundary except in the eastern part of the State, where beds of Early to Middle Pennsylvanian age rest on beds of Early to Late Mississippian age.

The lower depositional cycle began in the Late Devonian, was interrupted briefly at the Devonian-Mississippian boundary, and continued into the Late Mississippian (early Meramecian). The upper depositional cycle began in the Late Mississippian (middle Meramecian) and continued into the Middle Pennsylvanian (Atokan). In terms of western European chronostratigraphy, the lower depositional sequence is Late Famennian-early Viséan, and the upper depositional sequence is late Viséan-early Westphalian. Deposition was continuous across the lower Carboniferous-upper Carboniferous boundary in most of the State. The major regional break in deposition began in the middle Viséan and extended into the Namurian.

An isopach map of the Mississippian part of the succession (fig. 4) shows a general thickening of these rocks northward and westward from a zero edge in southeastern Wyoming. Two cross sections parallel to thickness trends (fig. 5; see also table 1) show northward- and westward-expanding wedges of predominantly shelf carbonate rocks of the lower depositional sequence overlain disconformably by thin predominantly terrigenous rocks of the upper depositional sequence.

The character of the lower depositional sequence changes rapidly in adjacent southeastern Idaho, where terrigenous rocks were deposited in a deep basin west of the carbonate shelf margin. The Mississippian part of the upper depositional sequence thickens rapidly in western Wyoming, where a wedge of carbonate rocks extends eastward from a thick body of Upper Mississippian shelf carbonate rocks in Idaho. The Pennsylvanian (Morrowan and Atokan) part of the upper depositional sequence (not shown on fig. 5) is a nearly tabular body of shaley and sandy carbonate rocks throughout most of the State.

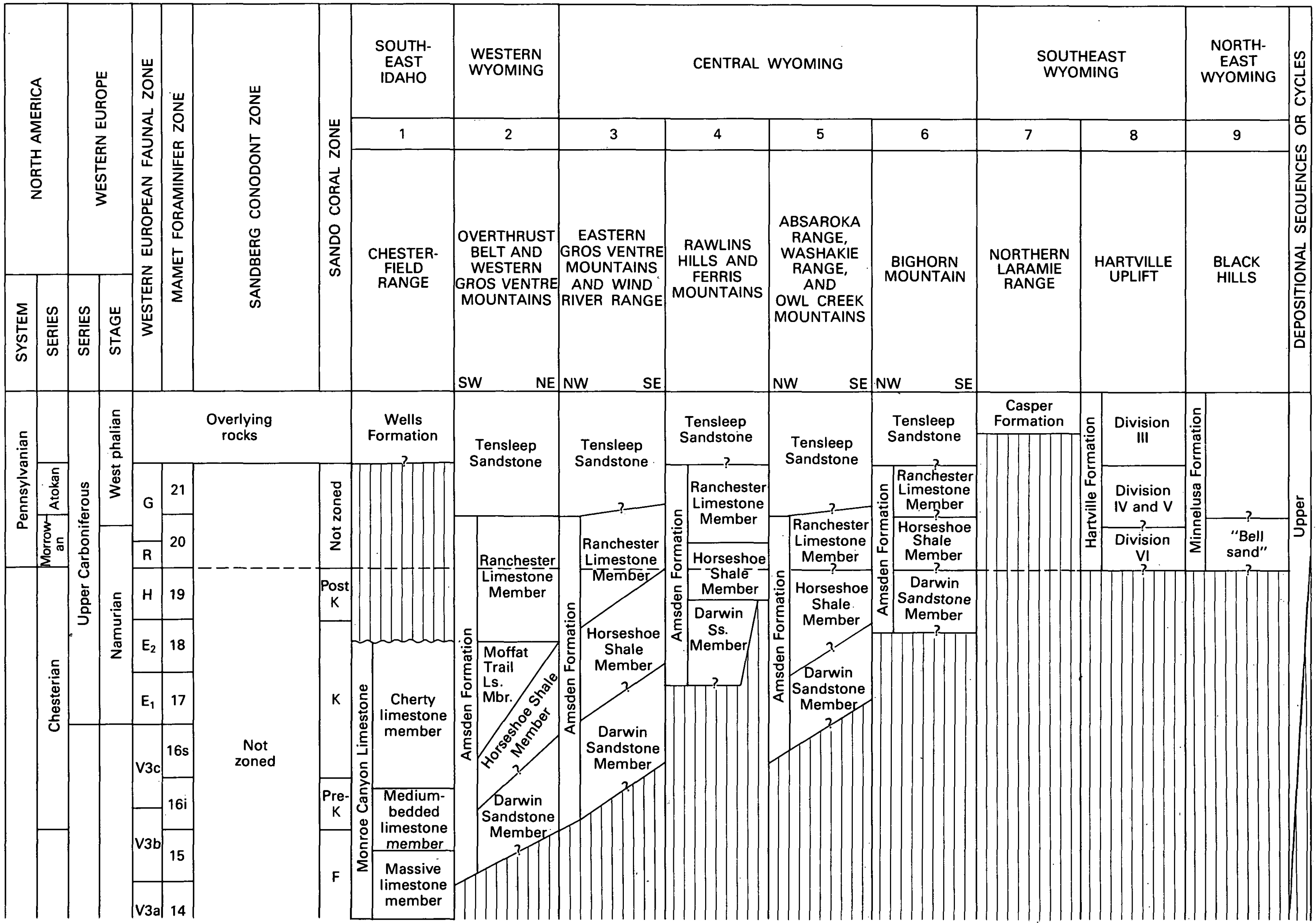
LOWER DEPOSITIONAL SEQUENCE

MADISON GROUP

In western Wyoming, the Madison is a group that includes the Lodgepole Limestone and the overlying Mission Canyon Limestone, formations whose type localities are in Montana. The Madison is about 375 m thick in this area. In parts of western Wyoming, conodont-bearing dark shale as much as 18 m thick is at the base of the Lodgepole Limestone, which rests disconformably on the Devonian Darby Formation. The dark shale represents the upper tongue of the Cottonwood Canyon Member of the Lodgepole, a unit of earliest Kinderhookian (early Tournaisian) age (fig. 3, column 2). The member contains conodonts of the *Siphonodella sandbergi* Zone (Sandberg and Mapel, 1967).

The Cottonwood Canyon Member is succeeded conformably by the Paine Member of the Lodgepole Limestone, which is at the base of the formation where the Cottonwood Canyon Member is absent. The Paine is about 50 m thick. The lower 3–5 m of the member consists of silty, glauconitic, crinoidal limestone and is overlain by cherty thin-bedded silty fine-grained limestone. The lower crinoidal part of the Paine contains rare corals and brachiopods of Coral Zone A of Sando and others (1969) and conodonts of the Lower *Siphonodella crenulata* Zone (Sandberg and Mapel, 1967). The poorly fossiliferous remainder of the Paine Member contains corals and brachiopods of Coral Zone B, foraminifers of Zone pre-7 (Mamet, *in* Sando and others, 1969), and conodonts of the *Siphonodella isosticha*-Upper *S. crenulata* Zone of Sandberg (written commun., 1977). The Paine Member is of Kinderhookian (middle Tournaisian) age.

The Paine Member is overlain conformably by the Woodhurst Member of the Lodgepole Limestone, which consists of about 80 m of cyclically interbedded, thin-bedded silty fine-grained limestone and oolitic crinoidal limestone. This unit contains a rich fauna of conodonts, foraminifers, brachiopods, corals, bryozoans, and gastropods which represent the *Siphonodella isosticha*-Upper *S. crenulata* Zone, the *Gnathodus typicus* Zone, the *Pseudopolygnathus* n. sp. Zone, foraminiferal Zones 7 and 8, and Coral Zone C_1. The Woodhurst Member includes beds of late Kinderhookian (middle Tournaisian) and early Osagean (middle and upper Tournaisian) age.


NORTH AMERICA
SYSTEM
SERIES
WESTERN EUROPE
SERIES
STAGE
WESTERN EUROPEAN FAUNAL ZONE
MAMET FORAMINIFER ZONE
SANDBERG CONODONT ZONE
SANDO CORAL ZONE
SOUTH-EAST IDAHO
WESTERN WYOMING
CENTRAL WYOMING
SOUTHEAST WYOMING
NORTH-EAST WYOMING
DEPOSITIONAL SEQUENCES OR CYCLES
1
2
3
4
5
6
7
8
9
CHESTER-FIELD RANGE
OVERTHRUST BELT AND WESTERN GROS VENTRE MOUNTAINS
EASTERN GROS VENTRE MOUNTAINS AND WIND RIVER RANGE
RAWLINS HILLS AND FERRIS MOUNTAINS
ABSAROKA RANGE, WASHAKIE RANGE, AND OWL CREEK MOUNTAINS
BIGHORN MOUNTAIN
NORTHERN LARAMIE RANGE
HARTVILLE UPLIFT
BLACK HILLS
SW
NE
NW
SE
NW
SE
NW
SE
Pennsylvanian
Atokan
Morrow-an
Chesterian
Upper Carboniferous
West phalian
Namurian
G
R
H
E2
E1
V3c
V3b
V3a
21
20
19
18
17
16s
16i
15
14
Overlying rocks
Not zoned
Not zoned
Post K
K
Pre-K
F
Wells Formation
Monroe Canyon Limestone
Cherty limestone member
Medium-bedded limestone member
Massive limestone member
Tensleep Sandstone
Amsden Formation
Ranchester Limestone Member
Moffat Trail Ls. Mbr.
Horseshoe Shale Member
Darwin Sandstone Member
Horseshoe Shale Member
Darwin Ss. Member
Casper Formation
Hartville Formation
Division III
Division IV and V
Division VI
Minnelusa Formation
"Bell sand"
Upper
?

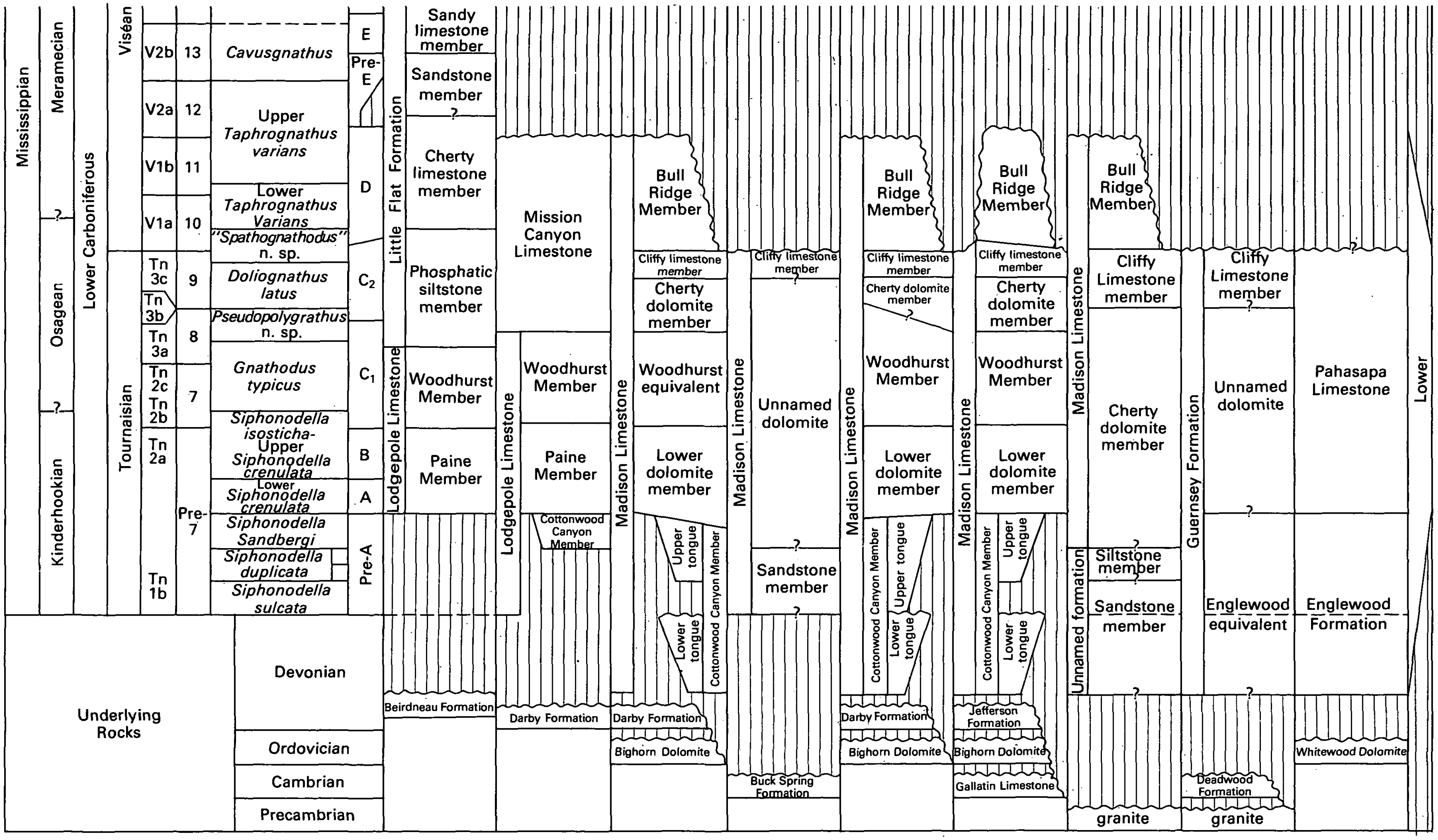


FIGURE 3.—Nomenclature and temporal relations of Mississippian rocks and lower part of Pennsylvanian rocks in Wyoming and adjacent parts of Idaho. [Vertical lines denote hiatus. Sources of data for stratigraphic columns (indicated by numbers) modified from: 1, Dutro and Sando (1963), Sando and others (1976), Poole and Sandberg (1977); 2, Sandberg and Mapel (1967), Sando and others (1975), Sando (1977); 3, Sandberg and Mapel (1967), Sandberg and Klapper (1967), Sando (1967a, 1977), Sando and others (1975); 4, Sando (1967), Sando and others (1975), M. W. Reynolds (written commun., 1968); 5, Sando (1967a, 1975), Sando and others (1975), Sandberg and Mapel (1967), Sandberg and Klapper (1967), Sandberg (1967); 6, Sando (1976b), Sando and others (1975), Sandberg (1967), Sandberg and Klapper (1967), Sandberg and Mapel (1976); 7, Maughan (1963), Mallory (1967), Sando and Sandberg (unpub. data, 1977); 8, Love and others (1953), Sando (unpub. data, 1977); 9, Sandberg and Mapel (1967), Sando (unpub. data, 1977). Foraminifer and coral zones modified slightly from Sando and others (1969). Conodont zones from Sandberg and Mapel (1967) and Sandberg (written commun., 1977).]

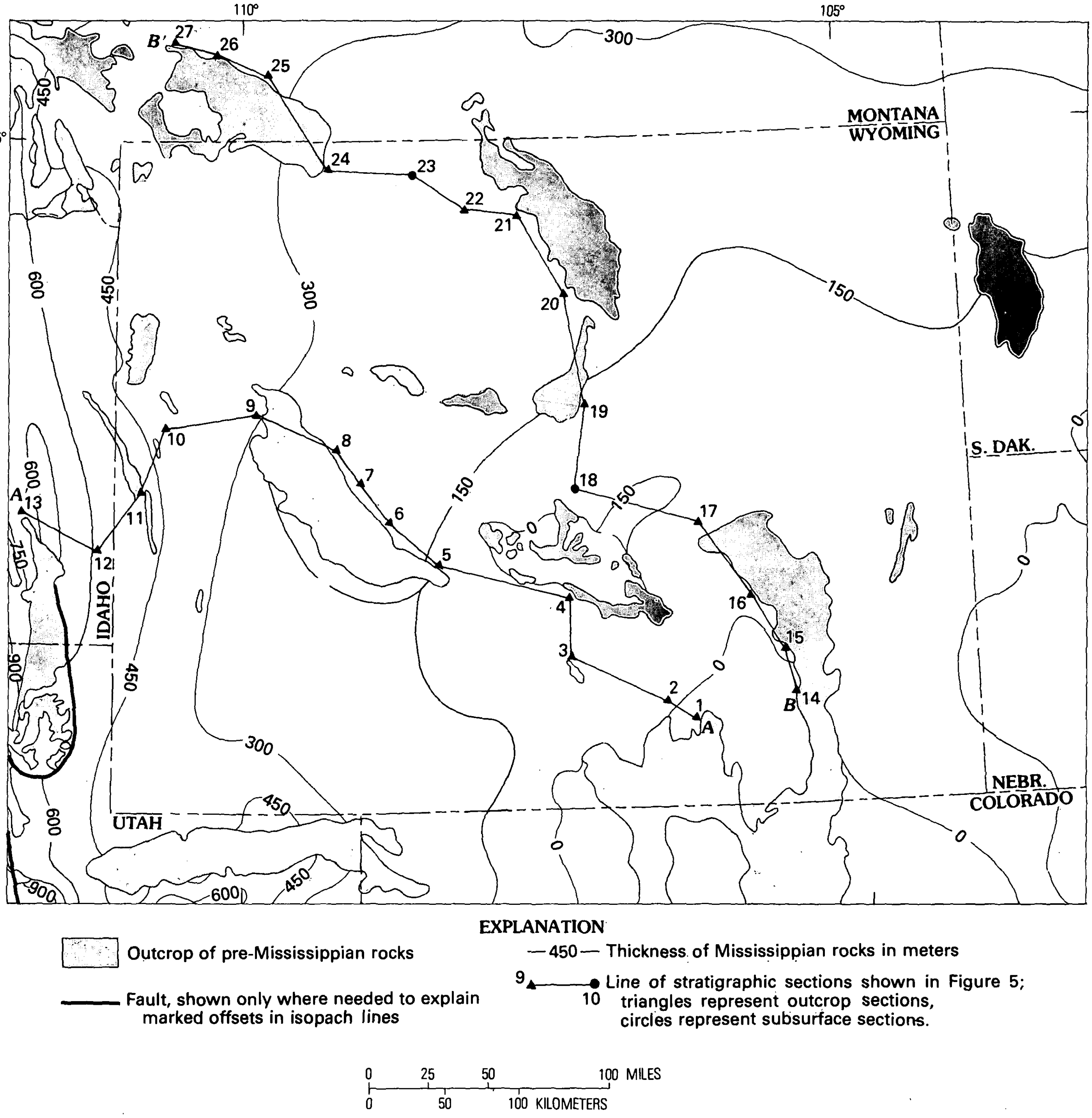


FIGURE 4.—Thickness of Mississippian rocks in Wyoming and adjacent areas, trends of cross sections, and locations of key stratigraphic sections (numbered) shown in figure 5 (modified from Sando, 1976a).

The Woodhurst Member of the Lodgepole Limestone is succeeded conformably by the Mission Canyon Limestone, which consists of interbedded, thick-bedded crinoidal oolitic limestone and cherty thin-bedded fine-grained dolomite and dolomitic limestone in the lower half and dolomite and dolomitic limestone interbedded with cherty predominantly fine-grained limestone in the upper half. The upper half is also characterized by two solution-breccia zones except in the Hoback Range, where three evaporitic zones are preserved (Sando, 1977). The total thickness of the Mission Canyon in western Wyoming is about 250 m. The formation is moderately fossiliferous, containing foraminifers,

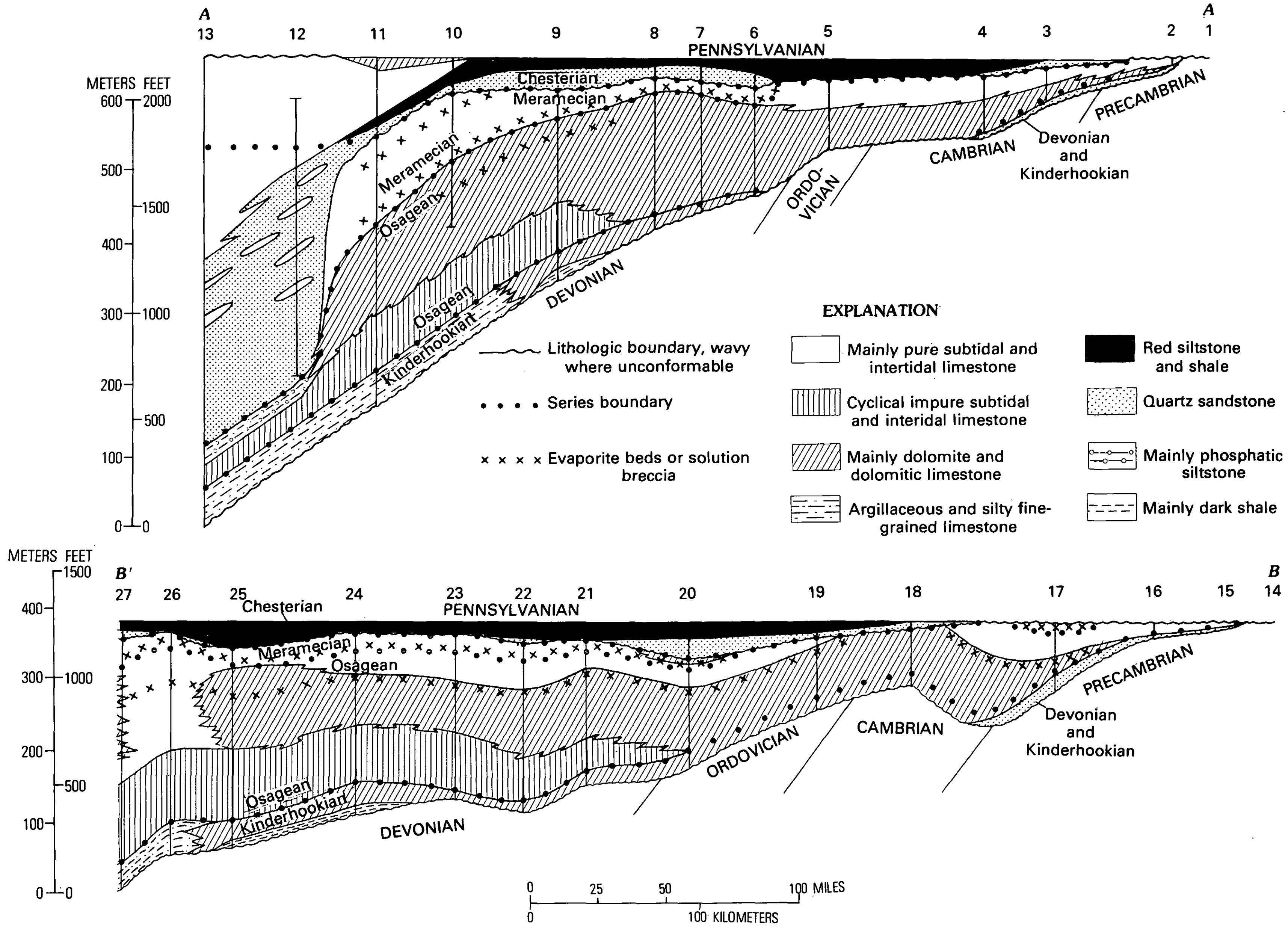


FIGURE 5.—Stratigraphic and structural relations of Mississippian rocks from southeastern Wyoming to southeastern Idaho (A–A′) and from southeastern Wyoming to southern Montana (B–B′). Modified from Sando (1976a). Locations of stratigraphic sections are shown on figure 4 and described in table 1.

TABLE 1.—*Geographic locations and sources of information for stratigraphic sections shown on figures 4 and 5*

[Locality numbers refer to numbers shown on figs. 4 and 5]

Section No.	Locality name	Geographic location	Source of information
1	Arlington	Approximately T. 18 N., R. 79 W., Carbon County, Wyo.	Maughan (1963)
2	Elk Mountain	Approximately T. 19 N., R. 82 W., Carbon County, Wyo.	Do.
3	Buck Spring	Sec. 33, T. 23 N., R. 88 W., Carbon County, Wyo.	Sando (1967a); Sando and others (1975)
4	Cottonwood Creek	Secs. 27 and 34, T. 27 N., R. 88 W., Carbon County, Wyo.	Reynolds, M. W. (written commun., 1968)
5	Sweetwater Canyon	Secs. 27 and 34, T. 29 N., R. 97 W., Fremont County, Wyo.	Sando (1967a), Sando and others (1975)
6	Sinks Canyon	Sec. 18, T. 32 N., R. 100 W., Fremont County, Wyo.	Do.
7	Washakie Reservoir	Sec. 18, T. 1 S., R. 2 W., Fremont County, Wyo	Do.
8	Bull Lake Creek	Secs. 2 and 3, T. 2 N., R. 4 W., Fremont County, Wyo.	Do.
9	Big Sheep Mountain	Sec. 6, T. 38 N., R. 108 W., and sec. 31, T. 39 N., R. 108 W., Sublette County, Wyo.	Blackwelder, Eliot (unpub. data); Richmond (1945)
10	Hoback Canyon	Sec. 3, T. 38 N., R. 115 W., Teton County, Wyo	Sando (1977); Sando and others (1975)
11	Haystack Peak	Sec. 19, T. 34 N., R. 117 W., Lincoln County, Wyo.	Do.
12	Wells Canyon	Secs. 10 and 11, T. 10 S., R. 45 E., Caribou County, Idaho	Gulbrandsen and others (1956)
13	Little Flat Canyon	Secs. 17 and 20, T. 7 S., R. 40 E., Bannock County, Idaho	Dutro and Sando (1963), Sando (1977), Sando and others (1976)
14	Laramie Range	Approximately T. 20 N., R. 72 W., Albany County, Wyo.	Maughan (1963)
15	Wheatland Reservoir	Sec. 17, T. 23 N., R. 73 W., Albany County, Wyo.	Do.
16	Marshall	Approximately T. 26 N., and T. 27 N., R. 75 W., Albany County, Wyo.	Do.
17	Casper Mountain	Sec. 9, T. 32 N., R. 79 W., Natrona County, Wyo.	Do.
18	National Co-op Refining Co., Wallace Creek No. 2 well.	Sec. 16, T. 34 N., R. 87 W., Natrona County, Wyo.	American Stratigraphic Co.
19	Middle Buffalo Creek	Secs. 20 and 21, T. 40 N., R. 86 W., Natrona County, Wyo.	Sando (1976b), Sando and others (1975)
20	Tensleep Canyon	Sec. 27, T. 48 N., R. 87 W., Washakie County, Wyo.	Sando (unpub. data, 1977), Sando and others (1975)
21	Shell Canyon	Secs. 9 and 17, T. 53 N., R. 90 W., Big Horn County, Wyo.	Do.
22	Sheep Mountain	Sec. 35, T. 54 N., R. 94 W., and sec. 2, T. 53 N., R. 94 W., Big Horn County, Wyo.	Do.
23	Ohio Oil Co., Easton unit 6 well	Sec. 28, T. 56 N., R. 97 W., Big Horn County, Wyo.	American Stratigraphic Co.
24	Clarks Fork Canyon	Secs. 5, 6, and 7, T. 56 N., R. 103 W., Park County, Wyo.	Sando (1972, 1975), Sando and others (1975)
25	Ben Bow Mine Road	Secs. 20 and 29, T. 5 S., R. 16 E., Stillwater County, Mont.	Sando (1972)
26	Baker Mountain	Secs. 34 and 35, T. 3 S., R. 12 E., Sweetgrass and Park Counties, Mont.	Do.
27	Livingston	Secs. 1 and 2, T. 3 S., R. 9 E., and sec. 35, T. 2 S., R. 9 E., Park County, Mont.	Do.

corals, brachiopods, gastropods, and bryozoans and rare conodonts. Foraminiferal Zones 8 through 11 and Coral Zones C_1, C_2 and D are represented in the sequence. The Mission Canyon ranges in age from middle Osagean into early Meramecian (late Tournaisian-early Viséan).

MADISON LIMESTONE

Throughout most of Wyoming, the lower depositional sequence or cycle is represented by a single formation, the Madison Limestone. Six members are recognized in the Madison in most of central Wyoming, but in the Rawlins hills and in southeastern Wyoming, the lower four members are not recognizable, and sandstone occurs at the base of the Mississippian (fig. 3). Although the members of the Madison were based on outcrop studies, examination of well logs from the Bighorn, Wind River, and Powder River basins indicate that most of the members are also recognizable in the subsurface. The

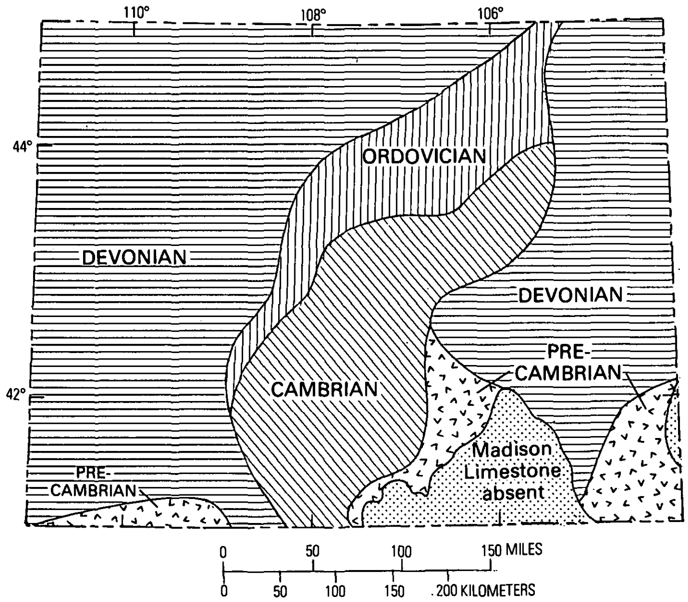


FIGURE 6.—Paleogeologic map showing ages of rocks underlying the Madison Limestone and equivalent rocks in Wyoming. Modified from Craig (1972, fig. 2) and Baars (1972, figs. 6 and 8).

Madison Limestone rests unconformably on a variety of formations ranging from Precambrian to Devonian in age; the subcrop becomes progressively younger northward and westward from the zero edge of the Madison in southeastern Wyoming (fig. 6).

Throughout most of central Wyoming, the basal unit of the Madison Limestone is the Cottonwood Canyon Member, which consists mostly of conodont-bearing shale, siltstone, and silty dolomite as much as 12 m thick. The member includes a lower tongue of latest Devonian (Famennian) age separated by a disconformity from an upper tongue of early Kinderhookian age (*Siphonodella sulcata, S. duplicata,* and *S. sandbergi* conodont zones).

The Cottonwood Canyon Member is succeeded conformably by the lower dolomite member, which is mostly thick-bedded crinoidal dolomite and dolomitic limestone as much as 30 m thick. The member contains a sparse fauna of conodonts, corals, and brachiopods and is of Kinderhookian (middle Tournaisian) age. It is approximately equivalent to the Paine Member of the Lodgepole, and a lateral transition into the Paine can be observed in outcrop sections in the Beartooth Mountains of southern Montana (Sando, 1972). In the Owl Creek and southern Bighorn Mountains, the underlying Cottonwood Canyon Member is absent, and the lower dolomite member rests directly on beds of Cambrian, Ordovician, or Devonian age.

The lower dolomite member is overlain conformably by the Woodhurst Member or equivalent beds throughout most of central Wyoming. In the Abasaroka Range and central and northern Bighorn Mountains, the lower part of this member consists of crossbedded crinoidal dolomite or thick-bedded oolitic limestone, and the upper part consists of typical Woodhurst facies silty, thin-bedded, cyclically interbedded fine-grained limestone and bioclastic crinoidal limestone. The Woodhurst Member attains a maximum thickness of about 90 m. In the Wind River Mountains, Owl Creek Mountains, and southern Bighorn Mountains, the entire member is mostly crossbedded crinoidal dolomite and limestone. The typical Woodhurst facies is a wedge that pinches out southward from Montana and eastward from western Wyoming (fig. 5). The typical Woodhurst facies contains conodonts (rare) and foraminifers, brachiopods, and corals (moderately abundant). Foraminifer Zones 7 and 8 are represented throughout most of central Wyoming, and Zone 9 has been identified at the top of the member in the Absaroka Range. Coral Zone C_1 is also represented. The Woodhurst is mostly Osagean in age but contains some beds of Kinderhookian age in the lower part. The Kinderhookian-Osagean boundary is commonly placed at the base of the member for convenience. In terms of western European chronostratigraphy, the Woodhurst is middle and upper Tournaisian in age.

The Woodhurst Member is succeeded conformably by the cherty dolomite member, whose maximum thickness is about 70 m. The member consists of shattered and brecciated, thin-bedded, fine-grained cherty dolomite and dolomitic limestone at most localities, but elsewhere includes some beds of crinoidal or fine-grained limestone. The commonly shattered character of the beds suggests that the sequence was originally evaporitic in part and has subsequently been leached. Fossils are extremely rare, but a few brachiopods and corals of Osagean age have been found at some localities. The age of the cherty dolomite, bracketed by dated units above and below, is late Osagean (late Tournaisian) at most localities.

The cherty dolomite member is overlain conformably by the cliffy limestone member, which attains a maximum thickness of about 65 m. The upper part of the member consists of a very widespread sequence of commonly cherty, fine- to coarse-grained, mostly thick-bedded bioclastic crinoidal limestone and dolomite that contains abundant spiriferoid brachiopods and some corals and foraminifers. In central Wyoming, a solution breccia that represents a leached evaporite-carbonate-terrigenous sequence occurs at the base of the member, except in the southern Wind River Mountains and Rawlins hills. The beds above the solution breccia are brecciated as a result of collapse. Red sandstone and siltstone were deposited in solution cavities throughout the member during pre-Amsden uplift and Amsden deposition (Sando, 1974). The member contains the Zone 9 foraminifer fauna and corals of Zone C_2 at most localities, except in the northern Bighorn Mountains, where foraminifer Zone 10 has been found at a few places. The cliffy limestone member is of upper Osagean (upper Tournaisian and lowermost Viséan) age. For convenience, the Osagean-Meramecian boundary is commonly placed at the top of the member.

The cliffy limestone member is overlain conformably by the Bull Ridge Member in much of central Wyoming. The Bull Ridge Member reaches a maximum thickness of about 25 m and consists of a lower unit of red and yellow shale and siltstone and minor thin carbonate beds and an upper unit of thin- to medium-bedded, sparsely cherty, fine- to medium-grained, commonly stromatolitic limestone or dolomite. The lower unit is commonly a solution breccia, and the upper unit is commonly collapsed, indicating the former presence of evaporite beds. Sinkholes and solution cavities related to pre-Amsden emergence and filled by Amsden sediments are common in the Bull Ridge Member (Sando, 1974). The member was partly or entirely removed by pre-Amsden erosion at many localities in central Wyoming, particularly in the southern Wind River Mountains, eastern Owl Creek Mountains, and southern Bighorn Mountains. The upper part of the Bull Ridge Member contains corals and brachiopods of Zone D; foraminifers of Zones 10, 11, and (rarely) 12; and conodonts (rare) of the Upper or Lower *Taphrognathus varians* Zone.

In the Rawlins hills and Ferris Mountains, the Madison includes a lower, poorly dated sandstone member thought to be of Kinderhookian age, overlain by dolomite equivalent to the lower dolomite member, Woodhurst Member, and cherty dolomite member, and capped by the cliffy limestone member (fig. 3). In the northern Laramie Range, the Madison conformably overlies an unnamed quartz sandstone and siltstone sequence of probable Devonian and early Kinderhookian age, formerly regarded as the Cambrian and Ordovician Deadwood Formation. The Madison here consists of the cherty dolomite member, cliffy limestone member, and, locally, the Bull Ridge Member. Farther south in the Laramie Range, the carbonate appears to interfinger with quartz sandstone and conglomerate that unconformably overlies Precambrian granite (Maughan, 1963); exact age relationships remain to be determined in this area.

GUERNSEY FORMATION

In the Hartville uplift and adjacent subsurface, rocks equivalent to the Madison Limestone are included in the Guernsey Formation, which unconformably overlies quartzite of the Cambrian and Ordovician Deadwood Formation and Precambrian granite (Love and others, 1953). The lower part of the Guernsey consists of red siltstone of probable Devonian and Mississippian age regarded as an equivalent of the Englewood Formation of the Black Hills. Above the siltstone is unnamed dolomite and dolomitic limestone that is overlain by the cliffy limestone member. The Guernsey attains a maximum thickness of about 90 m.

ENGLEWOOD FORMATION AND PAHASAPA LIMESTONE

In the Black Hills and adjacent subsurface, the base of the Carboniferous is included in the Englewood Formation, a sequence of red to purple dolomite, dolomitic limestone, limestone, and shale as much as 27 m thick (Klapper and Furnish, 1962; Sandberg and Mapel, 1967). The Englewood is overlain by the Pahasapa Limestone, which includes limestone and dolomite of Kinderhookian and Osagean age as much as 190 m thick.

UPPER DEPOSITIONAL SEQUENCE

AMSDEN FORMATION

In western and central Wyoming, the upper depositional sequence or cycle is represented by the Amsden Formation (fig. 2), a transgressive sequence of quartz sandstone, siltstone, shale ,and carbonate rocks that attains a maximum thickness of about 150 m. The Amsden includes three members, except in western Wyoming, where four are recognized (fig. 3).

The age and correlation of the Amsden Formation have been the subjects of much controversy. The interpretations summarized herein are based on a comprehensive study of the physical stratigraphy and paleontology of the Amsden Formation in Wyoming by Sando and others (1975). The physical stratigraphy in this detailed analysis was based on study of 50 stratigraphic sections in Wyoming. The age and correlation of the Amsden was based on study of 160 collections including well over

6,000 specimens of fossils representing more than 350 taxa by a team of 10 paleontologists, each expert in one of the biologic groups represented. Critical age determinations rested mainly on study of foraminifers, algae, and brachiopods by specialists who worked independently and had no preconceptions about the physical stratigraphy when they interpreted the ages of fossil assemblages at various localities. In another part of this paper, E. K. Maughan presents a different interpretation of the Amsden, based mostly on physical stratigraphy, that emphasizes two presumed unconformities not recognized by Sando and others (1975). We have examined Maughan's reinterpretations of the paleontologic evidence and cannot agree with his conclusions.

The basal unit of the Amsden at most localities is the Darwin Sandstone Member, which consists of unfossiliferous, thin- to thick-bedded, ordinarily crossbedded, white to red, fine- to medium-grained quartz sandstone as much as 65 m thick. The Darwin rests disconformably on the Bull Ridge Member or the underlying cliffy limestone member of the Madison (Sando and others, 1975, pl. 1). At some localities, the Darwin is absent or present only in sinkholes in the Madison, and the basal unit is the Horseshoe Shale Member, which ordinarily overlies the Darwin. The Darwin Sandstone ranges in age from late Meramecian (late Viséan) in western Wyoming to late Chesterian (Namurian) in east-central Wyoming, on the basis of paleontologic dating of the overlying and underlying beds.

The Darwin is overlain conformably by a transgressive red-bed sequence named the Horseshoe Shale Member. This unit consists of a much as 45 m of red siltstone, shale, and rare thin beds of fine-grained sandstone and limestone. Although fossils are generally rare in the Horseshoe, occurrences of ostracodes, mollusks, brachiopods, foraminifers, and, to a limited extent, corals support a late Chesterian (early and middle Namurian) age in western and west-central Wyoming and a Morrowan (late Namurian-early Westphalian) age in the Bighorn Mountains and Rawlins hills. Foraminifer Zones 18, 19, and 20 are represented.

In western Wyoming, the Moffat Trail Limestone Member is a transgressive wedge-shaped body of carbonate rock as much as 33 m thick that intervenes between the Horseshoe Shale Member and the Overlying Ranchester Limestone Member of the Amsden. The Moffat Trail Member is predominantly fossiliferous, cherty, thin- to thick-bedded, medium- to coarse-grained crinoidal bioclastic limestone. The fauna consists of abundant foraminifers, bryozoans, brachiopods, ostracodes, gastropods, and corals that represent foraminifer Zones 17 and 18 and coral Zone K. The Moffat Trail is of Chesterian (late Viséan-early Namurian) age.

The highest unit in the Amsden is the Ranchester Limestone Member, which conformably overlies the Moffat Trail Limestone Member in western Wyoming and the Horseshoe Shale Member elsewhere. The Ranchester is a heterogeneous sequence of interbedded cherty dolomite and limestone, sandstone, and shale that attains a maximum thickness of about 75 m. The member contains a sparse normal marine fauna dominated by brachiopods and mollusks. The distribution of brachiopods and foraminifers indicates a late Chesterian-Morrowan age in western Wyoming and a Morrowan-Atokan age in central Wyoming.

CASPER FORMATION

The Casper Formation is a sequence of quartz sandstone and carbonate rocks in the Laramie basin and on the flanks of the Laramie Range in southeastern Wyoming. The Casper includes beds of Pennsylvanian and Permian age; the Pennsylvanian part ranges from about 30 to 180 m in thickness (Mallory, 1967). Equivalents of the Amsden Formation (Morrowan-Atokan) are present in parts of southeastern Wyoming, but Missourian and Des Moinesian rocks disconformably overly the Madison on the Pathfinder and Front Range uplifts (Mallory, 1963, 1967).

FOUNTAIN FORMATION

The Fountain Formation, named for exposures along the Front Range in Colorado, is a sequence of red sandstone, conglomerate, and arkose on the flanks of the Pennsylvanian uplifts in southeastern Wyoming. According to Mallory (1967), the Fountain ranges in age from Atokan to Virgilian. Hence, the Fountain may contain beds equivalent to the upper part of the Amsden Formation.

HARTVILLE FORMATION

The name Hartville Formation is applied to a sequence of sandstone, shale, and carbonate rocks of Pennsylvanian and Permian age in the Hartville uplift and adjacent subsurface in southeastern Wyoming. The six divisions of the formation are numbered from I at the top to VI at the base (Love and others, 1953).

Division VI, also known as the Fairbank Formation, is mostly unfossiliferous quartz sandstone that disconformably overlies the Guernsey Formation. The age of this unit is Chesterian and (or) Morrowan, probably the latter, which indicates that it is equivalent to the Darwin Sandstone Member of the Amsden Formation and, in fact, is probably laterally continuous with the Darwin. Division VI attains a maximum thickness of about 30 m. Thickness varies greatly depending on the topography of the karst surface on which it was deposited.

Division VI is conformably overlain by Division V, which consists principally of cherty limestone containing red shale beds. The overlying Division IV consists of red shale containing limestone and dolomite beds. Divisions IV and V are commonly lumped together for stratigraphic analysis to form a limestone-shale interval that attains a maximum thickness of about 90 m. Brachiopods, conodonts, foraminifers, corals, and bryozoans found near the base of Division V suggest a Morrowan age, and fusulinids from Division IV are of Atokan age. The combined Division IV–V unit is probably equivalent to and continuous with the Horseshoe Shale Member and Ranchester Limestone Member of the Amsden Formation.

Division IV is overlain conformably by Division III, a sequence of carbonate and quartz sandstone of Des Moinesian age that is equivalent to the lower part of the Tensleep Sandstone.

MINNELUSA FORMATION

The Minnelusa Formation, named for exposures in the Black Hills, is a sequence of carbonate, sandstone, and solution breccias or evaporite beds of Pennsylvanian and Permian age (Agatston, 1954; Bates, 1955; Mallory, 1967). The name has been used in northeastern Wyoming on the flank of the Black Hills and in the subsurface of the Powder River basin. The Minnelusa disconformably overlies the Pahasapa Limestone in the Black Hills and the Madison Limestone in the subsurface. A basal sandstone of probable Morrowan age, the Bell sand of subsurface usage, is probably laterally continuous with the Darwin Sandstone Member of the Amsden Formation. Beds equivalent to the Horseshoe and Ranchester Members of the Amsden are probably present in the sequence above the Bell sand.

SUMMARY OF GEOLOGIC HISTORY

During the early part of Carboniferous time, most of Wyoming was part of the Cordilleran platform, a broad cratonic area of relatively thin marine deposition that extended from southern Canada southward into Mexico (fig. 7). Extending across the southeast corner of the state was the Transcontinental arch, an emergent linear positive area that divided most of the United States into eastern and western areas of deposition and acted as a barrier to faunal migrations. Immediately west of Wyoming was the Cordilleran geosyncline, whose eastern part was a foreland basin during Mississippian time. West of the foreland basin, the geosyncline was divided into Antler orogenic highland, inner-arc basin, and island arc. The paleogeographic development described below has been illustrated by a series of maps in Sando (1976a).

Late Devonian regression created a land area of low relief in Wyoming. During latest Devonian (Famennian) and early Kinderhookian (early Tournaisian) times, a sea advanced rapidly onto the Cordilleran platform from southwest of Wyoming over a terrane composed of Precambrian, Cambrian, Ordovician, and Devonian rocks. Terrigenous and silty carbonate rocks (Cottonwood Canyon Member of Madison Limestone) were deposited in shallow marine basins on the Cordilleran platform, and the Transcontinental arch was a land area bordered on the west by an apron of quartz sand derived from the arch. Mild epeirogenic movements at the end of the Devonian caused temporary emergence or stillstand over most of the State.

Continued transgression during late Kinderhookian (early middle Tournaisian) time caused the sea to expand and cover all but the southeast corner of Wyoming. Subtidal and intertidal carbonate sediments (Madison Limestone) were deposited over most of the State except for deeper water carbonate rocks (Paine Member of Lodgepole Limestone) deposited on the eastern slope of the foreland basin in western Wyoming.

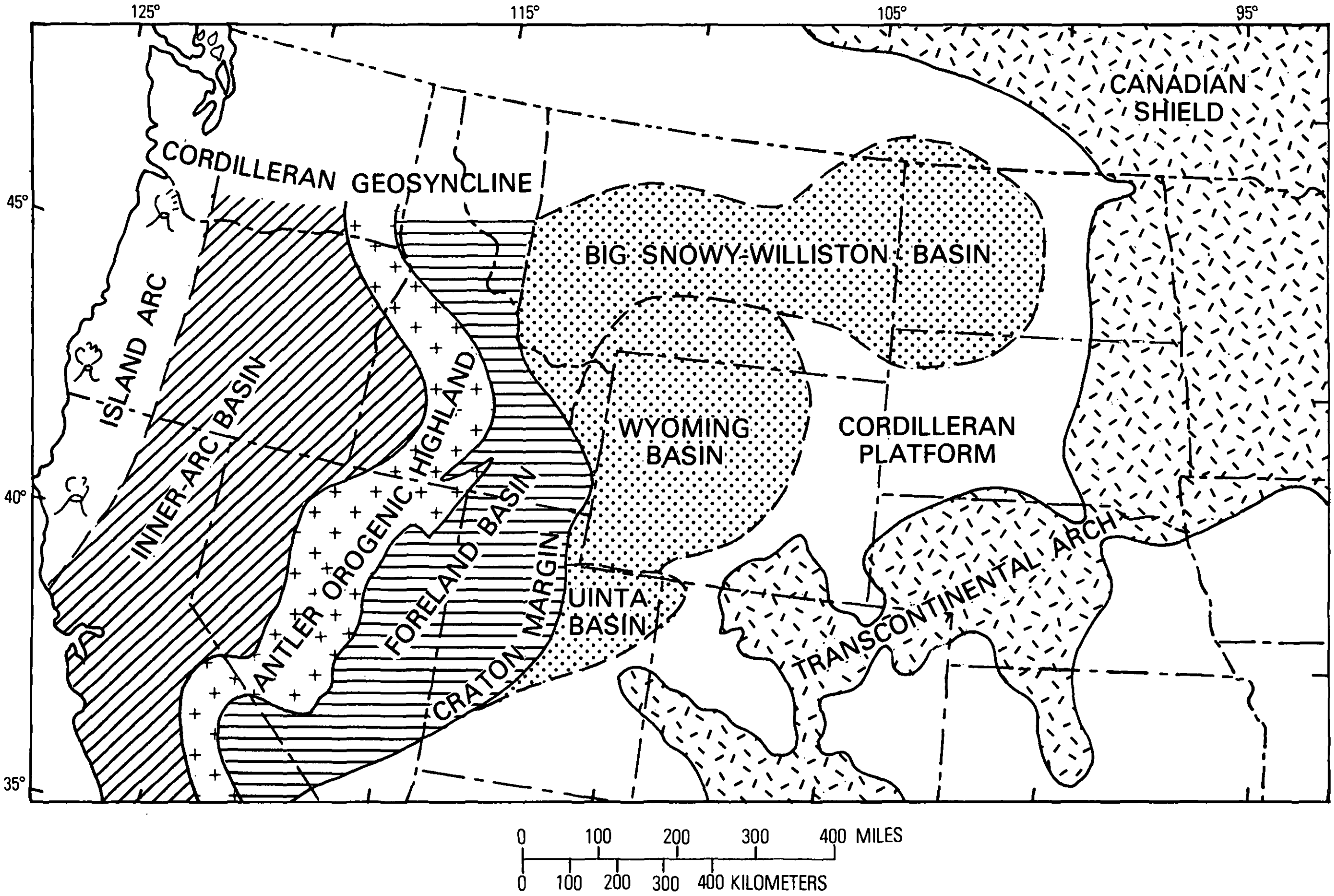


FIGURE 7.—Paleotectonic map of Carboniferous deposition in western United States. Modified from Sando (1976a) and Poole and Sandberg (1977).

Early Osagean (early Tournaisian) time marked the beginning of regression, characterized by westward progradation of shelf carbonate sediments (Madison Limestone). The foreland basin in Idaho became a starved basin that received fine terrigenous sediment (lower part of Little Flat Formation) during the latter part of the Osagean. Restricted circulation on the Cordilleran platform caused evaporite deposition during several periods beginning in the middle or late Osagean (middle or late Tournaisian) and extending into the early Meramecian (early Viséan). A reflux system associated with evaporite deposition caused dolomitization of underlying shelf carbonate rocks.

During latest early Meramecian (early middle Viséan) time, epeirogenic uplift drained the Cordilleran platform in Wyoming, while the foreland basin continued to receive marine sediments. A karst topography began to form in Wyoming, and a river system began to carry terrigenous sediment from the Transcontinental arch across the platform to the foreland basin.

Karst development continued throughout the Cordilleran platform until late Meramecian (late Viséan) time, when subsidence of the platform margin permitted encroachment of the Cordilleran sea into Wyoming and southwestern Montana. Shelf carbonate deposits began to form in the foreland basin and to prograde eastward. Continued subsidence caused the sea to transgress eastward in three developing basins, the Big Snowy-Williston basin in Montana, the Wyoming basin in Wyoming, and the Uinta basin in Utah and Colorado (fig. 7), each basin receiving terrigenous sediments from the Transcontinental arch on the east.

In Wyoming, a basal transgressive sand (Darwin Sandstone Member of the Amsden Formation and its equivalents), composed in part of reworked river sediment, was deposited across the Wyoming basin, filling irregularities and solution cavities in the underlying karst terrane. The sand was followed by an eastward-expanding red-bed facies (Horseshoe Shale Member of the Amsden Formation) deposited in a lagoonal environment, produced in part by restriction of the mouth of the Wyoming basin. The eastern margin of the eastward-prograding carbonate shelf of the foreland basin protruded into the mouth of the Wyoming basin in western Wyoming during middle Chesterian (late Viséan and early Namurian) time (Moffat Trail Limestone Member of the Amsden Formation). In late Chesterian (late Namurian) time, a restricted carbonate-terrigenous facies (Ranchester Limestone Member of the Amsden Formation) began to form at the mouth of the Wyoming basin.

The three facies of the upper depositional cycle—sandstone, red beds, and restricted carbonate-terrigenous rocks—continued to trangress eastward across the expanding Wyoming basin until Morrowan (latest Namurian and earliest Westphalian) time, when the shallow sea of the Wyoming basin breached a narrow land area that separated it from the Big Snowy-Williston basin in Montana. Hence, Pennsylvanian rocks rest disconformably on Mississippian rocks only in eastern Wyoming, where the upper depositional transgression did not cover the karst plain composed of lower depositional sequence rocks until Morrowan time. Expansion of the Wyoming basin continued into the Atokan (early Westphalian), followed by the influx of a flood of quartz sand (Tensleep Sandstone and equivalents) during later Atokan and later Pennsylvanian time.

PENNSYLVANIAN (UPPER CARBONIFEROUS) SYSTEM OF WYOMING

By Edwin K. Maughan

INTRODUCTION

Carboniferous rocks in Wyoming originated chiefly as marine sediments and are divided into lower, chiefly carbonate, deposits of Mississippian age, and upper, chiefly terrigenous, detrital sediments of Pennsylvanian age. Four cycles of deposition are recorded in these Carboniferous rocks, which accumulated upon the shelf between the North American craton and the Cordilleran miogeosyncline. The lower two depositional sequences, the Madison Limestone and the Big Snowy Group or their equivalents, are of Mississippian age; the upper two sequences, the Amsden Formation and the Tensleep Sandstone, or their equivalents, are of Pennsylvanian age. Regional unconformities separate the four sequences. Generally, red mudstone strata that mark the initial deposits of the Pennsylvanian rest unconformably either upon remnants of the upper Mississippian sequence, which comprises strata of Late Mississippian (Chesterian) age, or upon the lower Mississippian sequence, which comprises the Madison Limestone of Kinderhookian, Osagean, and Meramecian age. An exception is in the southern part of the Laramie and the Medicine Bow Ranges (fig. 8), where Pennsylvanian strata rest unconformably upon Precambrian rocks. The upper Pennsylvanian sequence generally rests unconformably upon strata of the lower Pennsylvanian sequence, but locally, erosion has completely stripped away the lower Pennsylvanian strata, and the upper sequence rests upon rocks as old as the Madison.

Upper Carboniferous strata are exposed along the flanks of most mountain ranges in the State and in thrust plates within the westernmost mountains (fig. 8). Drill logs and cores from boreholes in most structural basins provided lithologic and paleontologic data to establish continuity of the strata between outcrops throughout the State.

Pennsylvanian strata were deposited in Wyoming in shallow marine water of the continental shelf. Highlands of the ancestral Rocky Mountains lay to the south in Colorado, and a prong of these mountains, the ancestral Front Range (fig. 9), projected northwestward into south-central Wyoming approximately coincident with the present-day Sierra Madre. The low Siouxia landmass was east of Wyoming, but shallow seas between the ancestral Front Range and Siouxia extended through southeastern Wyoming and across eastern Colorado, western Nebraska, and Kansas into Oklahoma and the Ouachita geosyncline (McKee, Crosby, and others, 1975, pl. 15A). The Cordilleran miogeosyncline and related geanticlinal or island-arc areas lay to the west in central Idaho. Northward, in Montana lay a continuation of the shallow shelf where shoal deposits similar to those deposited on the shelf in Wyoming characterize most of the Pennsylvanian rocks. The Bannock highland (Williams, 1962) was a probable land area in southeastern Idaho and adjacent

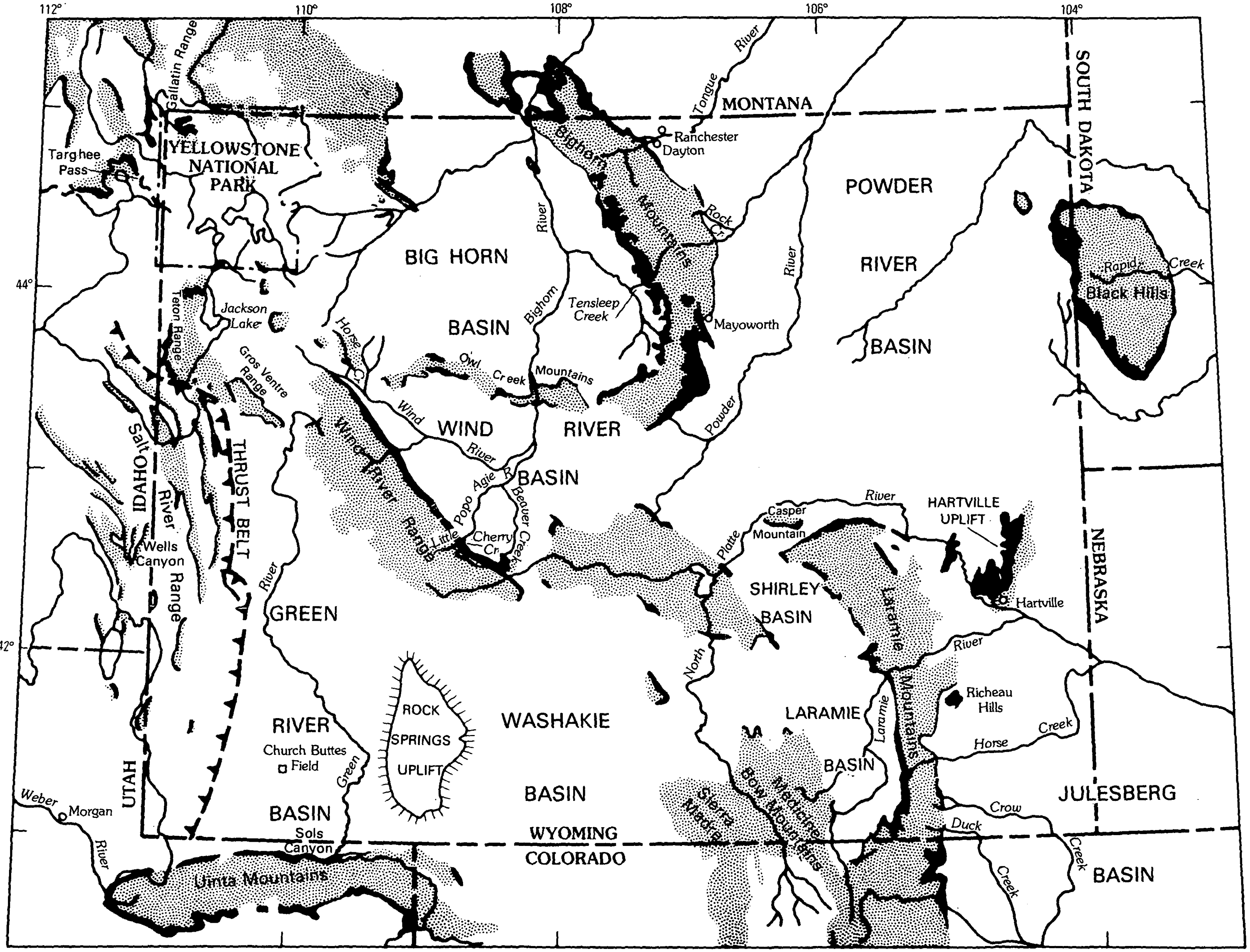


FIGURE 8.—Index to localities referred to in text. Outcrop of Pennsylvanian rocks shown in black and those of older rocks, by stipple pattern.

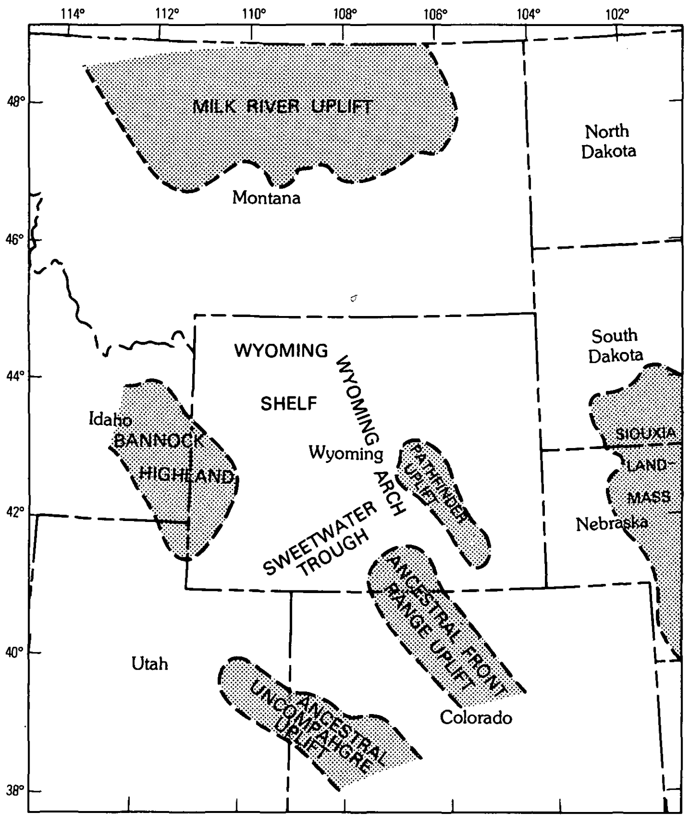


FIGURE 9.—Paleogeographic elements affecting deposition and preservation of Pennsylvanian rocks in Wyoming and adjacent States.

parts of western Wyoming in Early to Middle Pennsylvanian time, as was the Pathfinder uplift (Mallory, 1963) in east-central Wyoming; however, these land areas seem to have been low lying or were quickly eroded to base level and subsequently buried beneath sediments of the Upper Pennsylvanian sequence in Middle to Late Pennsylvanian time.

PREVIOUS WORK

Excellent summaries describing the Pennsylvanian rocks in most of Wyoming have been written by Williams (1962) and Mallory (1967, 1975). The present report is taken largely from these works, minor changes being drawn chiefly from newly available data or from slightly different interpretations of the data for which I am solely responsible. Other important contributions to the understanding of the vertical sequence and the lateral correlations of the Pennsylvanian rocks have been those of Condra, Reed, and Scherer (1940), who established correlations between surface exposures of the Minnelusa, Hartville, and Casper Formations; Love, Henbest, and Denson (1953), who carefully documented the fusulinids in the Hartville Formation; Love (1954), who correlated Pennsylvanian rocks in an east-west section in Wyoming; Agatston (1954), who summarized much surface and subsurface data of the Pennsylvanian rocks in northern and eastern Wyoming; and Foster (1958), who divided the Minnelusa Formation into three members bounded by regional unconformities and described the thickness and lithofacies of these members in northeastern Wyoming. Faunal data for the lower Pennsylvanian sequence and the lower part of the upper sequence were contributed by Sando, Gordon, and Dutro (1975). Many other investigations are referred to in the text where appropriate.

ACKNOWLEDGMENTS

I am grateful to H. D. Thomas, former Wyoming State Geologist, who initially encouraged my study and suggested answers to some of the questions regarding Pennsylvanian rocks in the Laramie Range and adjacent parts of southeastern Wyoming. F. F. Wilson helped to develop the concepts relative to the stratigraphic, paleogeographic, and tectonic framework along the northern Front Range in Colorado and in southeastern Wyoming; A. E. Roberts made similar contributions in Montana. Discussions with W. W. Mallory, H. R. Wanless, and J. D. Love were especially useful to test and extend these concepts in other parts of Wyoming. Many individuals and publications have provided fragments of information and have contributed to this report.

TECTONIC FRAMEWORK

Tectonic movements in Wyoming during the Pennsylvanian were subtle and seem to reflect epeirogeny peripherally related to strong orogenic disturbances in the mobile Cordilleran geosyncline. Repeated movement of the ancestral Rocky Mountains south of Wyoming also seems to have been approximately contemporaneous with orogenic movements in the geosyncline. Thus, low-magnitude epeirogenic adjustments in Wyoming that included arching, downwarping, and possibly faulting, subtly affected the provenance of the sediments, lithological constituents, environments of deposition, and preservation of the strata from place to place in the State.

Pennsylvanian deposition succeeded regional uplift, erosion, and karstification of the Mississippian strata. All of Wyoming seems to have been emergent at the beginning of the Pennsylvanian Period. Chesterian age rocks equivalent to the Big Snowy

Group of Montana and the Great Blue, Doughnut, and part of the Manning Canyon Formations of northern Utah are absent in most of the State, except for a wedge of Chesterian strata in the westernmost part and local remnants of the basal sandstone of the Chesterian sequence in the center and the east. Chesterian-age rocks of the upper Mississippian sequence probably were deposited across most of Wyoming; they were subsequently removed by erosion during latest Mississippian and Early Pennsylvanian uplift, except for basal sandstone remnants, which are the Darwin Sandstone Member of the Amsden Formation and the probably equivalent Fairbank Formation, in eastern Wyoming. In the western Wyoming thrust belt, the beveled wedge of these Upper Mississippian strata is preserved unconformably below Pennsylvanian strata where the Moffat Trail Limestone lies above red beds and sandstone strata that seem equivalent to the Darwin Member. Elsewhere in the State, where pre-Pennsylvanian uplift and erosion removed the Chesterian strata, Pennsylvanian rocks lie unconformably upon Kinkerhookian-to-Meramecian Madison or equivalent rocks. In southeastern Wyoming, in the vicinity of the ancestral Front Range uplift (fig. 9), Mississippian-age strata are entirely absent, and Pennsylvanian strata lie unconformably upon Precambrian rocks.

As the Pennsylvanian sea encroached upon the Wyoming shelf in late Morrowan time, paralic sedimentation gave way to marine littoral deposition across all of the State, except in the south-central part, where the northwest-projecting prong of the ancestral Front Range uplift remained. Initially, the upper Morrowan sediments, composed chiefly of highly oxidized and reworked soils that had formed during the Late Mississippian to Early Pennsylvanian uplift of the Wyoming shelf, were red terrigenous muds and very fine grained quartzose and arkosic sands deposited in shallow water of the advancing sea. However, carbonate sediments were the chief deposits in the Sweetwater trough (Mallory, 1975, p. 266) that projected northeastward from Utah into southwestern and central Wyoming. By early Atokan time, the sea had inundated the Wyoming shelf and adjacent parts of the cratonic margins along the Cordilleran geosyncline; the influx of terrigenous sediments was diminished, and carbonate sediments (chiefly lime mud) were widely deposited.

Middle to late Atokan time saw a second pulse of epeirogenic movement and renewed uplift that brought the ancestral Front Range and the ancestral Uncompahgre elements to their maximum relief. This Atokan epeirogeny was nearly contemporaneous with mid-Pennsylvanian orogenic movements in the mobile belt of the Cordilleran geosyncline in Nevada. The Wyoming shelf was warped and possibly locally faulted by this epeirogeny, so that other uplifts of low relief, such as the Bannock highland in western Wyoming and eastern Idaho (Williams, 1962, p. 159, 172) and the Pathfinder uplift in east-central Wyoming (Mallory, 1963), were rejuvenated or newly formed as parts of the ancestral Rocky Mountains. The previously formed Morrowan- and early Atokan-age sediments were stripped from these uplifts and were beveled in areas adjacent to them. Other areas that had been arched upward but had not necessarily been exposed above sea level were beveled also. Maximum relief of the Pathfinder uplift was probably about 150–200 m, as determined from the thickness of the earlier Pennsylvanian sediments preserved in adjacent downwarped areas and from the thickness of later Pennsylvanian sediments that lap onto the uplift. Uplift of the Bannock highland seems to have been of comparable magnitude, although it may have been as much as 300 m.

Deposition of the upper sequence of Pennsylvanian strata in Wyoming began at the time of the mid-Pennsylvanian epeirogeny. As the ancestral Rocky Mountains rose, sediments stripped from them were deposited in adjacent lows. These initial sediments of late Atokan to early Des Moinesian age were alternating beds of argillaceous and magnesian or lime mud and some thin beds of quartz sand. As the low areas filled and the uplifted areas were leveled in late Atokan and Des Moinesian time, quartz sand from distant sources to the west and northwest spread across western and central Wyoming to form an extensive sand blanket. Tongues of sand projected eastward from this large sand body into the alternating sand and carbonate sediments of eastern Wyoming. The ancestral Front Range uplift rose to its maximum relief and was a local source of arkosic sediment deposited at this time around the flanks of the uplift, but these ancient mountains probably contributed relatively little to the bulk of the sediments elsewhere in Wyoming.

Sand continued to be deposited in western and central Wyoming in Late Pennsylvanian time, but in eastern Wyoming, chiefly carbonate sediments were deposited during Missourian time and overlapped onto the Middle Pennsylvanian sand deposits of central Wyoming. This relation suggests diminishing delivery of sand from the western sources in Missourian time; however, another pulse of epeirogeny is recorded by an increase in the terrigenous components of the Virgilian sediments. This late Pennsylvanian epeirogeny led to the ending of Pennsylvanian deposition in Wyoming as uplift took place along a north-trending axis in central Wyoming, the Wyoming arch (Thomas, 1949). Virgilian-age sediments either were not deposited or were removed from most areas in the State, except in the far west and east. Missourian-age strata were entirely or mostly removed from the broad crest of the Wyoming arch, where rocks of Des Moinesian age are unconformably overlain by Permian strata.

Permian rocks record transgressive onlap of the Pennsylvanian strata on the Wyoming arch. Wolfcampian deposits occur in both western and eastern Wyoming, and sediments of probable late Wolfcampian to early Leonardian age were deposited thinly across a broad saddle on the arch in the central part of the State. Northward, the arch merged with the Milk River uplift in Montana (Maughan, 1966a), and southward it merged into the ancestral Front Range. Strata of late Leonardian age wedge out against higher parts of the arch in north-central and south-central Wyoming. Lower Guadalupian strata overlap the older Permian strata and rest on Pennsylvanian rocks in both northernmost Wyoming and adjacent parts of south-central Montana and in southernmost Wyoming and Colorado in the vicinity of the ancestral Front Range. The ancestral Front Range persisted as a land feature, and upper Permian strata rest unconformably upon Precambrian rocks beyond the limits of deposition of the Pennsylvanian strata on the flanks of this ancient land area.

STRATIGRAPHY

Pennsylvanian strata in Wyoming (fig. 10) comprise the Amsden Formation and the Tensleep Sandstone in the Bighorn Mountains and adjacent parts of central Wyoming westward to the Gros Ventre and Teton Ranges. In eastern Wyoming, they include part of the Minnelusa Formation in the Black Hills and adjacent basins, part of the Hartville Formation in the Hartville uplift, and part of the Casper Formation and the Fountain Formation in the Laramie Mountains. In southwestern Wyoming, Pennsylvanian rocks in the subsurface of the Green River basin include the Round Valley Limestone, Morgan Formation, and Weber Sandstone. In the ranges of the thrust belt in westernmost Wyoming, Pennsylvanian rocks are included in the Wells Formation. In the Gallatin Range in Yellowstone National Park, Pennsylvanian rocks comprise the Amsden Formation and Quadrant Quartzite.

CASPER, HARTVILLE, AND MINNELUSA FORMATIONS

The Casper, Hartville, and Minnelusa Formations are lithologically and sequentially similar and should be identified as a single formation. However, early studies and descriptions of these formations at isolated, widely separated exposures in the Laramie Mountains, the Hartville uplift, and the Black Hills have led to the use of separate names in each of these areas.

The Minnelusa Formation was the first to be studied. The name was initially used by Winchell (1875) and was later modified by Darton (1901) for the sandstone and limestone above the Pahasapa Limestone of Mississippian age and below the Opeche Shale of Permian age exposed along Rapid Creek in the Black Hills. Minnelusa is the Dakota Sioux name for Rapid Creek. Condra, Reed, and Scherer (1940) divided the Minnelusa into six units, which they correlated with similar divisions in the Hartville Formation. Two regional unconformities within the Minnelusa (fig. 10), provide boundaries for lower, middle, and upper members of this formation (Foster, 1958; Maughan, 1966b). The lower member of the Minnelusa is of Morrowan and Atokan age; the middle member is of Des Moinesian, Missourian, and Virgilian age; and the upper member is of Wolfcampian age.

The Hartville Formation was named by W. S. T. Smith (1903) for exposures of sandstone, carbonate rock, and mudstone between the Guernsey Limestone of Mississippian age and the Opeche Shale of Permian age in the hills around the town of Hartville, Wyo. Condra and Reed (1935) and Condra and others (1940) divided the Hartville into six units and gave names to them derived from geographic features in the vicinity of Guernsey Reservoir on the North Platte River in the Hartville uplift. They named the lowest unit the Fairbank Formation and designated the other units as "groups." The groups of Condra and others subsequently have been modified in rank, and the names have been used to identify members of the Hartville Formation (Tranter and Petter, 1963; Hoyt, 1963). The Fairbank Formation probably is of Mississippian age (Love and others, 1953). An informal threefold division into lower, middle, and upper members has been used (Maughan, 1966b, p. 97) and is adopted in this report in order to simplify comparisons with similar subdivisions of the Minnelusa and Casper

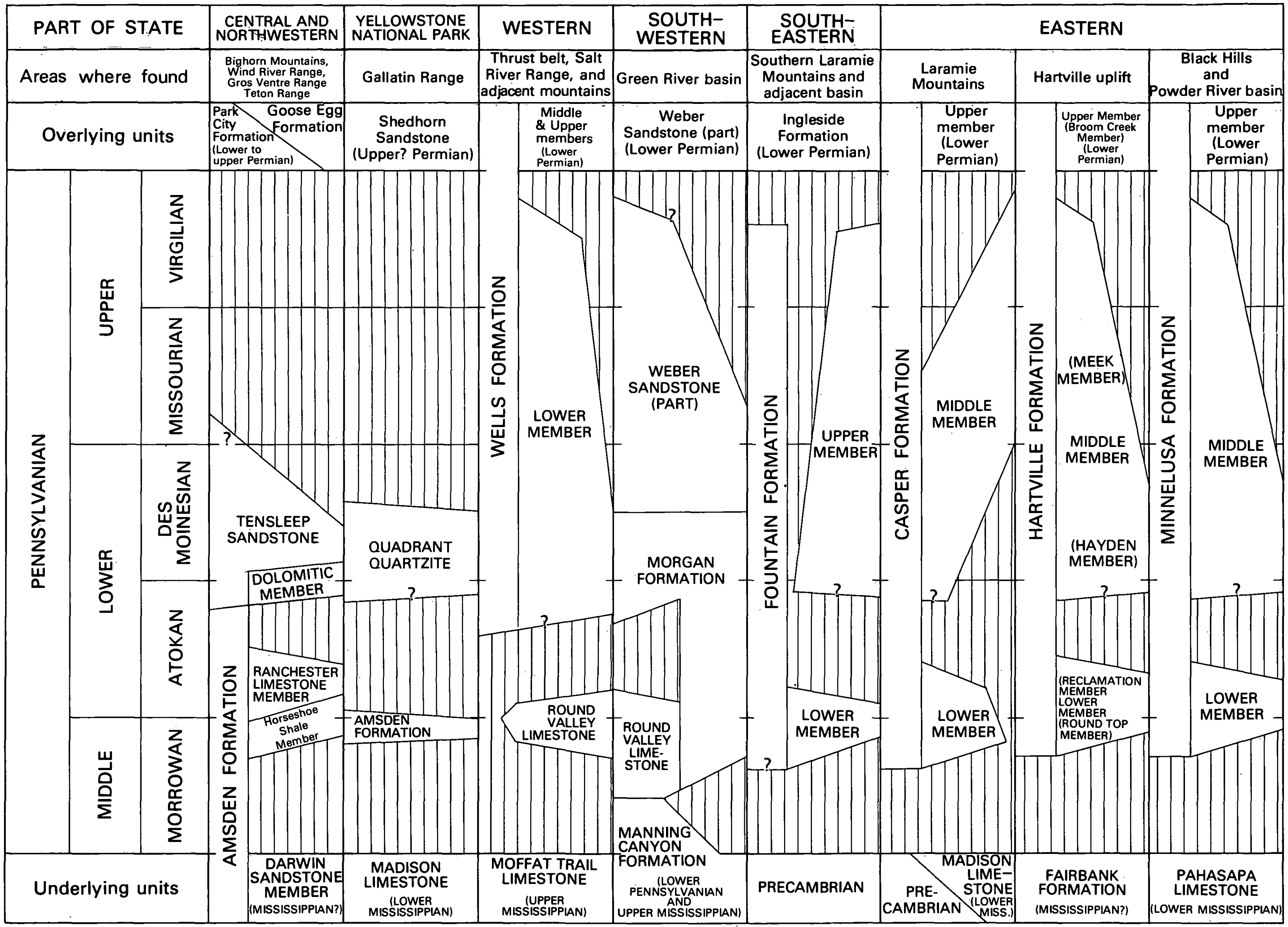


FIGURE 10.—Summary of nomenclature and correlation of Pennsylvanian rocks in Wyoming. Boundaries queried where ages uncertain; names in parentheses indicate usage of Tranter and Petter (1963).

Formations. The lower member, which excludes the Fairbank Formation and includes, in ascending order, the Reclamation and Roundtop Members of Tranter and Petter (1963), is of probable Morrowan and of Atokan age; the middle member, which includes the Hayden, Meek, and Wendover Members, is of Des Moinesian, Missourian, and Virgilian age; and the upper member, which includes the Broom Creek and Cassa Members, is of Lower Permian age (Love and others, 1953).

The Casper Formation was named by Darton (1908) for strata he believed to be equivalents in the Casper and Laramie Mountains of the Lower to Middle Pennsylvanian Amsden and Tensleep Formations. The Casper is more inclusive than Amsden and Tensleep as it also comprises strata of Late Pennsylvanian and Early Permian ages. In the southern Laramie Range, the Casper Formation lies unconformably upon Precambrian rocks, and in the northern part of the Laramie Range and on Casper Mountain, the formation is unconformable upon Madison Limestone of Early Mississippian age. The Casper Formation may be divided into three members similar in lithology and age to the members of the Hartville and Minnelusa Formations (Maughan, unpub. data, 1977). However, the lower member is absent at many places in the Laramie Range, and strata of the middle member ranging from Des Moinesian to Missourian in age form the base of the Casper Formation at these places. Onlap of the middle member northward along the west flank of the Laramie Mountains was recognized by Thomas and others (1953); similar onlap occurs on the east flank of the range. The lower member is present in outcrops along the east flank south of Horse Creek to the Colorado border, and probable remnants of the lower member are exposed at the base of the formation at places elsewhere in the range. The absence of the lower member and the onlapping of the middle member are evidence for the Pathfinder uplift. Uplift took place after deposition of the lower member. Remnants of the lower member that lie unconformably below the middle member must have formed prior to deposition of the onlapping middle member of the Casper. Members of the Casper Formation tongue southward into the lower and upper parts of the Fountain Formation of Pennsylvanian age and into the Ingleside Formation of Early Permian age in Colorado (Maughan and Wilson, 1960).

Basal strata in the Minnelusa, Hartville, and Casper Formations are chiefly red mudstone interstratified with fine-grained sandstone and thin beds of light-gray to pale-purple and pale-red-purple limestone. Detrital sediments predominate in the lowermost part of the Casper Formation because deposition took place near sources of terrigenous sediments in the ancestral Front Range; carbonate rocks predominate in the Minnelusa because of its more central position in the depositional basin. The mudstone and carbonate beds that constitute the lower member of these formations are commonly as thick as 70 m. However, thickness differs considerably from place to place, largely because of erosion prior to deposition of the overlying middle member. The lower member of the Casper Formation is beveled, and, locally, in the northern part of the Laramie Mountains, it is absent from the area of the Pathfinder uplift.

The limestone of the lower member is mostly light gray, although it commonly may be pale purple or pale red-purple. Commonly it contains red to dark-brown, white, and dark-gray nodules and stringers of chert. The limestone beds are generally 20–50 cm thick and are commonly interbedded with red or gray clayey mudstone ranging from a thin film to a few centimeters in thickness.

The late Morrowan and Atokan age of the lower member, which constitutes the lower Pennsylvanian sequence in eastern Wyoming, is best documented in sections of the Hartville Formation by the fusulinid zones of *Millerella, Profusulinella,* and *Fusulinella* (Love and others, 1953). *Millerella marblensis, Millerella pinguis, Profusulinella,* and *Fusulinella* (with one possible exception) are restricted to the lower member. *Millerella marblensis* and *M. pinguis* range through the lower part (Reclamation Member) and into the upper part (Roundtop Member) to an horizon about 34 m above the base of the lower member. *Profusulinella,* which is represented in a collection as low as 6.7 m above the base, indicates an Atokan age for the higher beds in the lower member; the lower 6 m probably is of Morrowan age. *Fusulinella* ranges from 10.5 m above the base to the top of the lower member, and the species represented indicate early Atokan age.

Fusulinella? cadyi, of Des Moinesian age, is an exception in the genus of *Fusulinella,* which usually indicates Atokan age. Love and others (1953) suggested that this Des Moinesian species may be regarded as possibly nearer the genus *Fusulina* than it is to the genus *Fusulinella.*

The lower member of the Casper Formation has not yielded fusulinids, although about 50 m of undated strata below the middle member on the eastern flank of the Laramie Mountains is presumably of early Atokan and Morrowan age. *Fusulinella* is represented in collections from the lower Minnelusa in the southern Black Hills (Thompson, 1936; Jennings, 1959), indicating a late Morrowan or early Atokan age of these strata in that area.

The middle member of the Casper, Hartville, and Minnelusa Formations comprises interbedded dolomite, calcareous cemented quartzose sandstone, and mudstone. Generally these strata are light to medium gray and yellowish gray, except in the Casper Formation where the middle member commonly is pinkish gray, especially in outcrops in the southern part of the Laramie Mountains. In that area, the strata tongue into the red arkosic upper member of the Fountain Formation.

Beds in the lower part of the middle member are mostly 20–50 cm thick, but in the upper part of the member they reach a maximum thickness of 10 m. Thin carbonaceous shale beds found in the lower part of the middle member in central eastern Wyoming are exposed in the Hartville uplift and in the southern Black Hills. These dark-gray carbonaceous shale beds grade laterally to pale olive green where exposed in the northern Black Hills and in the Laramie Mountains. In the subsurface of the southern Powder River basin, these carbonaceous strata are conspicuous on radioactivity logs.

Foraminifera from the upper Pennsylvanian sequence in eastern Wyoming indicate Des Moinesian to Virgilian age. *Fusulina* occurring as much as 54 m above the base of the middle member of the Hartville Formation indicates Des Moinesian age for the lower part of the member; *Wedekindellina ultimata* and cylindrical forms of *Triticites* indicate Missourian age for an approximately 40-m-thick middle part; and ventricose forms of *Triticites* indicate Virgilian age for as much as 20 m that constitutes the upper part of the middle member.

The zones of *Fusulina, Wedekindellina,* and of *Kansanella* (*Triticites* of cylindrical form) indicate the Des Moinesian to Missourian age of the middle member of the Casper Formation (Thomas and others, 1953; Thompson and Thomas, 1953; L. G. Henbest, written commun., 1964). Most collections from the lowermost strata of the middle member of the Casper are of possible late Atokan age according to Henbest (written commun., 1964). The zones of *Fusulina, Wedekindellina,* and *Kansanella* also occur in the middle member of the Minnelusa Formation (Thompson, 1936; Jennings, 1959; Verville and Thompson, 1963).

The upper member of the Minnelusa, Hartville, and Casper Formations is of Wolfcampian age and rests unconformably upon strata of the middle member, which in some areas are as old as Des Moinesian. At the base of the upper member in most of eastern Wyoming is a red sandy to argillaceous mudstone known as the red marker. This distinctive stratum, believed to be a regolith or redistributed regolith, marks an erosional contact between Pennsylvanian and Permian rocks. The red marker crops out in the southern Black Hills, the Hartville uplift, and the northern Laramie Mountains. The red marker is readily identified in subsurface logs in parts of the basins adjacent to surface exposures. The red marker extends northward to about the middle of the Powder River basin and seems to die out there and to the west on the eastern flank of the Wyoming arch, just short of surface exposures of this horizon on the southeast flank of the Bighorn Mountains. Its extent southward in the subsurface of the Julesberg basin is uncertain because of sparse drilling into Pennsylvanian rocks of southeastern Wyoming, but the red marker has been identified in westernmost Nebraska (Hoyt, 1963). Red beds similar to the red marker are found locally in subsurface logs of the Laramie basin and at the base of Permian rocks in some exposures on the flanks of the adjacent Medicine Bow Mountains, beyond the area in central Wyoming where the red marker seems to be regionally persistent.

FOUNTAIN FORMATION

The name Fountain Formation is used in south-central Wyoming for arkosic sandstone and mudstone of Pennsylvanian age where they crop out on the flanks of the southern part of the Laramie Mountains and in the Medicine Bow Mountains. This formation was named by Cross (1894) for red arkose exposed along Fountain Creek near Colorado Springs, Colo. The Fountain is exposed along the east flank of the Rocky Mountains in Colorado northward 230 km to the Wyoming State line. The Fountain Formation in Colorado is of Pennsylvanian and Early Permian age (Maughan and Wilson, 1960), and the rocks grade northward from alluvial-fan deposits through fan-delta sediments into littoral marine deposits. In the vicinity of Lyons, Colo., the Fountain is divided into three units: (1) the lower Fountain of probable Morrowan and Atokan age; (2) the upper Fountain north of Lyons of Des Moinesian to Virgilian age; and (3) part of the upper Fountain south of Lyons, which northward merges into the Ingleside Formation, of Wolfcampian age. The lower Fountain tongues into the lower member of the Casper Formation in Wyoming, the upper Fountain tongues into the middle member of the Casper in the vicinity of the Colorado-Wyoming State line, and the Ingleside merges into the upper member of the Casper (Maughan and Wilson, 1960).

ROUND VALLEY LIMESTONE

The Round Valley Limestone was named by Sadlick (1955, p. 56–57) for a thin-bedded limestone unit near Morgan, Utah, comprising rocks that lie unconformably above the Upper Mississippian Doughnut Formation and unconformably below the predominantly red mudstone and carbonate rocks of the Morgan Formation. The Manning Canyon shale of Mississippian and Pennsylvanian age conformably underlies the Round Valley at many places in the Wasatch Mountains and the Uinta Mountains in Utah, including the section at Sols Canyon (Anderman, 1955) illustrated in figure 11. The Round Valley previously had been included with the Morgan Formation, and this practice is currently followed in well logs that penetrate Pennsylvanian strata in southwestern Wyoming. Subsurface logs in the Green River basin show chiefly limestone in the lower Pennsylvanian (Verville and others, 1973; Verville and Momper, 1960, fig, 4, col. 2), assigned here to the Round Valley, which lies unconformably upon Mississippian limestone and below the restricted Morgan Formation. The Round Valley in Utah commonly has some red mudstone at the base and some dark-gray and locally reddish-brown mudstone in partings and thin layers interbedded with the thin-bedded limestone. The amount of interbedded mudstone in the formation increases eastward, and these rocks become mostly red mudstone eastward in the subsurface of the Green River basin. The strata gradually change from the dominant limestone facies of the Round Valley Limestone in southwestern Wyoming into the red mudstone and limestone facies that constitutes the Horseshoe Shale and Ranchester Limestone Members of the Amsden Formation in central Wyoming (fig. 11) and the lower member of the Casper Formation and equivalent strata in eastern Wyoming.

Foraminifera in the Round Valley Limestone belong in the zone of *Millerella* and indicate a Morrowan age for these rocks. Thompson (1945) identified *Millerella advena, M. inflecta, M. circuli, M.* cf. *M. pressa,* and *M.* aff. *M. marblensis* in the eastern Uinta Mountains from strata of the Round Valley Limestone that he called Belden Formation. Verville and Momper (1960) and Verville and others (1973) recognized the zone of *Millerella* and Morrowan age in rocks that are dominantly limestone in the lower part of the Morgan Formation in the subsurface of southwestern Wyoming. The zone of *Profusulinella* seems to be absent in southwestern Wyoming, although conodonts indicate that strata of early Atokan age are present, at least locally, in the subsurface of eastern Uinta County (Verville and others, 1973). Sadlick (1955, 1957, 1959) indicated an Early Pennsylvanian age, chiefly on the basis of brachiopod faunas in the Round Valley Limestone.

MORGAN FORMATION

The name Morgan Formation is commonly used in studies of the subsurface of southwestern Wyoming for Pennsylvanian rocks that lie unconformably upon the Mississippian Madison Limestone and that grade into overlying Middle Pennsylvanian Weber Sandstone. The formation was named by Blackwelder (1910) for exposures near Morgan, Utah, and the base was placed above a limestone unit of Pennsylvanian age, subsequently named the Round Valley Limestone (Sadlick, 1955). Gray limestone and black to dark-gray shale of the Round Valley Limestone have been included as basal strata of the Morgan Formation, as used in the subsurface of southwest Wyoming. Limestone beds of the Round Valley Formation have been excluded in this report from the Morgan of the subsurface; the restricted formation is like the Morgan at its type section and like outcrops along the flanks of the Uinta Mountains where the Morgan comprises reddish-brown sandstone, mudstone, and some dolomite or limestone. Sandstone is generally fine to medium grained, commonly calcareous, and white to light pink but weathers red. Dolomite and limestone beds are generally light gray, commonly cherty in white to gray and red to brown nodules and stringers.

The Morgan rests unconformably upon the Round Valley (Sadlick, 1955), and Sadlick (1957) considered the Morgan to be flysch derived from the ancestral Uncompahgre uplift. Williams (1962) suggested that sandstone and shale deposited at the type section is thicker than that deposited in the western Uinta Mountains because it was derived from the Bannock highland. He also stated that deposition of the Morgan took place in the Sweetwater trough between the Bannock and Uncompahgre land areas.

The Morgan Formation is of Middle Pennsylvanian age. The Hells Canyon Formation, an equivalent of the Morgan Formation on the southeastern

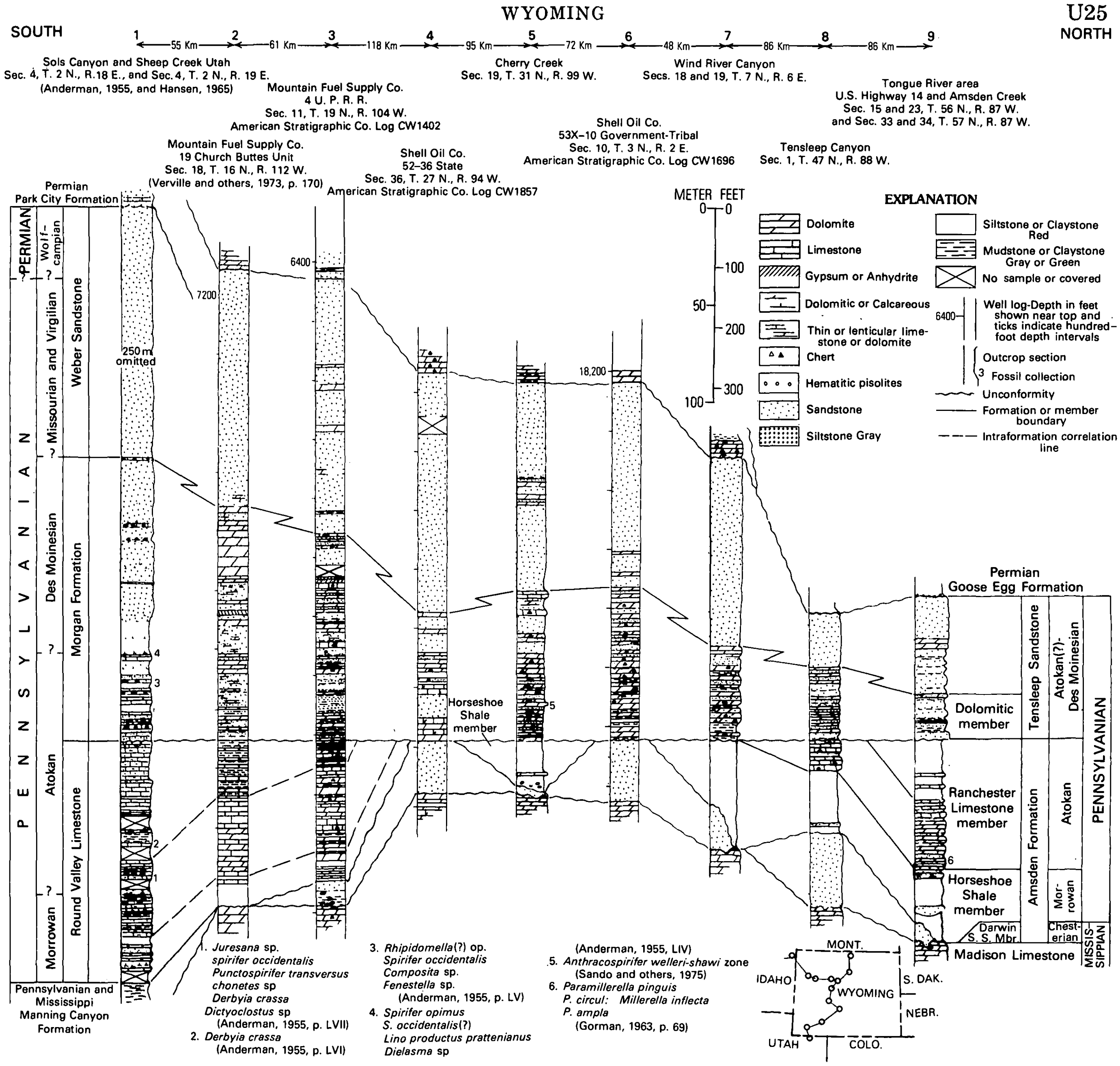


FIGURE 11.—Correlation of sections of Pennsylvanian strata in Wyoming from the north flank of the Uinta Mountains, Utah, to the Tongue River area on the northeastern flank of the Bighorn Mountains, Wyo.

flank of the Uinta Mountains includes *Pseudostaffella* cf. *P. keytei* var. *maccoyensis*, *Fusulinella iowensis* var. *leyi*, *F. lounsberyi*, *F. uintaensis*, and *F. haywardi* (Thompson, 1945, p. 43), fusulinids that are similar to species from the basal part of the Des Moinesian of New Mexico and Oklahoma. Sadlick (1955, 1957) and Kinney (1955, p. 43) assigned an early Des Moinesian age to the Morgan Formation. Sadlick (1955, p. 58) stated that "No fusulinids of Atokan age are known from [the Morgan Formation in] the Uinta Mountains." However, Verville and others (1973) stated that upper Atokan conodonts, fusulinids, and smaller forams as well as lower Des Moinesian forms occur in the Morgan Formation in eastern Uinta County, Wyo., in the Mountain Fuel Supply Co., No. 19 Church Buttes Unit well. Deposition of the Morgan strata probably was earlier in the Sweetwater trough than in areas of outcrop in the eastern Uinta and in the Wasatch Mountains—localities that were on the flanks of the ancient Uncompahgre and Bannock highlands.

WEBER SANDSTONE

The name Weber Sandstone (pronounced WEE-burr) is used in Wyoming in subsurface descriptions in the southwestern part of the State. The name was first used by Clarence King (1876) in referring to sandstone of the middle Coal Measures in Weber Canyon, Utah. Blackwelder (1910) defined the formation at a type locality in the upper narrows of the Weber River near Morgan, Utah, and described it as quartzite and sandstone lying above the redbeds of the Morgan Formation and below the dominantly carbonate rocks of the Permian Park City Formation. The Weber is exposed eastward of the type area on the flanks of the Uinta Mountains into northwestern Colorado. Age of the Weber in the type area is Middle (Des Moinesian) to Late Pennsylvanian, but in the Uinta Mountains, sandstone strata of Early Permian age are included (Bissell and Childs, 1958). Sandstone strata that commonly are identified as the Weber Sandstone in the subsurface of the Green River basin are indicated as ranging from about middle Des Moinesian to Early Permian in age (Verville and Momper, 1960).

WELLS FORMATION

Pennsylvanian rocks in the mountain ranges of the thrust belt in western Wyoming form the lower part of the Wells Formation. This formation was named by Mansfield (1927, p. 71) for sandy, cherty limestone and calcareous sandstone above the Mississippian Brazer Limestone and below the Permian Phosphoria Formation. The type section is in Wells Canyon, Caribou County, Idaho. The formation was divided by Mansfield into three members. The lower member, which constitutes the Pennsylvanian upper sequence in western Wyoming, is of Middle Pennsylvanian (Des Moinesian) age and the two upper members are of Lower Permian age, according to Williams (1962, p. 171). The Wells generally lies unconformably upon the Brazer Formation, but at places in eastern Idaho and western Wyoming, strata of the lower Pennsylvanian sequence intervene.

Local remnants of the lower Pennsylvanian sequence commonly are mapped with the Wells, although they consist of limestone and red mudstone beds that resemble the Round Valley Limestone and its equivalent, the Ranchester Limestone Member of the Amsden Formation. These remnants probably could be identified as Round Valley or Ranchester and mapped separately from the Wells. In eastern thrust plates of the thrust belt, red mudstone dominates over limestone in the lower part of these basal strata. These rocks are differentiated at some places into the Horseshoe Shale and Ranchester Limestone Members of the Amsden Formation.

The bulk of the Pennsylvanian rocks in the thrust belt are sandy carbonate rocks that are mostly sandy dolomite and dolomitic limestone interbedded with sandstone and mudstone that grade upward into mostly medium- to thick-bedded sandstone. For the most part, these rocks seem to be of Middle Pennsylvanian age, although some strata near the top of the sequence are Late Pennsylvanian (Wanless and others, 1955, p. 35) and are as young as Virgilian, as determined from sparse collections of fusulinids (G. J. Verville, oral commun., 1977).

The pattern of Middle Pennsylvanian deposition in western Wyoming seems to be like the pattern of deposition elsewhere in the northern Rocky Mountains region. The Upper Pennsylvanian sequence, consisting of part of the Wells Formation, was deposited upon an erosional unconformity succeeding mid-Pennsylvanian regional orogenic disturbance. At the type section of the formation in Wells Canyon, an area within the Bannock highland, Des Moinesian-age rocks lie unconformably upon Chesterian-age rocks (Williams, 1962). In the eastern part of the thrust belt, an area that would have been on the flank of the Bannock highland, the Wells Formation includes basal strata that have an Atokan age, probably late Atokan. In the Green River basin, an area within the Sweetwater trough, equivalent strata included in the Morgan Formation show little or no hiatus between late Atokan and underlying early Atokan strata (Verville and others, 1973), which indicates continuous or near continuous deposition from the Lower into the Upper Pennsylvanian sequence. These relations indicate transgressive overlap of middle Atokan to Des Moinesian strata from the trough onto the adjacent Bannock highland.

AMSDEN FORMATION

The Amsden Formation was named by Darton (1904). The Amsden comprises red shale, white to dark-gray limestone, and dolomitic limestone that lie unconformably upon the Madison Limestone and are overlain by the Tensleep Sandstone. The type locality is the vicinity of Amsden Creek, a tributary of the Tongue River, on the east flank of the Bighorn Mountains near Dayton, Wyo. Because of poor exposures, sparse fossils, and complex stratigraphic relations, the Amsden Formation remains the center of much controversy.

The Darwin Sandstone Member was added to the Amsden by Blackwelder (1918) for white-to-red sandstone unconformably resting upon the Madison Limestone. This sandstone was thought by him to grade into the overlying red shale. Blackwelder (1918, p. 442) indicated that the Darwin Member should be mapped separately from the overlying weakly resistant shale, sandstone, and limestone that compose the rest of the Amsden Formation.

A reference section for the Amsden on Amsden Creek, as described by Gorman (1963), is composed of four units: (1) the Darwin Sandstone Member at the base; (2) red mudstone, siltstone, and sandstone; (3) a dominantly carbonate rock unit; and (4) a dominantly clastic unit characterized by appreciable quartz sandstone and some shale and dolomite. Other sections in the type area were given by Agatston (1954, p. 569), who used the top of the carbonate rock unit as the top of the Amsden. Foster (1958) also showed an unconformity directly above the carbonate unit and correlated the Amsden with the lower member of the Minnelusa Formation; Maughan and Roberts (1967) showed correlation from the Amsden type locality with equivalent Pennsylvanian-age strata in central Montana; Mallory (1967) divided the Amsden above the Darwin into two parts and named the dominantly red mudstone lower part the Horseshoe Shale Member, and the dominantly carbonate rock upper part, the Ranchester Limestone Member; Sando and others (1975, p. A11) presented graphic comparisons of several measured sections of the Amsden in the type area.

The correlation of the Horseshoe and Ranchester Members by Sando and others (1975) with strata of late Meramecian and early Chesterian age in western Wyoming differs from the correlation presented in this report. The Moffat Trail Limestone and underlying red beds that include sandstone correlated with the Darwin are included by Sando and others (1975) as members of the Amsden Formation in the thrust belt. These rocks are a beveled wedge of the Upper Mississippian sequence of Chesterian age. They correlate lithologically, stratigraphically, and paleontologically with the similar Doughnut and Humbug Formations of northern Utah and the similar Big Snowy Formation of southwestern Montana. On the other hand, the Horseshoe Shale Member unconformably overlies the beveled upper Mississippian sequence (figs. 11, 12) and does not underlie or tongue into the Moffat Trail as shown by Sando and others (1975). Only the Darwin Sandstone extends east of the thrust belt, where it unconformably lies immediately beneath the Horseshoe Shale Member and is included as a member of the Amsden Formation. In the thrust belt in western Wyoming, the Chesterian-age red beds that include Darwin-like sandstone and the overlying Moffat Trail should not be included in the Amsden Formation.

The Darwin Sandstone Member comprises white to light-yellowish-gray, commonly red-stained, fine- to medium-grained, moderately sorted sandstone. Sand grains are mostly well-rounded to moderately well rounded quartz. Clay is common, and ferromagnesian grains range from sparse to abundant. The rock is weakly to moderately well cemented by calcite, and in places it is siliceous. Bedding consists mostly of low-angle cross laminae in planar beds as much as half a meter thick, but at most localities, crossbeds containing wedge-planar and trough crossbeds, having depositional dip angles as high as 20°, occur as sets that are as much as 3 m thick, and angles as high as 45° have been reported (Houlik, 1973, p. 503).

The Darwin Sandstone Member lies unconformably upon limestone beds of the Madison Group and was deposited above a karst surface on the Madison. It includes angular limestone blocks and fragments along the contact at most places. Red mudstone, siltstone, and sandstone as well as angular blocks and fragments of limestone occur as irregularly shaped cavern fillings that extend well downward into the upper part of the Madison. Terra rosa beneath sandstone at the base of the Darwin is uncommon, however, and at most places the quartzose sandstone rests directly upon the limestone.

The Darwin is unconformably overlain by red mudstone of the Horseshoe Shale Member. An unconformity above the Darwin is indicated in part by truncation of strata at the top of the member and by paleoweathering of the upper 10–50 cm of the sandstone. Most exposures of the weathered upper contact of the Darwin show moderate yellow staining not seen elsewhere in the sandstone, and weathering is evident at the top of the Darwin in cores from south-central Wyoming (Reynolds and others, 1975).

Thickness of the Darwin reaches a maximum of 50 m (Keefer and Van Lieu, 1966, p. B37), but the member is absent at many places throughout northwestern Wyoming, commonly within short distances of thick remnants of the sandstone. At the type section of the Amsden, Gorman (1963) indicated an unconformity 8.3 m above the base of the

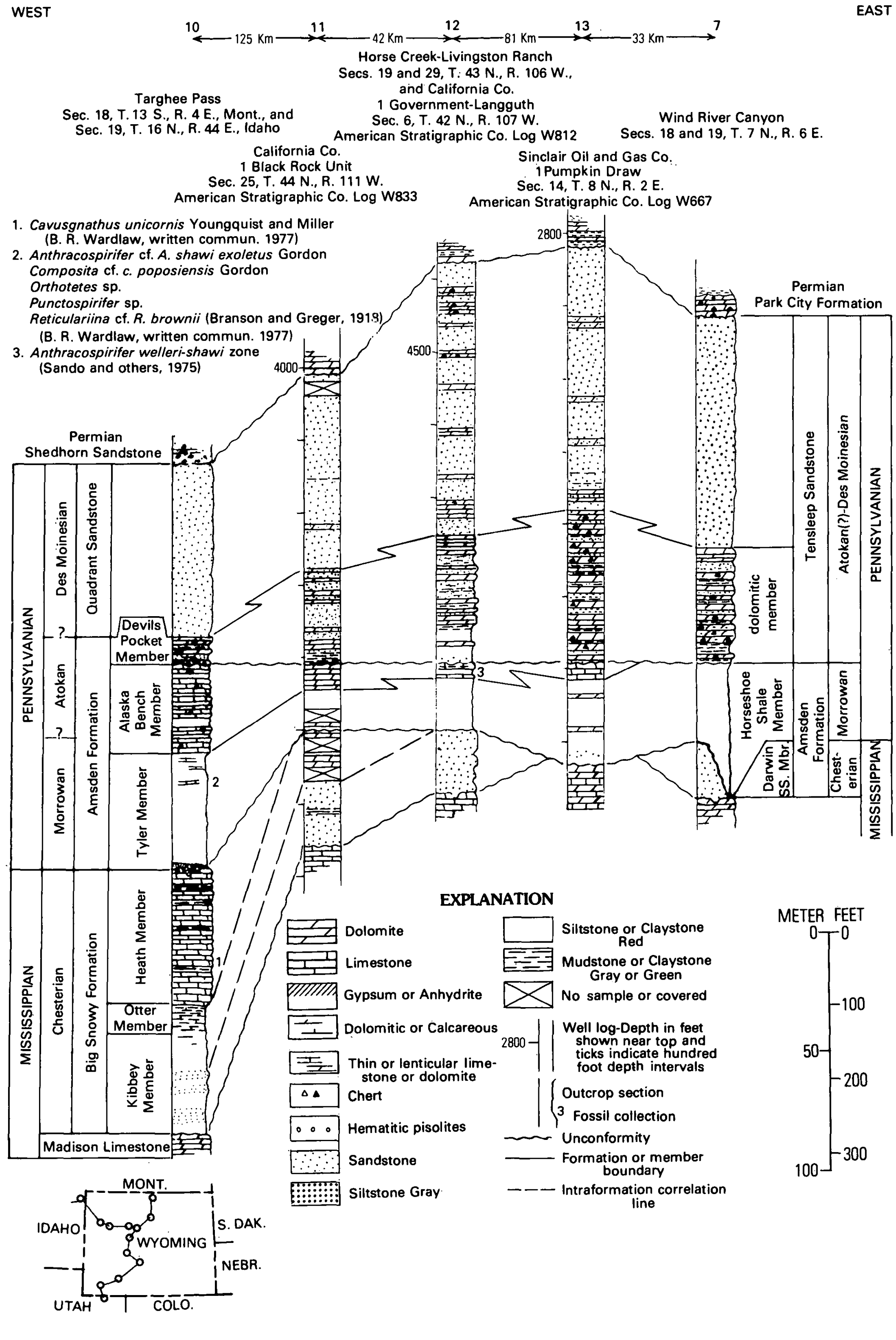


FIGURE 12.—Correlation of sections of Pennsylvanian rocks in northwestern Wyoming from Targhee Pass, Mont. and Idaho, to Wind River Canyon, Wyo.

Darwin. He indicated the total thickness of the Darwin as 13.1 m, but the Darwin-like sandstone above the unconformity is sand that seems to have been eroded from the Darwin and redeposited as a lensing channel-fill deposit in the base of the Horseshoe Shale Member. Because of erosion along this unconformable surface, the Darwin is absent 600 m west of Gorman's section, as shown by Maughan and Roberts (1967). Similar beveling of the Darwin Sandstone is evident in exposures of the Amsden in roadcuts along U.S. Highway 14 about 3 km southeast of the Amsden Creek locality. Beveling was recognized in the vicinity of Wind River Canyon (Maughan, 1972a, b), and it has been noted where the Darwin is locally absent in the vicinity of Horse Creek and Livingston Ranch in the Wind River basin and at many other places throughout the region.

Thickness variation of the Darwin has been attributed generally to deposition on the irregular erosional surface of the Madison (Agatston, 1954, p. 517; Keefer and Van Lieu, 1966, p. B37); however, the basal Darwin strata seem nearly conformable with the underlying Madison strata wherever the contact is not complicated by karst features (Houlik, 1973, p. 504–505). This applies even though the contact represents a significant hiatus, as indicated by the time required for the formation of karst and the regional transection of faunal zones in the Madison (Sando and others, 1975, p. A21). Most of the variation in the thickness of the Darwin is better attributed to erosional beveling at the top of the Darwin.

Age of the Darwin Member is considered to be Mississippian, probably early Chesterian, but this probable age determination is based on regional physical relations and lithologic similarities rather than on any direct paleontologic evidence. Fossils diagnostic of age have not been found in the Darwin. Brachiopods collected from the overlying Horseshoe Shale Member are believed to be of Pennsylvanian age, rather than Mississippian, as discussed elsewhere in this report. Thus, the age of the Darwin remains little less uncertain than when Blackwelder in 1918 (p. 423) noted that "the stratigraphic position of the sandstone indicates its age is early Pennsylvanian or late Mississippian—probably the former." Beds of sandstone that are very similar to the Darwin Sandstone compose part of an unconformable red-bed sequence that lies upon the Mississippian Madison Group and below limestone beds of Chesterian age in the western Wyoming thrust belt. This sandstone is correlated with the Darwin by Sando and others (1975).

Sandstone similar to the Darwin also makes up part of a red-bed sequence in the Kibbey Member of the Big Snowy Formation of Chesterian age, which lies upon the Madison Group in southwestern Montana and adjacent areas in Wyoming south of Yellowstone National Park (fig. 12, col. 10–12). These lithologically similar sandstones in the Big Snowy Formation, in the Chesterian of the thrust belt, and in the Darwin suggest that the strata are equivalent. Therefore, the age of the Darwin is probably Mississippian rather than Pennsylvanian. The correlation is possible only by comparing lithologic similarities, probably similar depositional environments, and regional stratigraphic relations.

The Horseshoe Shale Member of the Amsden Formation was named by Mallory (1967) and comprises chiefly red mudstone, siltstone, and fine-grained sandstone between the Darwin Sandstone Member and the overlying Ranchester Limestone Member at the principal reference section of the Amsden Formation. The name is from nearby Horseshoe Mountain, but the strata do not crop out there. Thin beds, generally less than 20 cm thick, of purplish- or reddish-gray argillaceous lime mud occur at some places, mostly in the upper part of the member. West of the type section, thin limestone beds are found within the member. Thickness of the Horseshoe generally ranges from 15 to 45 m, but the member is as much as 95 m thick near Casper and in the Shirley basin (Mallory, 1967, pl. 2B).

A few centimeters to a decimeter of pebble conglomerate may lie at the base of the Horseshoe Shale in northwestern Wyoming. Along Highway 14 on the northeast flank of the Bighorn Mountains, 24 m of medium-grained sandstone and mudstone, which originated as lenticular channel and overbank deposits, probably were derived chiefly from erosion of the Darwin. They constitute the lower part of the Horseshoe where it lies unconformably upon Madison Limestone and local remnants of the Darwin Sandstone Member. Above the channel and overbank deposits at Highway 14 and at most other places, the Horseshoe Shale Member mostly is composed of blocky to poorly fissile mudstone and siltstone and lesser amounts of clayey and very fine grained sandstone together with hematite and clay cement. Generally the member is covered with soil or talus and rarely is well exposed.

Pisolitic hematite, or possibly hematitic bauxite nodules, in beds that reach a maximum thickness of a meter, are common in the lower part of the member. The pisolitic beds lie at the base of the Horseshoe at many places; however, where channel and overbank deposits form the lower part of the member, the pisolitic beds, if present, lie between these and the overlying blocky mudstone that forms the upper part of the member. Biggs (1951, p. 21) suggested that the iron pisolites were formed in a shallow enclosed basin, an explanation similar to that proposed for deposition of the Silurian Clinton ores of the Appalachian region.

The Ranchester Limestone Member of the Amsden Formation was named by Mallory (1967) for the upper predominantly limestone part of the Amsden Formation at the principal reference section on Amsden Creek. The member is named for the town of Ranchester, about 16 km east of the type locality. The member comprises a lower predominantly limestone unit and an upper predominantly shaly unit. The Ranchester in the Tongue River area is unconformably overlain by sandstone, dolomite and shale strata that compose the lower part of the Tensleep Sandstone, although at most places in Wyoming, these overlying strata consist chiefly of dolomite and commonly are included in the Ranchester. In this report, these dolomitic strata are excluded from the Amsden Formation and are included as a dolomitic member of the Tensleep.

The Ranchester varies widely in thickness and at places is absent because of erosional beveling prior to deposition of the basal dolomitic member of the Tensleep. The Ranchester Limestone Member is part of the lower Pennsylvanian sequence. The unconformity above the Ranchester and beneath the dolomitic member was shown by Foster (1958) to extend throughout the Powder River basin and by Maughan and Roberts (1967), through Montana. This unconformity extends westward from the Bighorn Mountains and is an element in the Pennsylvanian strata in western Wyoming, as shown on the correlation of sections in Wyoming (figs. 11 and 12).

The Ranchester comprises mostly light- to medium-gray and reddish- to purplish-gray limestone and dolomitic limestone. Dolomite predominates at some places, but generally it is a minor component in this unit. The limestone beds are mostly 0.2–1.0 m thick and commonly are interstratified with red, at some places purplish, argillaceous mudstone that occurs as parting films or in thicknesses as great as 15 m. Limestone beds commonly contain nodules and stringers of red, brown, white, and dark-gray chert. The jasperoid chert is a useful lithologic key that aids identification of the member. Carbonate rock predominates where only the lower 20 m or so of the member is present, but in a few places, as at Amsden Creek and along Highway 14, 3 km southeast of the type section, the carbonate unit is about 45 m thick, and the upper part of the member comprises almost equal amounts of carbonate rock and mudstone. Quartzose sandstone rarely occurs in the Ranchester Limestone Member.

The Horseshoe Shale Member and the lower unit of the Ranchester Limestone Member correlate with the lower member of the Minnelusa Formation and lower member of the Hartville Formation (Foster, 1958) and constitute the lower Pennsylvanian sequence in central and northwestern Wyoming. Maughan and Roberts (1967, pl. 4) showed correlation of the Horseshoe and Ranchester with the lithologically similar Cameron Creek Member of the Tyler Formation and the Alaska Bench Limestone of Morrowan and early Atokan age in central Montana. Correlation of the Horseshoe and Ranchester members of the type locality of the Amsden Formation with the Pennsylvanian Round Valley Limestone in southwestern Wyoming and the Uinta Mountains is shown in figure 11. Similar stratigraphic relations westward to the Pennsylvanian Cameron Creek and Alaska Bench equivalents of the Amsden Formation in southwestern Montana are shown in figure 12.

Regional stratigraphic relations show that the Horseshoe in central Wyoming must be the same age as or slightly younger than equivalent red beds of late Morrowan to Atokan age in the lowermost member of the Hartville and Minnelusa Formations in eastern Wyoming and probably are slightly younger than limestone and red beds of Morrowan age in the Round Valley Limestone in southwestern Wyoming and northern Utah. That the Horseshoe Shale Member is younger is emphasized by onlapping of successively younger Lower Pennsylvanian conodont zones at the base of the Pennsylvanian strata from the Uinta Mountains northward into the Green River basin (G. J. Verville, oral commun., 1977). Paleogeographic evidence indicates that this onlapping probably continues to the outcrops at Cherry Creek near the southern end of the Wind River Range and into the Wind River basin, where the base of the Horseshoe Shale Member seems to be younger than the approximately equivalent Morrowan strata in the Green River basin and no older than about middle Morrowan. Similar physical onlapping of the Pennsylvanian Tyler Formation of the Amsden Group in central Montana with the Horseshoe Shale Member at Amsden Creek has been shown by Maughan and Roberts (1967, pl. 1).

The age of the Horseshoe Shale and the Ranchester is Morrowan, probably late Morrowan and early Atokan, although the members are sparsely dated paleontologically. Gorman (1963) identified fusulinids that characterize the zone of *Millerella* from limestone at the base of the Ranchester. These fusulinids are similar to the fauna from the same stratigraphic horizon in the Hartville uplift (Love and others, 1953). Fusulinids of the zone of *Millerella* also occur near the base of equivalent strata in central and southwestern Montana, where they are part of the Alaska Bench Limestone (Scott, 1945a,b; Mundt, 1956, p. 50). Early Pennsylvanian fossils occur on the east flank of the Bighorn Mountains within 1.5 m of the base of the Horseshoe at South Rock Creek (Sando and others, 1975, pl. 6) and 8 m above the base at U.S. Highway 14 (Gorman, 1963). Early Pennsylvanian age is indicated for the Horseshoe in the Wind River basin as the result of paleontological studies by Burk (1954) from collections made by Biggs (1951) 8.5 m above the base of the member at Beaver Creek, 8.7 m, at Cherry Creek, and 11.3 m, at Horse Creek.

The brachiopod assemblages from some Wyoming localities have been assigned a Mississippian age (Shaw and Bell, 1955; Shaw, 1955; Sando, 1967; Sando and others, 1975; Gordon, 1975). If these assemblages are of Mississippian age, they must be very late Chesterian; however, the Pennsylvanian age of the brachiopod faunas of the Horseshoe Shale Member in the Wind River basin is indicated by the regional relations shown in figures 11 and 12. As emphasized by Burk (1954, p. 4), the assemblages include many species known elsewhere in strata considered to be Pennsylvanian. Burk stated (p. 4) that "No genus or species is exclusively Mississippian or older" and "none of the previously identified faunas have a range which conflicts with a Pennsylvanian age for the Amsden." The Chesterian age assigned to faunal assemblages from the Horseshoe Shale and Ranchester Limestone Members collected at Cherry Creek (fig. 11) and at Livingston Ranch (fig. 12) on the flanks of the Wind River basin (Sando and others, 1975; Gordon, 1975) is at variance with the regional, physical-stratigraphic correlations presented in this report. The strata that have yielded the *Anthracospirifer welleri-shawi* faunal assemblage seem to be no older than middle Morrowan.

In agreement with the observations of Burk (1954), *Anthracospirifer welleri* (=*Spirifer welleri*) and other elements of the *A. welleri-shawi* zone have been identified in Pennsylvanian strata of the Morrowan to lower Atokan Cameron Creek Member of the Tyler Formation and Alaska Bench Limestone in the Big Snowy Mountains in central Montana (Maughan and Roberts, 1967). Elements of the *Anthracospirifer welleri-shaw* zone—*Anthracospirifer* cf. *A. shawi exoletus, Composita* cf. *C. poposiensis,* and *Reticulariina* cf. *R. brownii*—have also been identified (B. R. Wardlaw, written commun., 1978) from the Pennsylvanian Tyler equivalent of the Amsden Formation at Targhee Pass on the Montana-Idaho border (fig. 12, col. 10). *Pugnoides quinqueplecis* in several collections assigned to the *A. welleri-shawi* zone in Wyoming (Sando and others, 1975) has been found only in Pennsylvanian strata in central Montana (Maughan and Roberts, 1967, pl. 4). Thus, the *A. welleri-shawi* zone seems to be Early to Middle Pennsylvanian rather than Late Mississippian.

The age of the Horseshoe Shale Member at Berry Creek in the northern part of the Teton Range is considered to be Pennsylvanian, for the strata illustrate the same regional stratigraphic relations as at Targhee Pass and at Livingston Ranch. The Horseshoe at Berry Creek also includes the *A. welleri-shawi* zone in association with foraminiferal zone 19 of Mamet (Sando and others, 1975, p. A42). Zone 19, which correlates with the European *Homoceras* zone (Mamet, 1975, p. B6), had been thought to indicate a Late Mississippian age, even though this zone had not been identified in youngest beds of the type Chester in Illinois. The occurrence of foraminiferal zone 19 at Berry Creek probably confirms the Pennsylvanian age of these rocks because Lane (1977, p. 179) recently has shown that the *Rachistognathus primus* conodont zone occurs both in the base of the type Morrowan in Arkansas and in the *Homoceras* zone.

Assignment of a Mississippian age to the Horseshoe at Livingston Ranch (Sando and others, 1975), on the basis of foraminiferal data, is also contradictory to the regional correlations. All the taxa in the foraminiferal fauna listed from Livingston Ranch (Mamet, 1975, p. B9) also occur in collections of more diverse foraminiferal assemblages from the Moffat Trail Limestone of Chesterian age and from the Ranchester Limestone Member of early Atokan age. The Foraminifera designated as indicative of foraminiferal zone 18 at Livingston Ranch on the basis of the published list of taxa are not necessarily limited to the Late Mississippian (B. L. Skipp, oral. commun., 1978; P. L. Branckle, oral commun., 1978) and are not contradictory of a Pennsylvanian age.

TENSLEEP SANDSTONE

The Tensleep Sandstone was named by Darton (1904, p. 396–397) for sandstone above the Amsden Formation and below red beds that are now named Goose Egg Formation of Permian and Triassic age. The type section is in the canyon of Tensleep Creek on the western flank of the Bighorn Mountains. The formation comprises light-gray to yellowish-gray sandstone, medium-to coarsely crystalline pinkish to gray dolomite and dolomitic sandstone in mostly thick to massive beds. Dolomitic strata are mostly in the lower part of the formation, and in this report these beds are considered to be a dolomitic member of the Tensleep, although they have commonly been regarded as part of the Amsden Formation.

The dolomitic member comprises interbedded yellowish- to pinkish-gray and very light gray, fine- to medium-crystalline dolomite, green and red mudstone, and white to yellowish-gray, fine- and medium-grained quartzose sandstone. Dolomite generally predominates in the unit, but at some places, either sandstone or mudstone constitute the bulk of the rocks. Sandstone beds are increasingly abundant in upper parts of the unit. Greenish-gray to pale-green mudstone in beds as much as 0.5 m thick form a seemingly characteristic part of the lowermost few meters of this unit everywhere in Wyoming, although moderate-red to moderate-reddish-orange and pale-red-purple mudstone commonly occur at the base of the unit and may locally form basal red beds as much as 5 m thick. The dolomite unit reaches a maximum thickness of about 40 m, and pinkish, medium- to coarsely crystalline dolomite beds 2–4 m thick occur throughout the overlying Tensleep Sandstone.

Sandstone of the Tensleep comprises mostly fine- and medium-grained, well-sorted quartz sand. Both silica and carbonate minerals cement the sandstone; in a general way, siliceous cement dominates in the northwestern part of the State and carbonate cement dominates toward the southeast. Dolomite beds as much as 4 m thick occur in the upper part of the Tensleep, and at some places, these beds are at or near the top of the formation. Sandstone and dolomite beds in the lower part of the Tensleep range mostly from 2 to 5 m in thickness, and the sandstone commonly has small-scale, low-angle planar cross stratification. Beds reach a maximum thickness of 10 m in the upper part, and the sandstone includes a variety of crossbedding forms, including large-scale, high-angle, wedge-planar cross stratification indicative of offshore bar, beach, and aeolian dune deposits (Reynolds and others, 1975). Contorted crossbedding is commonly seen in the upper part of the Tensleep. In the Bighorn basin, transport of the sand was from the northwest (Todd, 1966), and this, as well as the regional geometry of the Tensleep and equivalent sandstone formations, suggests that the source of the sand was principally northwest of Wyoming, possibly from geanticlinal elements within the mobile belt of northern Idaho and adjacent areas (Maughan, 1975, p. 287–288).

Fusulinids of the zone of *Fusulina-Wedekindellina* indicate the Des Moinesian age of the Tensleep (Henbest, 1954, 1956; Agatston, 1954, p. 528), and Love (1954) has shown regional relations of the strata and fusulinid and other faunas from the Tensleep and associated strata. In a few places, an advanced form of *Profusulinella* in the lower part of the Tensleep suggests late Atokan age for the base of the formation (Henbest, 1956, p. 61, collections f–9791 and f–12103).

Rocks as young as late Des Moinesian may be included in the Tensleep Sandstone in western Wyoming (Love, 1954, col. 5), but those strata in far western Wyoming that have yielded fusulinids of Late Pennsylvanian age (Love, 1954, col. 3; Wanless and others, 1955, col. 19) are in the thrust belt, where the Des Moinesian and younger Pennsylvanian rocks are part of the Wells Formation (Schroeder, 1976; Jobin, 1972).

In the southeastern Bighorn Mountains, Pennsylvanian strata, possibly younger than Des Moinesian age (Mallory, 1967, p. G23), and Wolfcampian-age rocks are included in the Tensleep Sandstone near Mayoworth (Verville, 1957). There is no paleontological evidence in this area for either Missourian- or Virgilian-age rocks in the Tensleep, although a thicker section in this area suggests the possibility of Upper Pennsylvanian strata. Agatston (1954, p. 526) noted the increase in the number of limestone beds in the lower part of the Tensleep on the southeast flank of the Bighorn Mountains. Increase of limestone in this area can be attributed to gradation from the mostly sandstone facies of the Tensleep into the interbedded carbonate rock and sandstone facies of the middle member of the Minnelusa.

The Wolfcampian-age strata on the southeastern flank of the Bighorn Mountains (Verville, 1957) probably should not be included as part of the Tensleep Sandstone because these rocks unconformably overlie thick to massive crossbedded, slightly calcareous sandstone typical of the Tensleep, and the sandy strata above are of differing character. The Wolfcampian strata comprise thin planar beds of moderately to very dolomitic sandstone and sandy dolomite. They seem to be equivalent strata to the basal member (the "Nowood Member") of the Park City Formation (Maughan, 1967, p. 140) and probably could be identified as a tongue of the upper member of the Minnelusa. They should be separated from the Tensleep rather than incorporated as a local exception to the regionally established Des Moinesian age of the Tensleep Sandstone.

Pennsylvanian rocks of the upper sequence are overlain unconformably everywhere in Wyoming by rocks of Permian age.

ECONOMIC PRODUCTS FROM CARBONIFEROUS ROCKS

By David R. Lageson

The principal economic products from Carboniferous rocks in Wyoming are petroleum and high-calcium industrial limestone. Other commodities of minor importance include building stone and aggregate. Ground water from Mississippian limestones may have future economic importance in eastern Wyoming counties because of increasing demands for water in agriculture and in growing communities.

According to statistics of the Wyoming Oil & Gas Conservation Commission, oil and gas production from Mississippian reservoirs accounts for approximately 6 and 1 percent of the total oil and gas production in the State, respectively. Similarly, oil and gas production from Pennsylvanian reservoirs accounts for approximately 37 and 2 percent of the total State oil and gas production, respectively. Carboniferous strata have been, and will continue to be, important exploration targets for the petroleum industry in Wyoming. (See table 2.)

The exploration and development history of Carboniferous reservoirs in Wyoming has paralleled petroleum development in the Rocky Mountain region as a whole. Oil seeps and prominent "sheepherder" anticlines were the primary exploration targets during the late 1800's and early 1900's. Subsequently, as well control increased and as seismic technology became available, stratigraphic traps and seismic anomalies became exploration targets. More recently, deep drilling along basin axes and sophisticated stratigraphic prospects in wildcat areas are producing new discoveries. In addition, there may be considerable potential for petroleum in Carboniferous strata in the overthrust belt of western Wyoming, an area currently receiving a great deal of exploration interest.

Table 2.—*Oil and gas production from Carboniferous reservoirs in Wyoming*

[Data from the Wyoming Oil & Gas Conservation Commission. These data represent 1975 production and were the most current available at the time this report was compiled. bbl, billion barrels; mfc, million cubic feet]

Formation	Calendar year oil production (bbls)	Year total-oil (percent)	Calendar year gas production (mcf $\times$ 1,000)	Year total-gas (percent)
Madison	8,756,540	6.44	3,119,545	0.97
Amsden	147,578	.11	92,184	.03
Tensleep	34,000,464	25.01	5,608,017	1.75
Minnelusa	15,414,228	11.34	301,309	.09
Morgan	47,506	.04	285,287	.09
Weber	338,576	.25	1,249,982	.39
Total	58,704,892	43.19	10,656,324	3.32
Total of all producing formations in Wyoming)	135,809,697	99.9	317,646,049	99.3

The Powder River basin has produced significant quantities of petroleum from Pennsylvanian reservoirs. The Minnelusa Formation received intense exploration interest throughout the 1960's and, as a result, is currently producing oil from approximately 170 wells in the basin (Wyoming Oil & Gas Conservation Commission, 1975). Minnelusa traps typically result from either structural closure or truncation of reservoir beds over buried paleotopographic highs.

Mississippian limestones, although thus far unproductive in the Powder River basin, have excellent reservoir characteristics. Potential Mississippian reservoirs have not been adequately tested because of excessive depths (McGregor, 1972). However, many potentially productive anticlines along the basin margins have been flushed with fresh ground water, precluding the possibility of Mississippian production.

In the Bighorn basin, the Pennsylvanian Tensleep Sandstone is the principal oil producer. Anticlinal traps account for most of the Tensleep production in the basin; combination structural stratigraphic traps involve thinning or truncation of Tensleep reservoir beds at the post-Tensleep disconformity over large anticlines (Welden, 1972). Diagenetic modifications of the Tensleep Sandstone, including multiple-stage cementation, are very important in determining reservoir quality throughout the basin.

The Mississippian Madison Limestone, an excellent petroleum reservoir, is productive from several anticlines in the Bighorn basin. Drilling experience has shown that production from the Madison occurs only where the structural closure is greater than the thickness of the Phosphoria (Permian) to Madison interval (Weldon, 1972). Shale members of the Phosphoria Formation, which are rich in organic matter, appear to be the source rock for Paleozoic oil in the Bighorn basin.

In the Wind River basin, the Tensleep Sandstone is again one of the most productive reservoirs. Large asymmetric anticlinal traps along the west flank of the basin, where the Tensleep is the deepest productive horizon, produce most of the Tensleep oil (Benner, 1972).

Petroleum production from Carboniferous reservoirs across southern Wyoming has progressed sporadically. Tensleep production at Lost Soldier field was established in the 1920's. Mississippian oil was discovered at Lost Soldier in 1948. Gas production at Lost Soldier is now used to repressure the Tensleep reservoir after removal of condensate (Skeeters and Hale, 1972). Subsequent Tensleep discoveries include Hatfield, Baily Dome, and Sugar Creek, the last of which extended Pennsylvanian production south of Lost Soldier and west of Hatfield. Other Pennsylvanian reservoirs in southern Wyoming include Allen Lake, Medicine Bow, and Quealy Dome in the Hanna and Laramie basins. In the Green River basin of southwestern Wyoming, Paleozoic production is at prohibitive depths except on major structures (Skeeter and Hale, 1972). The Table Rock deep unit, a structure developed off the east flank of the Rock Springs uplift, is a significant gas discovery in the Madison and Weber Formations. Pennsylvanian sour gas was discovered in 1969 at Church Buttes field at a depth of 5,460 m. Mississippian strata have good reservoir qualities but have thus far produced only nonflammable gas.

High-calcium industrial limestone is another important commodity from Wyoming's Carboniferous rocks. The Mississippian Madison Limestone is the best potential source of high-calcium limestone in the State. High-calcium limestone is in demand in Wyoming for its use in refining sugar from sugar beets and in lime scrubber systems for air-pollution control of coal-fired electric-generating powerplants. In air-pollution control, approximately 100,000 tons of limestone would have to be mined each year to supply a 1000-MW generating plant, depending on variables such as heat value of the coal and purity of the limestone. As the use of Wyoming's coal increases, high-calcium-limestone production may also be expected to increase.

Pisolitic hematite from the Darwin Sandstone may have future economic value. Biggs (1951) has reported discontinuous lenses of pisolitic hematite 6–10.5 m above the base of the Darwin Sandstone along the northeast flank of the Wind River Mountains.

Ground water from the Madison Limestone in the Powder River basin is a questionable and disputed resource. A proposed project involving the removal of Madison water for use in a coal-slurry pipeline is questioned on the basis of the recharge ability of the Madison aquifer. Madison water might also have future significance in terms of the increasing water demands of agriculture and communities in eastern Wyoming. Many questions still remain to be answered before this resource becomes a reality.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— Colorado

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Colorado

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—COLORADO

By John Chronic[1]

ABSTRACT

Carboniferous rocks of Colorado have two contrasting facies, which roughly correspond to the Mississippian and the Pennsylvanian.

The Mississippian was a time of widespread and uniform carbonate deposition, which originally covered almost all the State; most carbonate beds are referred to as the Leadville Limestone. Later, most of Colorado, especially the central part, was domed above the sea, a karst surface and residual soil formed, and, in some places, Mississippian strata were completely removed by erosion.

Slight subsidence at about the beginning of Pennsylvanian time brought marginal coal-swamp deposition and then shallow-marine conditions in the basins. These basins bordered two elongate ranges of the rising ancestral Rockies, one roughly parallel to but west of the present Front Range, and the other extending northwest from what is now the San Luis Valley to beyond the present Uncompahgre Plateau.

Until about the middle of Pennsylvanian time, these mountains rose intermittently, shedding coarse alluvium into the adjacent basins in a series of complex marine and nonmarine cycles; deposits reached a maximum thickness of much more than 3,000 m (10,000 ft). In central and western Colorado, 29 evaporite cycles, including several deposits of potash, are present along the Utah-Colorado border. A multitude of names are used for the complex Pennsylvanian sequence.

The later part of the Pennsylvanian is poorly dated in Colorado because the strata lack fossils and datable igneous rocks; therefore, the upper boundary is poorly defined. Other evidence suggests that late in the period the two uplifts and three basins slowed in their differential movement and that the entire central part of the State was broadly domed and raised above sea level; nonmarine red beds, composed largely of debris from the ancestral Rockies, were deposited.

Carboniferous rocks of Colorado yield significant amounts of oil and gas; uranium has been precipitated in economic quantities in the carbonaceous shale. Many of the richest mines for gold, silver, lead, and zinc in Colorado are in Carboniferous rocks enriched during Laramide intrusive and tectonic mountain-building activity. Building stone and evaporitic minerals are produced from Carboniferous rocks as well, but no economic coal of that age is present.

INTRODUCTION

Carboniferous rocks in Colorado were first reported on the maps of Jules Marcou and Edward Hitchcock in 1853, but not until 1870 were they documented along the Front Range by F. V. Hayden. By about 1900, geologists knew that the Carboniferous of Colorado represented two very different regimes, the earlier one, Mississippian, consisting of a very widely distributed but thin limestone sequence deposited in shallow open-marine conditions without reefs or reefing, and the later one, Pennsylvanian, a highly variable and tectonically complex thick sequence of cyclic marine and continental sedimentary rocks.

Now, after 125 years of study, we know that, after deposition, Carboniferous rocks blanketed the State. Since then, however, they have been extensively eroded from the mountains and have been almost entirely covered by later deposition on the plains and plateaus; their present outcrops are "more or less discontinuous areas exposed along the flanks of mountain ranges or sometimes in the channels of rivers" (Girty, 1903, p. 27).

To paraphrase Girty, the most conspicuous feature in the distribution of Carboniferous rocks outcropping in Colorado is their occurrence in a discontinuous band stretching from the northwestern corner of the State to the center of its southern border. The northwestern part of the band is at the eastern end of the Uinta Mountains, where thick fossiliferous strata of Carboniferous age are present.

Southeast of the Uinta Mountains is an isolated fault block, Juniper Mountain, of extremely fossiliferous Pennsylvanian strata. Southeast of this block is a large area of Carboniferous outcrops having a very irregular outline, exposed chiefly along the White and Colorado Rivers and their tributaries and around the White River Plateau. Continuous with this to the south, and really forming a part of it, is the area produced by the upturned beds of the Sawatch and Elk Mountains, including Maroon Bells southwest of Aspen.

East and southeast of these areas, east of the Sawatch Range, a narrow linear area of outcrop ex-

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tends from near Vail south through Leadville to the Arkansas River valley and along the west side of South Park. Continuous with this to the south, but widening in exposure, is the Sangre de Cristo Range, extending to the southern border of the State and composed mostly of Carboniferous and Permian strata.

Another well-defined area of Carboniferous exposure is in southwestern Colorado, in the San Juan Mountains, and northwestward along the Dolores River to Gypsum Valley on the Utah line.

A last outcrop is the strip of sedimentary rocks upturned along the eastern margin of the Front Range. Carboniferous rocks are discontinuous along this belt; they extend from the Wyoming line southward to near the Huerfano River at the southern end of the Wet Mountains.

All the outcrops of Carboniferous rocks in Colorado were exposed by Cenozoic mountain building, and many thousands of square miles of Carboniferous strata have been eroded from the State in the last 60 million years. Many miles of truly outstanding Carboniferous outcrops are in the Colorado mountains, and notable caverns, some well above timber line, have been cut into the massive limestones of this age.

In most places, Carboniferous rocks are exposed at altitudes of 1820 m (6,000 ft) to more than 4,250 m (14,000 ft). Steep drainage basins are normal in these areas, and outcrops are well exposed in canyons and cliffs. Probably the best exposures are in central Colorado, from the McCoy area on the Colorado River southeast along the Mosquito Range and the Sangre de Cristo Range. An outstanding and very accessible exposure of Pennsylvanian strata also is present at Molas Pass south of Silverton. Most of the red arkose and coarse sandstone along the Front Range, from Garden of the Gods northward to the Flatirons at Boulder, are probably of Carboniferous age, although they are generally unfossiliferous.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Colorado Geological Survey.

HISTORY

U.S. Government geologists have made many of the studies of Carboniferous rocks of Colorado. In addition to and overlapping the extensive activities of the Hayden survey (Hayden, 1870), were investigations of the Wheeler, King, and Powell surveys, all contributing information on the strata, principally in the central and western parts of the State. The U.S. Geological Survey, established in 1879, began studies in Colorado, especially in and around Leadville. One of the first and most significant reports was by S. F. Emmons in the U.S. Geological Survey Second Annual Report published in 1882. Emmons' final report on this area, in 1886, was U.S. Geological Survey Monograph 12. Then followed several folios and monographs that included details of local occurrence of the Carboniferous System without containing anything generally significant.

In 1903, George H. Girty published "Carboniferous Formations and Faunas of Colorado," U.S. Geological Survey Professional Paper 16. All known occurrences and all fossils collected by the earlier surveys were described and summarized, and an excellent chronologic bibliography was included. This work is the most important reference on the subject of the present report, even though many of the names of formations and fossils have been changed by subsequent research.

Since Girty's classic work, many papers have been published on Carboniferous strata of Colorado, only the most significant of which can be mentioned here. Most of the authors in the early 1900's still used the term Carboniferous, but by 1930, most Colorado Carboniferous rocks were called either Mississippian or Pennsylvanian, and, as in the rest of the United States, the name Carboniferous fell into disuse.

In 1930, Roth and Skinner published the first descriptions of Colorado Pennsylvanian fusulinids and microfossils from the McCoy area, calling the containing rocks the McCoy Formation. In 1934, Read and Johnson described rocks and fossils of Pennsylvanian age from the Mosquito Range, and Brainerd and Johnson summarized the Mississippian of Colorado. Vanderwilt (1935) described the stratigraphy of Pennsylvanian strata in the Elk Mountains, and Johnson (1940, 1944, 1945) continued his long series of publications on the Paleozoic of central Colorado and the algae of the Carboniferous.

Arnold, in 1940 and 1941, called attention to the early Pennsylvanian floras of central Colorado, and in 1945, Thompson described the fusulinids and strata of northwestern Colorado. Brill (1944, 1952) studied the Pennsylvanian and Permian of central Colorado and named the two principal Pennsylvanian stratigraphic units the Belden and Minturn Formations.

McLaughlin (1952) and Lehman (1953) published descriptions of microfossils from the Glen Eyrie Formation near Manitou. Wengerd and Strickland, in 1954, described the varied stratigraphy of the Pennsylvanian in the Paradox salt basin in southwestern Colorado and adjacent areas after oil was discovered there.

The Rocky Mountain Association of Geologists (RMAG) held a field conference and published an excellent "Symposium on Pennsylvanian Rocks of Colorado and Adjacent Areas" (Curtis, 1958). The 22 articles in this symposium represent a comprehensive summary of knowledge on the Pennsylvanian to the date of publication. In 1961, another RMAG field conference resulted in the "Symposium on Lower and Middle Paleozoic Rocks of Colorado" (Berg and Rold, 1961), in which the Mississippian strata of the State are dealt with in a reasonably complete series of articles.

In 1962, Stevens summarized and illustrated the brachiopods of the McCoy area, and, in 1965, Murray and Chronic published the results of a study on conodonts from the Minturn Formation in western Colorado.

Finally, in 1972, Mallory summarized the Pennsylvanian System (p. 111–127) and Craig (*in* Mallory, 1972, p. 100–110) summarized the Mississippian System in "Geologic Atlas of the Rocky Mountain Region." Much of the stratigraphic information and many of the maps presented here are from this masterful volume; the paleontologic data are from the RMAG Symposiums (Curtis, 1958, and Berg and Rold, 1961).

The Colorado Geological Survey has had two periods of active existence. During the earlier period, work was done at the University of Colorado, and publications were mostly summaries, bibliographies, and reports in bulletin form on the geology of various mining areas. In the first report, published in 1909, Junius Henderson summarized information on "The Foothills Formations Of North Central Colorado," revising and adding to this report in 1920. Other early Survey reports document the occurrence of Carboniferous rocks and fossils in mining districts of the State, but the rocks were not seriously studied.

After a hiatus of 35 years, a newly organized Colorado Geological Survey began work in 1969 and has published reports, maps, and other geologic information since. Most data presented to date have been of practical nature, emphasizing oil, coal, gas, minerals, and geologic hazards. No contributive general information on the Carboniferous has been published, and none is now being prepared, although data on economic products from Carboniferous rocks are being gathered and are available at the Survey offices.

GEOLOGIC SETTING

Carboniferous rocks of Colorado lie on strata ranging in age from Precambrian to Late Devonian (fig. 1). In the eastern part of the State, Devonian rocks are generally absent, and Precambrian, Cambrian, and Ordovician beds underlie the Carboniferous rocks; in the west, the Late Devonian Ouray or Chaffee Formations underlie the Carboniferous rocks.

In eastern Colorado, the lower contact is unconformable, but there is little or no angularity between the underlying beds and the Carboniferous and no marked evidence of erosion, even though much time passed between deposition of the two sets of strata. Indeed, data on fossils from diatremes in the northern Front Range, indicate that Silurian marine strata were deposited over at least that part of the State and then completely removed by erosion before Carboniferous deposition.

On the west, deposition appears to have been continuous during Devonian and Carboniferous time; the Ouray Formation of the San Juan Mountains contains fossils of both Devonian and Mississippian age. The exact contact between Devonian and Carboniferous is impossible to determine in many places because no angular relationship or weathered zone is present. In many places, a darker gray zone of carbonate, some of which is soft, is found at about the contact level, but in other places, the Upper Devonian Ouray or Chaffee Formations appear stratigraphically continuous with the massive Leadville Limestone above.

Rocks overlying the Carboniferous in Colorado are generally red beds of continental origin, except in the subsurface of southeastern and southwestern Colorado where, in a few wells, Early Permian limestones have been found that contain fusulinids, suggesting a marine origin (fig. 2). In most places, except very near the Front Range and Uncompahgre uplifts, red beds, whose age is known or strongly suspected to be Permian, are composed of sandstone or shale, indicative of the sharp weakening of tectonism in these areas toward the end of Carboniferous time. In the subsurface of southeastern Colorado, a large area known as the Apishapa uplift contains no known Carboniferous rocks. There, Precambrian

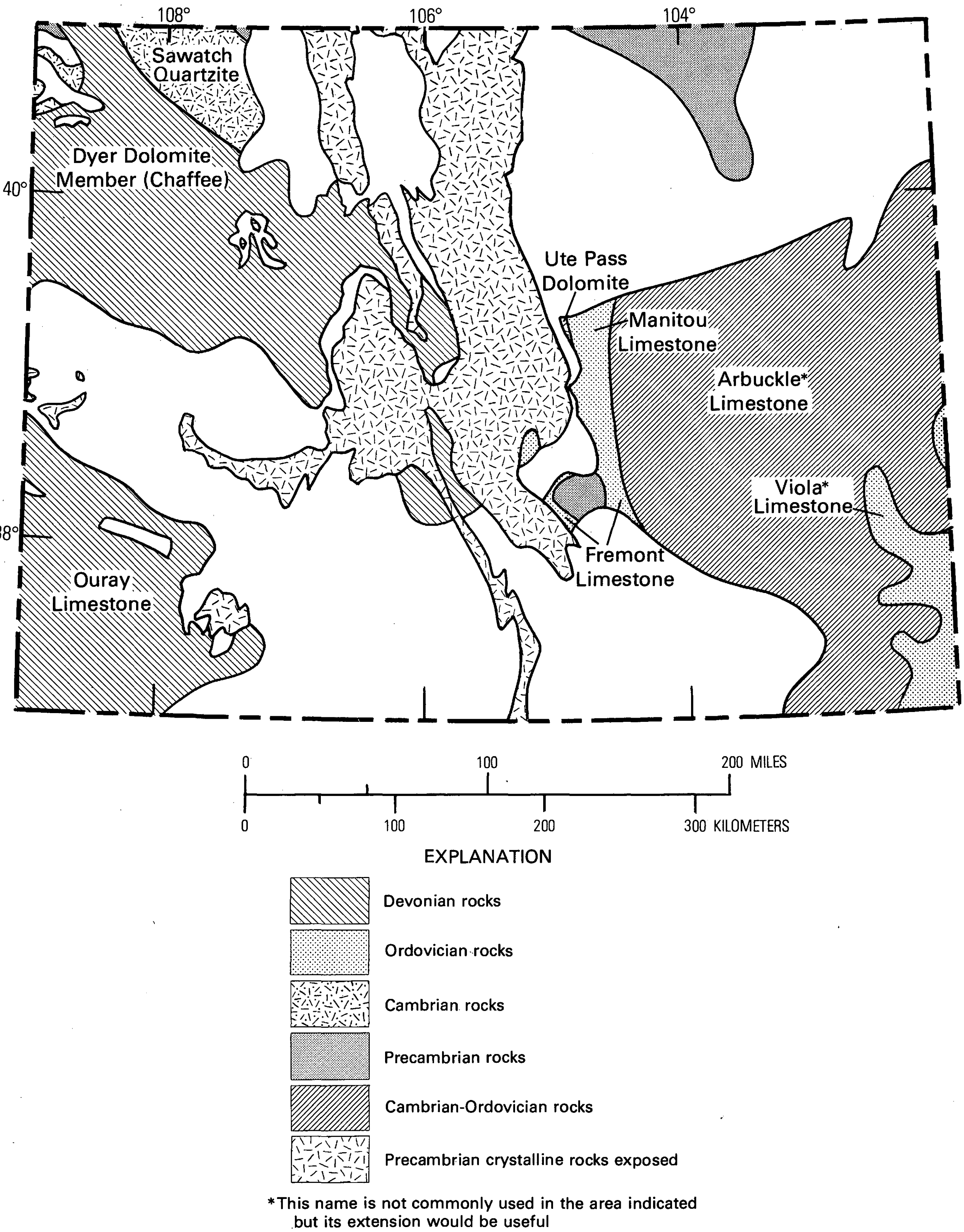


FIGURE 1.—Sub-Mississippian paleogeology (modified from Mallory, 1972, p. 102).

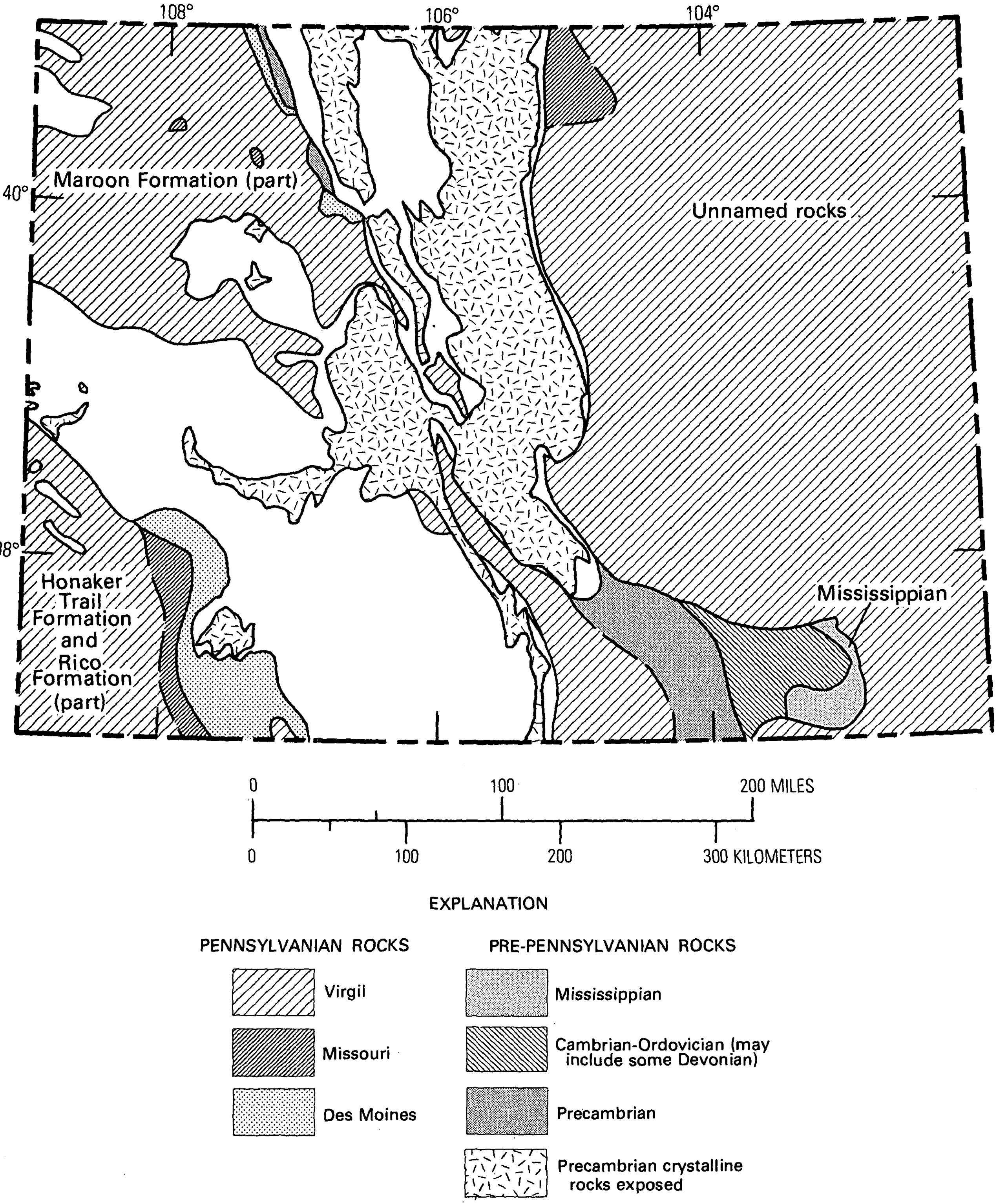


FIGURE 2.—Sub-Permian paleogeology (modified from Bailey Rascoe, Jr., and Donald L. Baars, *in* Mallory, 1972, p. 149).

and earlier Paleozoic strata are in contact with those of Permian age.

The upper surface of the Carboniferous in Colorado cannot be determined accurately because it is unfossiliferous and unsuited for radioactive age dating. Because most of the State was a lowland area receiving alluvial clastic sediments from the residual hills of the Front Range, Uncompahgre, and Apishapa uplifts, much of the upper surface will probably be found to be irregularly unconformable, when and if suitable means of age determination are established.

From the beginning of the Paleozoic Era until about the end of the Mississippian Period, Colorado was relatively stable and had almost no relief and little tectonic movement, except for regional uplift and subsidence. Mississippian strata are not well dated in Colorado, but the early and late parts of the period were apparently times of relative positive movement, whereas the middle part was a time of submergence.

Some evidence indicates that local graben and horst formation took place along the south side of the Uncompahgre uplift during middle Mississippian time, possibly the first signal of later violent Carboniferous events in Colorado.

Late in Mississippian time, Colorado slowly rose above the sea and became a lowland on which the recently deposited carbonate rocks were deeply eroded, almost entirely by solution. Many caves and a widespread karst surface formed at this time.

In Early Pennsylvanian time, the seas readvanced locally into basins between the Front Range and Uncompahgre uplifts, which were in an incipient stage of uplift.

As vertical uplift in these elongate regions, as well as in the Apishapa uplift, accelerated, and deep basins formed between them and at their sides, erosion moved great masses of sediment into the basins, resulting in the great quantity and variable character of Middle and Upper Pennsylvanian strata.

Orogenic movements must have ceased by about the end of Middle Pennsylvanian time, and the entire State was regionally uplifted above sea level. As Middle Pennsylvanian seas receded from the central part of the State, widespread continental deposition took place, and seas remained only in the southeastern and southwestern corners. Tectonic stability must have prevailed by the end of Carboniferous time, for the land was gradually flattened by erosion during the Permian and Triassic Periods.

The cyclic nature of sedimentation in much of the Pennsylvanian Period in Colorado reflects to a remarkable extent the general cyclicity of deposition throughout the world at this time. Few cycles in Colorado rocks contain coal, but most are highly variable units in which red beds, conglomerates, and arkoses alternate with marine limestone, which is usually gray or greenish gray. More than 60 cycles have been recorded in the probable Pennsylvanian rocks of the Arkansas River valley, and at least 40 in many other central Colorado sections. Most of these cycles whose ages have been determined are of Early and Middle Pennsylvanian age; later cycles are completely unfossiliferous.

For probably 100 million years after the Carboniferous Period, Colorado was extremely stable tectonically, but almost always somewhat above sea level. Then, in Early Cretaceous time, the sea invaded the State and covered it completely for about 40 million years.

Along with the surrounding Rocky Mountain area, the State was uplifted locally, and the present mountain ranges formed about 50 to 60 million years ago. Intervening areas, the present intermontane basins, remained slightly above sea level. Then, about 20 million years ago, in Miocene time, the whole State was regionally uplifted to about its present altitude. Violent volcanic activity took place, especially in the southwest, and finally, during the last few million years, glacial erosion and deposition gave the State its present topography. Carboniferous strata that formerly covered the entire State were almost completely eroded away from the mountainous area.

LITHOSTRATIGRAPHY

Carboniferous stratigraphy in Colorado is a study in contrasts. Almost all Mississippian rocks are called Leadville Limestone. Because of the many isolated outcrops and, in the Pennsylvanian System, the abrupt lateral and vertical facies changes, a number of stratigraphic units are in use for Colorado Carboniferous rocks (figs. 3 and 4).

On the east, in subsurface, Kansas terminology is generally used, but along the Front Range, the outcrop names Glen Eyrie and Fountain are used for the 150- to 1,250-m (500- to 4,000-ft)-thick Pennsylvanian section, a marine and continental cyclic sequence overlain by colorful arkosic red beds that form the Garden of the Gods near Colorado Springs, the Redrocks west of Denver, and the Flatirons at Boulder.

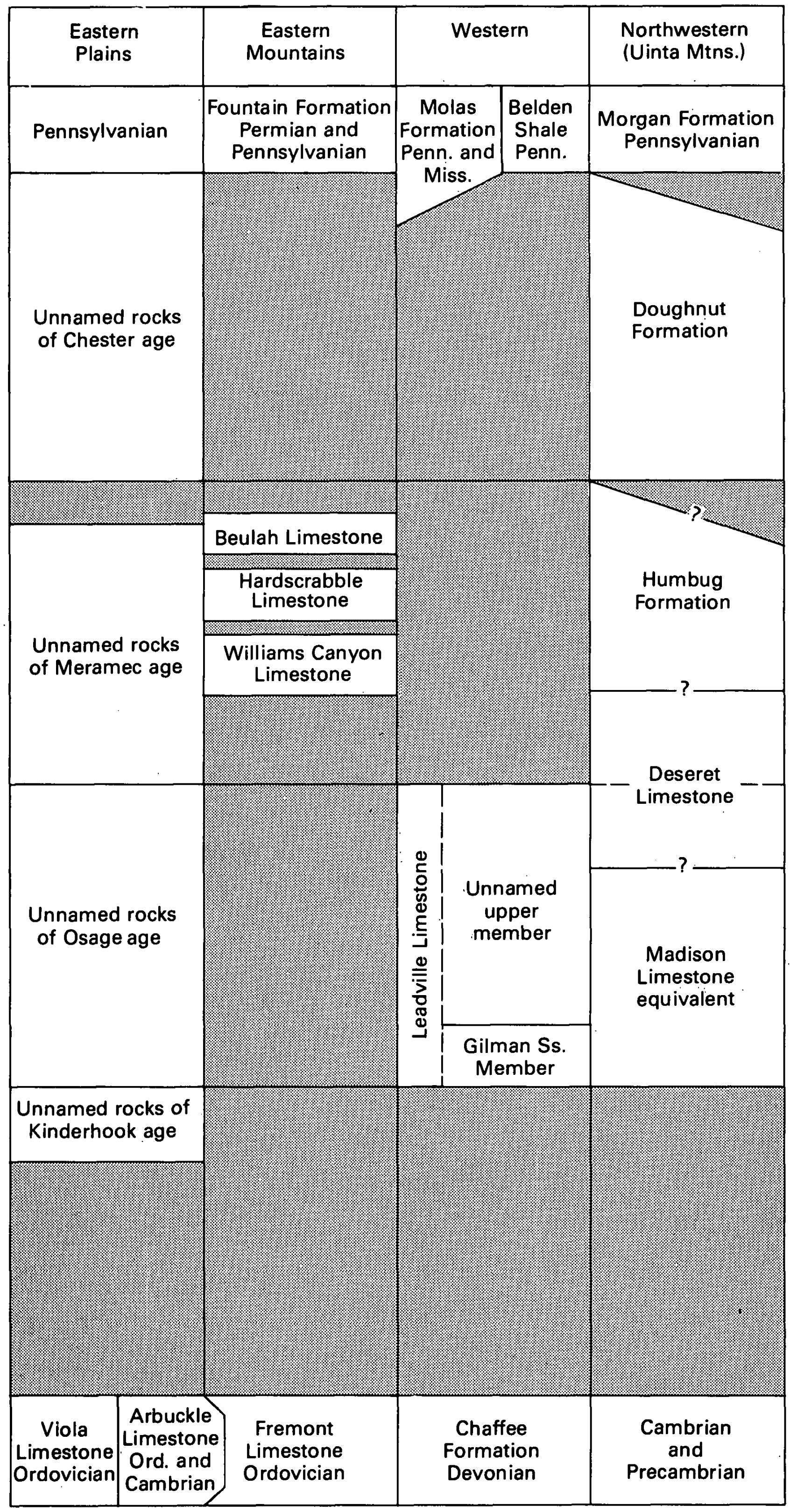


FIGURE 3.—Mississippian correlation chart (modified from L. W. Craig, *in* Mallory, 1972, p. 101).

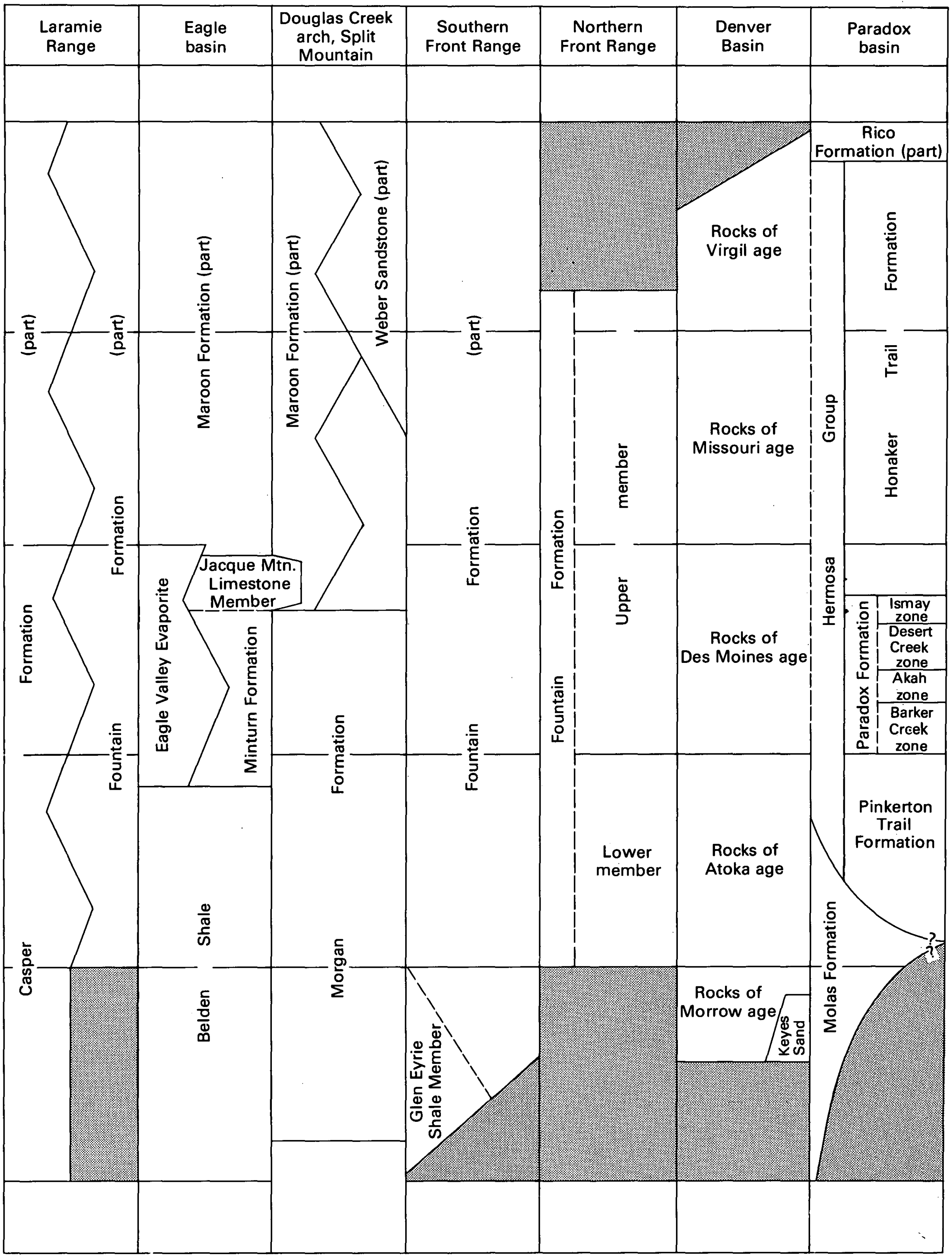


FIGURE 4.—Pennsylvanian correlation chart (modified Mallory, 1972, p. 113).

In the Maroon basin of central Colorado, the sequence of Kerber, Belden, Minturn, and Maroon Formations is used for Pennsylvanian strata. In the south-central part of the State, terminology of New Mexico is used, and in the northwest, that of Utah.

In southwest Colorado, other names have come into use because of the physical separation of the area from the Maroon basin. Molas, Pinkerton Trail, Paradox, Hermosa, and Honaker Trail are usually used for the sequence of Pennsylvanian strata in the Paradox basin, but other names for larger and smaller units are also used.

Figures 5–17, from the "Geologic Atlas of the Rocky Mountain Region" (Mallory, 1972), convey more clearly than any other way the nature of the Carboniferous rocks of Colorado.

In general, the contact between the Mississippian and Pennsylvanian in Colorado represents an erosion surface that has karst characteristics in many places (fig. 11). Commonly, there is an iron-rich, red soil zone, which contains chert in many places

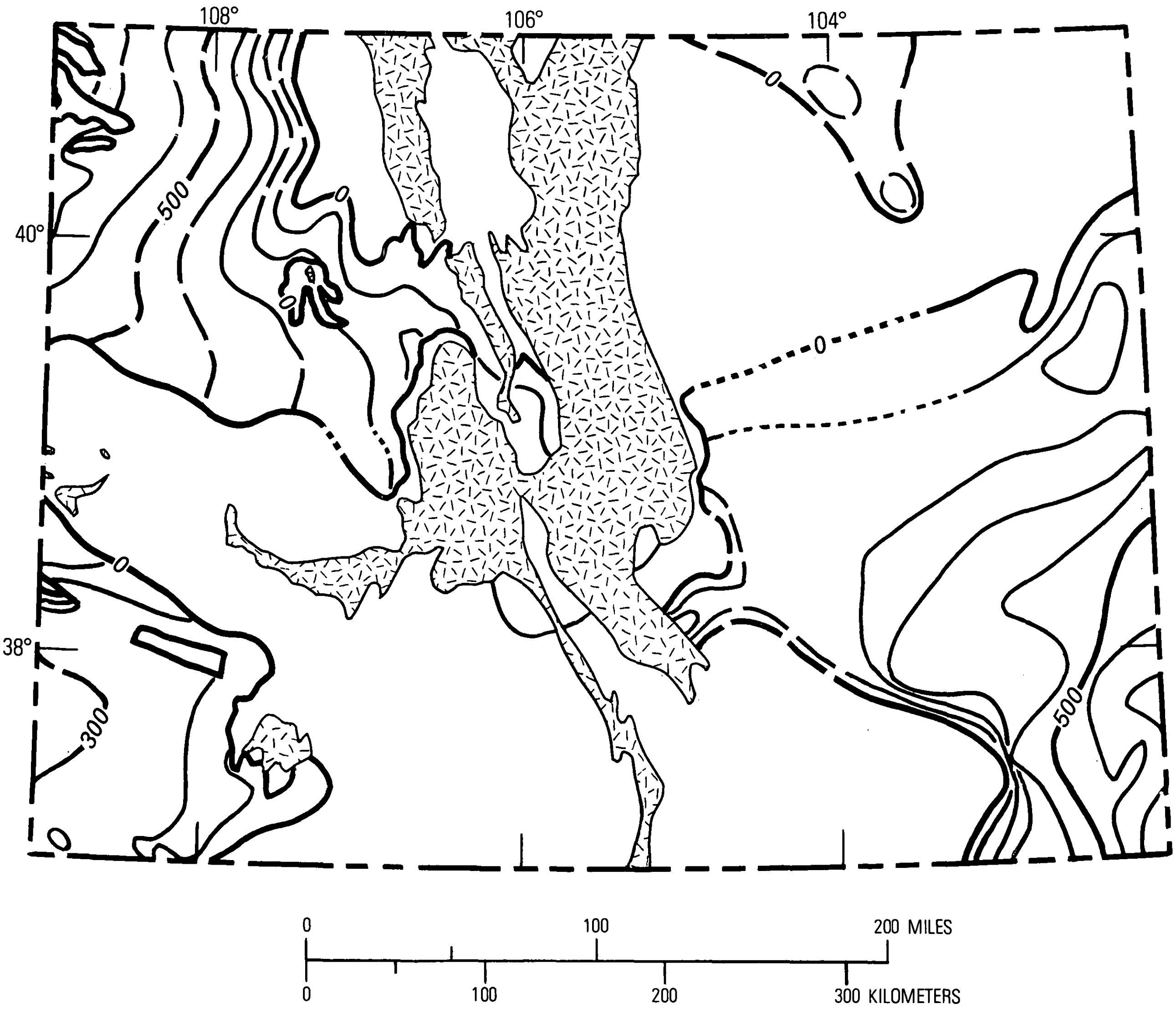


FIGURE 5.—Thickness of the Mississippian System (modified from L. W. Craig, *in* Mallory, 1972, p. 103). Isopach interval 100 feet, dashed where distant from control, dotted where conjectural. Precambrian crystalline rocks are patterned. Note the extensive areas in north-central, northeastern, southern and western Colorado where no Mississippian rock now exists; also note the thickening of the system to the southeast and west, toward the State margins, and the faulted graben on the western state line.

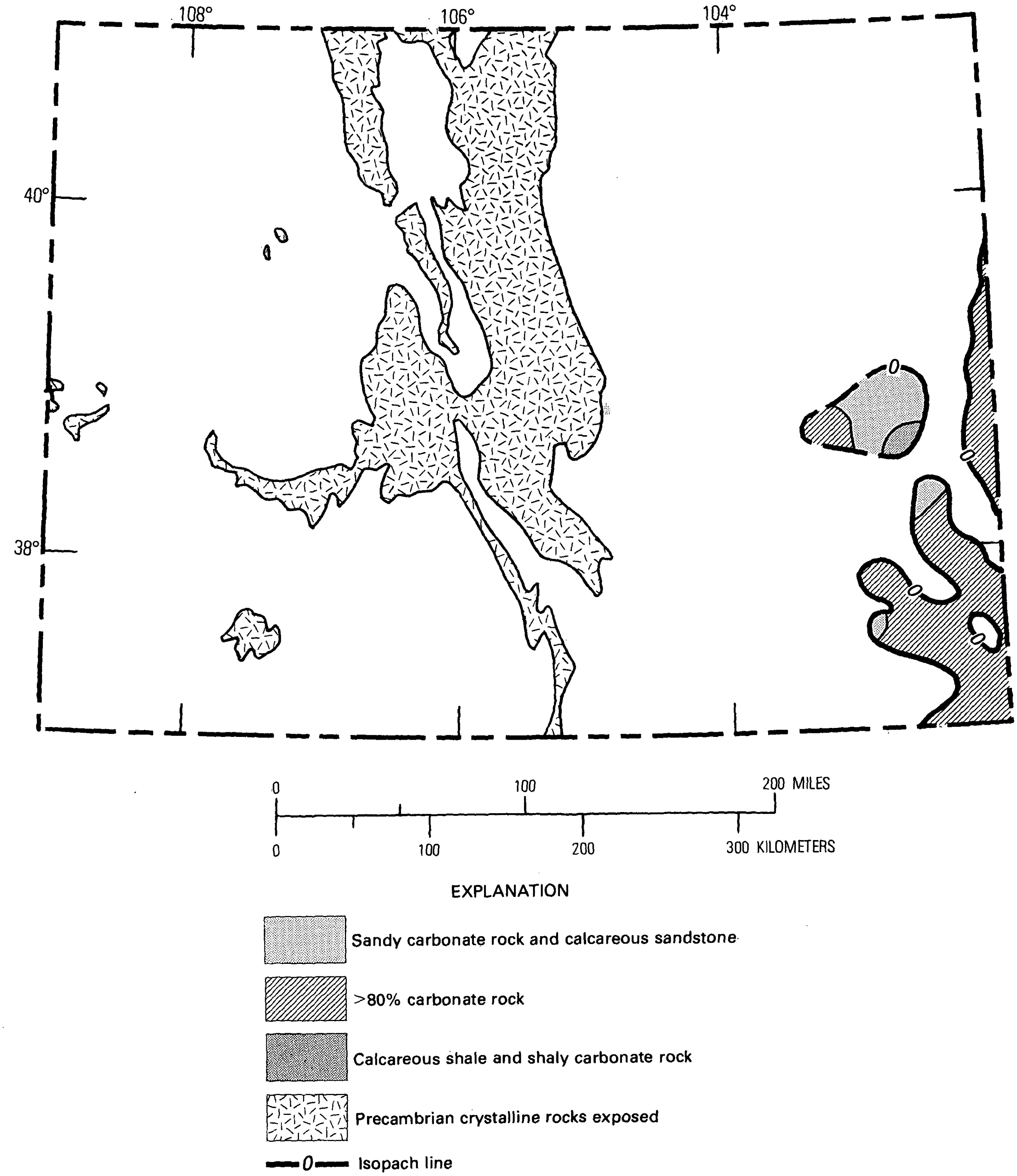


FIGURE 6.—Thickness and lithofacies of rocks of Kinderhook age (modified from L. W. Craig, *in* Mallory, 1972, p. 104).

and may be fossiliferous and of Mississippian age. This deposit is called the Molas Formation in central and western Colorado; in some places, the contact zone and the Molas may be enriched with ores because of its unusual physical and geochemical nature.

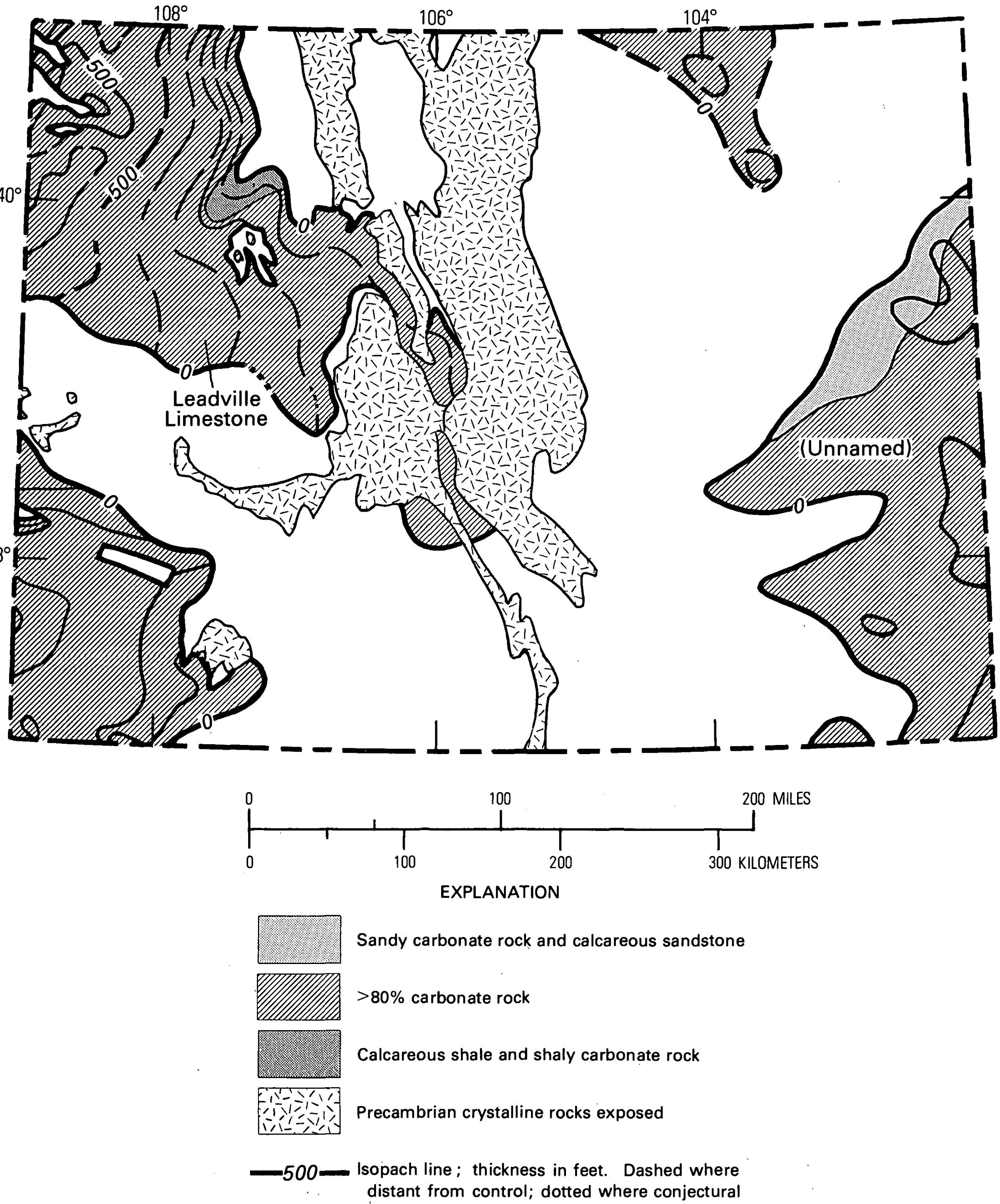


FIGURE 7.—Thickness and lithofacies of rocks of Osage age (modified from L. W. Craig, *in* Mallory, 1972, p. 105).

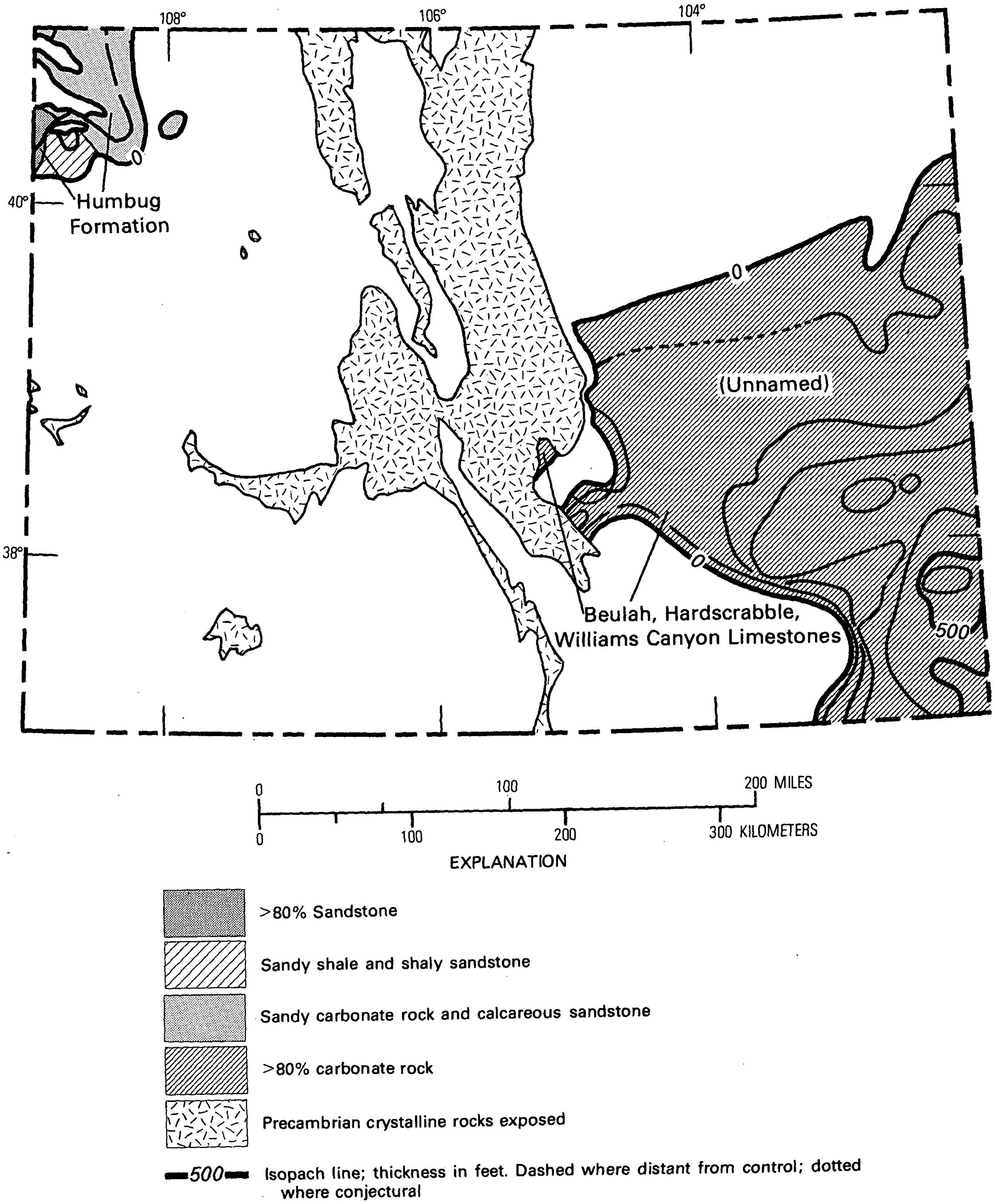


FIGURE 8.—Thickness and lithofacies of rocks of Meramec age (modified from L. W. Craig, *in* Mallory, 1972, p. 107).

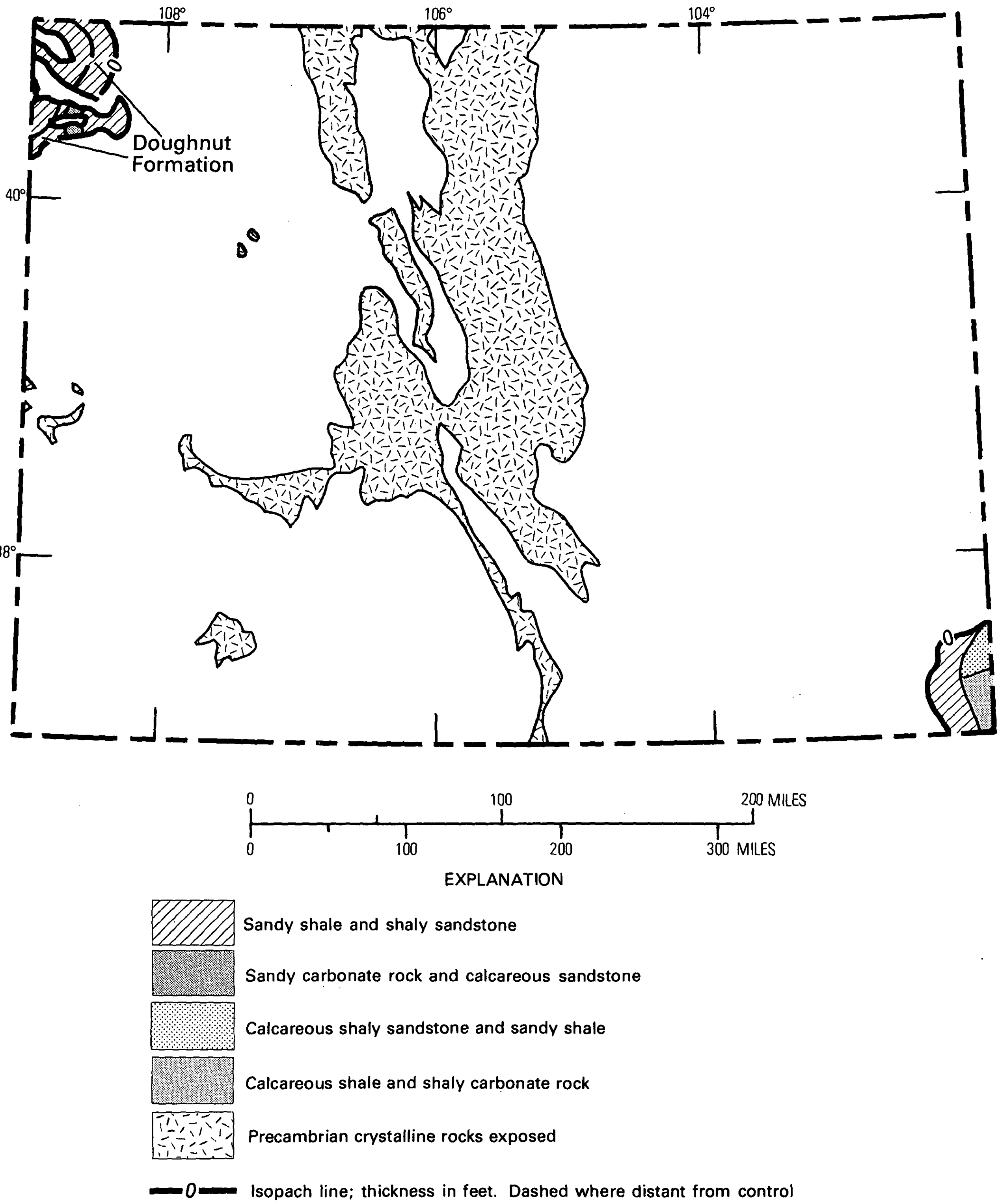


FIGURE 9.—Thickness and lithofacies of rocks of Chester age (modified from L. W. Craig, *in* Mallory, 1972, p. 108).

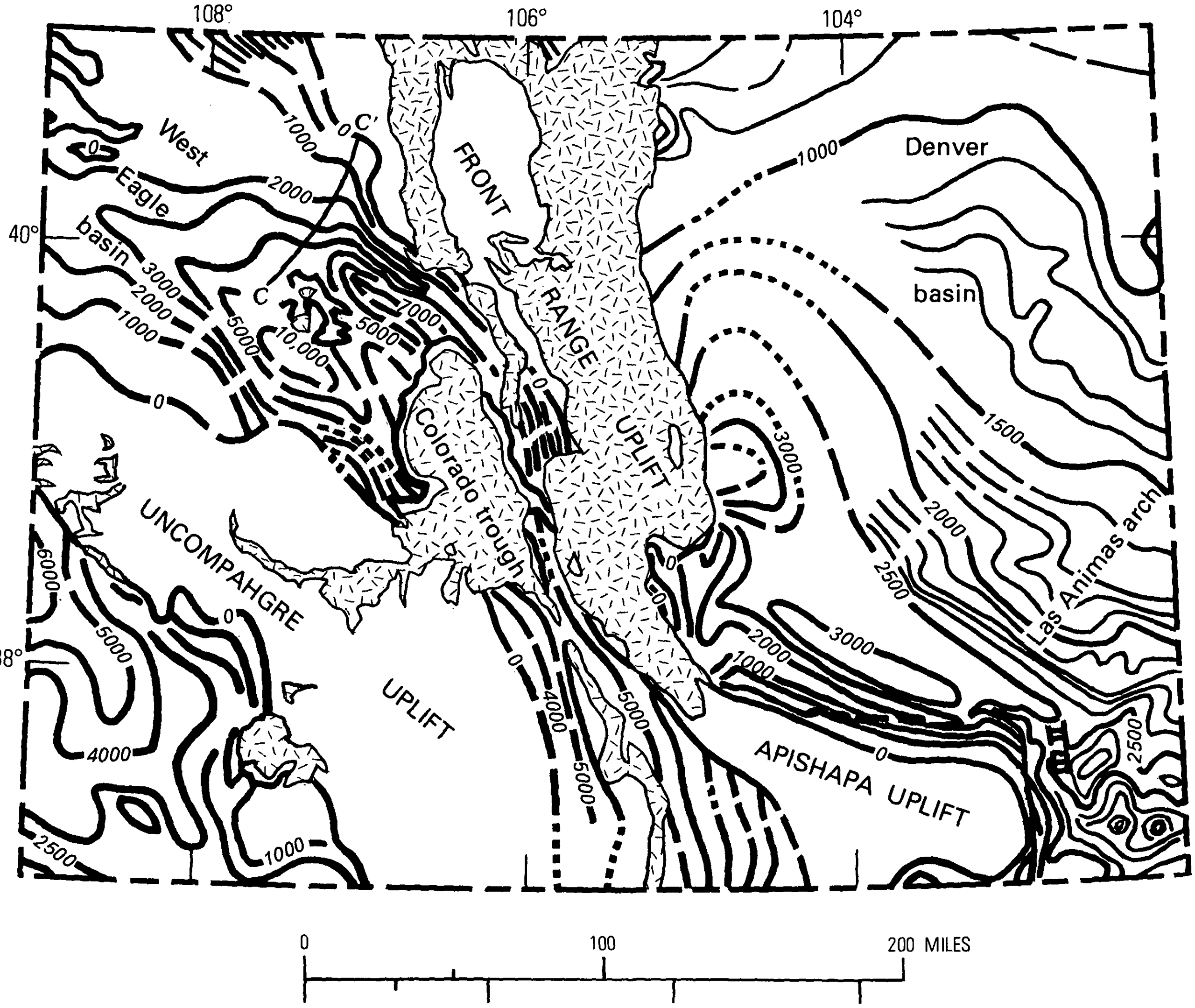


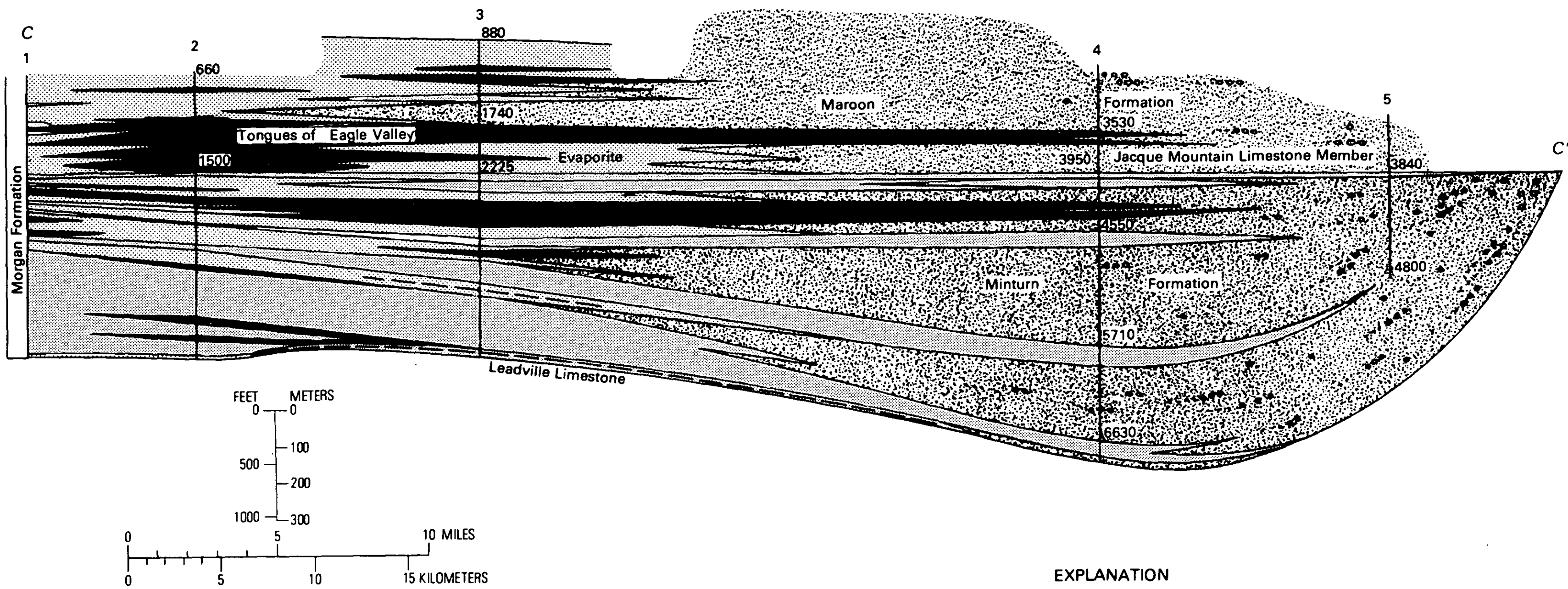
FIGURE 10.—Thickness of the Pennsylvanian System (modified from Mallory, 1972, p. 115). Isopach interval 100 feet, dashed where distant from control, dotted where conjectural. Precambrian crystalline rocks are shaded. Cross section *C–C′* in northwest, across central Colorado trough, is shown in figure 11. The Front Range, Uncompahgre, and Apishapa uplifts were above sea level and at times were mountains during the Pennsylvanian Period. Location of fig. 11 shown by line *C–C′*.

In some areas, especially along the northern Front Range, Mississippian rocks were undoubtedly present but were completely eroded away before Pennsylvanian clastic deposition, as some chert pebbles contain Mississippian fossils, especially the brachiopod *Eumertia* sp. and the bivalve *Conocardium* sp., in the basal Pennsylvanian red beds.

Every known class of sedimentary rock can be found in the Carboniferous strata of Colorado. The principal rocks exposed, however, are the massive gray limestone of the Mississippian and the coarse reddish-gray conglomeratic arkose of the Pennsylvanian. The Pennsylvanian, as indicated in figures 13 to 17, is extremely variable; the very coarse conglomerate of the Minturn and Sangre de Cristo Formations of south-central Colorado contrast strongly with the massive reef and offshore limestones of northwestern Colorado and the thick cyclic salt beds of the Paradox basin.

In Mississippian time, facies changes were scarce and slight and consisted mostly of minor textural changes in the almost pure carbonate. The lower beds of the Mississippian tend to be very fine grained; the overlying beds are commonly oolitic and contain endothyrid foraminifera.

* Fig. 11



CONTROL POINT IDENTIFICATION TABLE

Key no.	Name	Sec.	T.	R.
	Colorado			
1.	Miller Creek	7	2S	92W
2.	Miracle and Wooster, Quiett 1 and North Elk Cr	26	1S	92W
3.	Buford Oil, Govt. 1	16	1N	91W
4.	Phillips Petroleum, Poose Cr. 1	10	2N	88W
5.	Robinson, Kagie 1	29	4N	87W

EXPLANATION

DETRITAL ROCKS

Arkosic conglomerate, grit, sandstone, and shale, usually bright red, of the Maroon Formation.

Arkosic conglomerate, grit, sandstone, siltstone, shale, and intercalated carbonate beds, usually red, maroon, tan, or gray, of the Minturn Formation.

Sandstone with some siltstone and shale, white, gray, buff, pink, or orange, of the Morgan Formation.

PRECIPITATED ROCKS

Gypsum, anhydrite, and sulfatiferous siltstone and mudstone.

Limestone (and, locally, dolomite).

FIGURE 11.—Stratigraphic section through the Pennsylvanian System in the Eagle basin, northwest Colorado (modified from Mallory, 1972, p. 125). Location given on fig. 10.

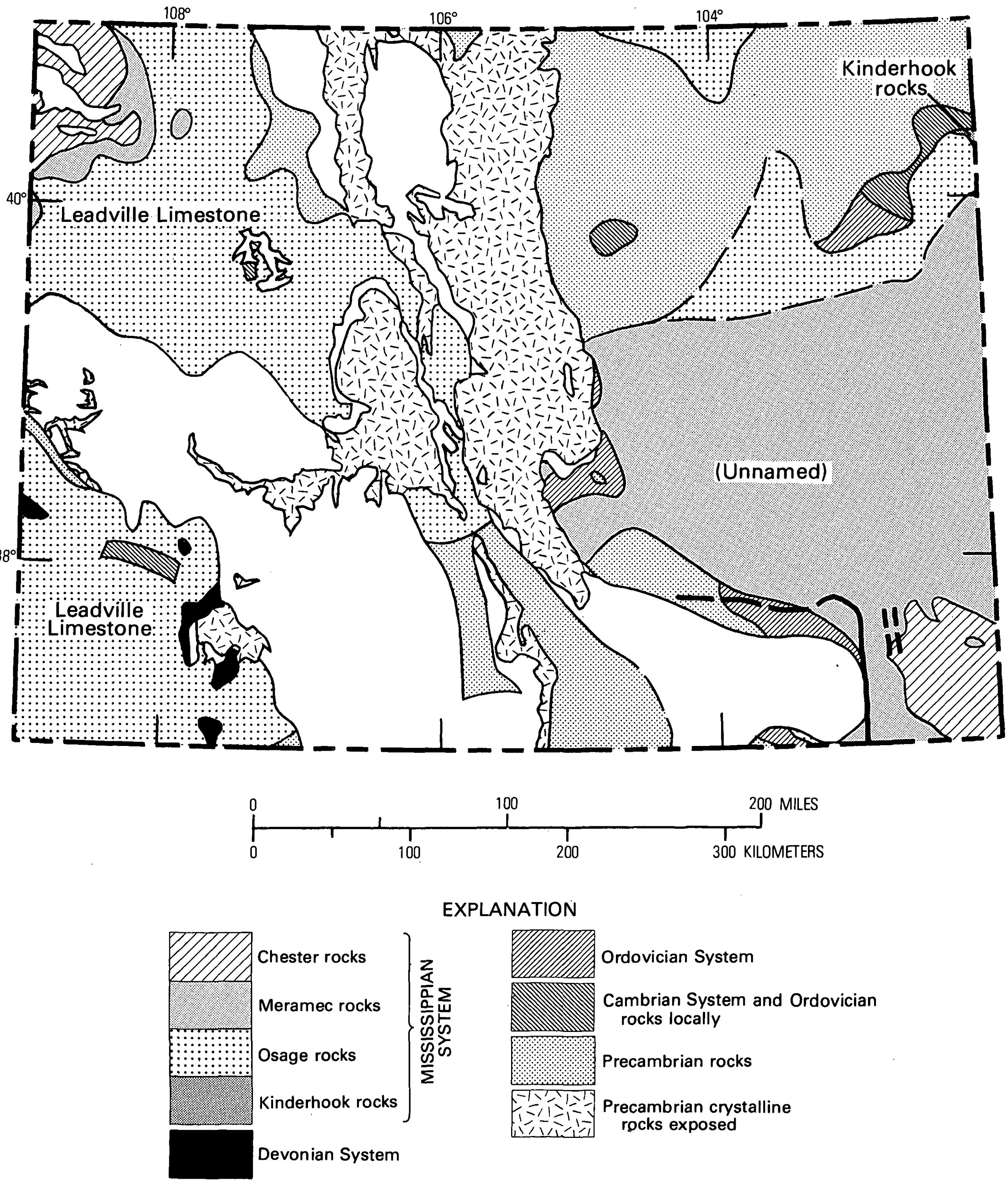
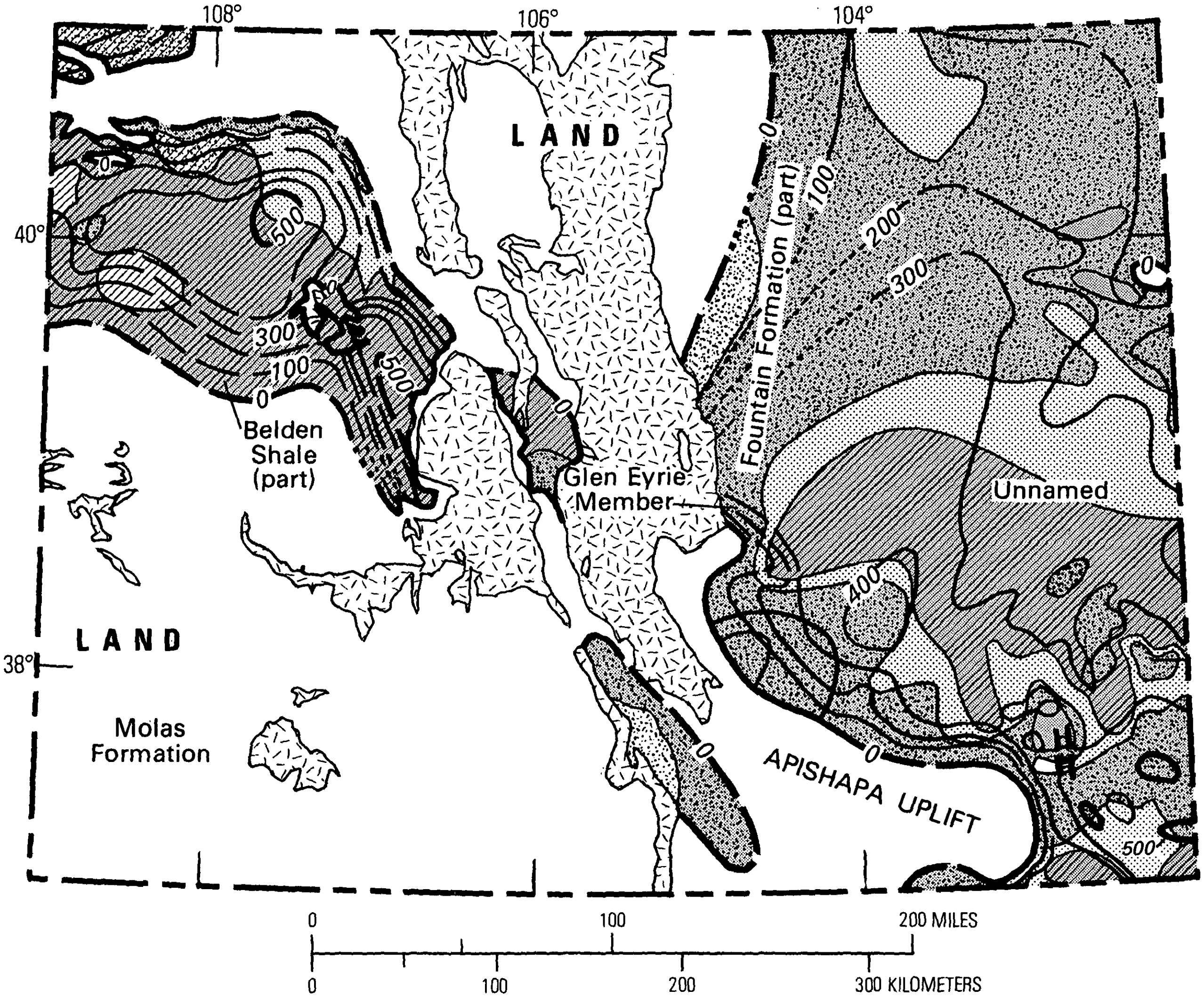


FIGURE 12.—Sub-Pennsylvanian paleogeology (modified from Mallory, 1972, p. 114).

Abrupt facies changes are the rule in the Pennsylvanian of central Colorado. Outcrops, roughly following the edges of Pennsylvanian highlands, may change within a very few miles from continental coarse conglomerates to marine limestones, alternating in cyclic repetition with red shales and sandstones. In many places, correlation of the beds in the clastic strata for more than a few feet is difficult,



EXPLANATION

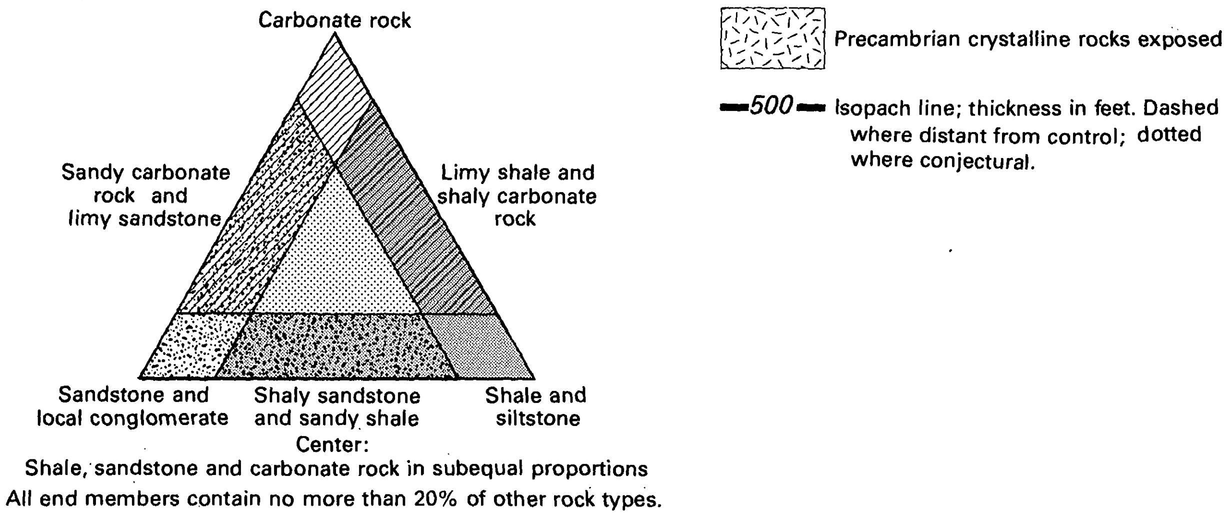


FIGURE 13.—Thickness and lithofacies of rocks mostly of Morrow age (modified from Mallory, 1972, p. 116).

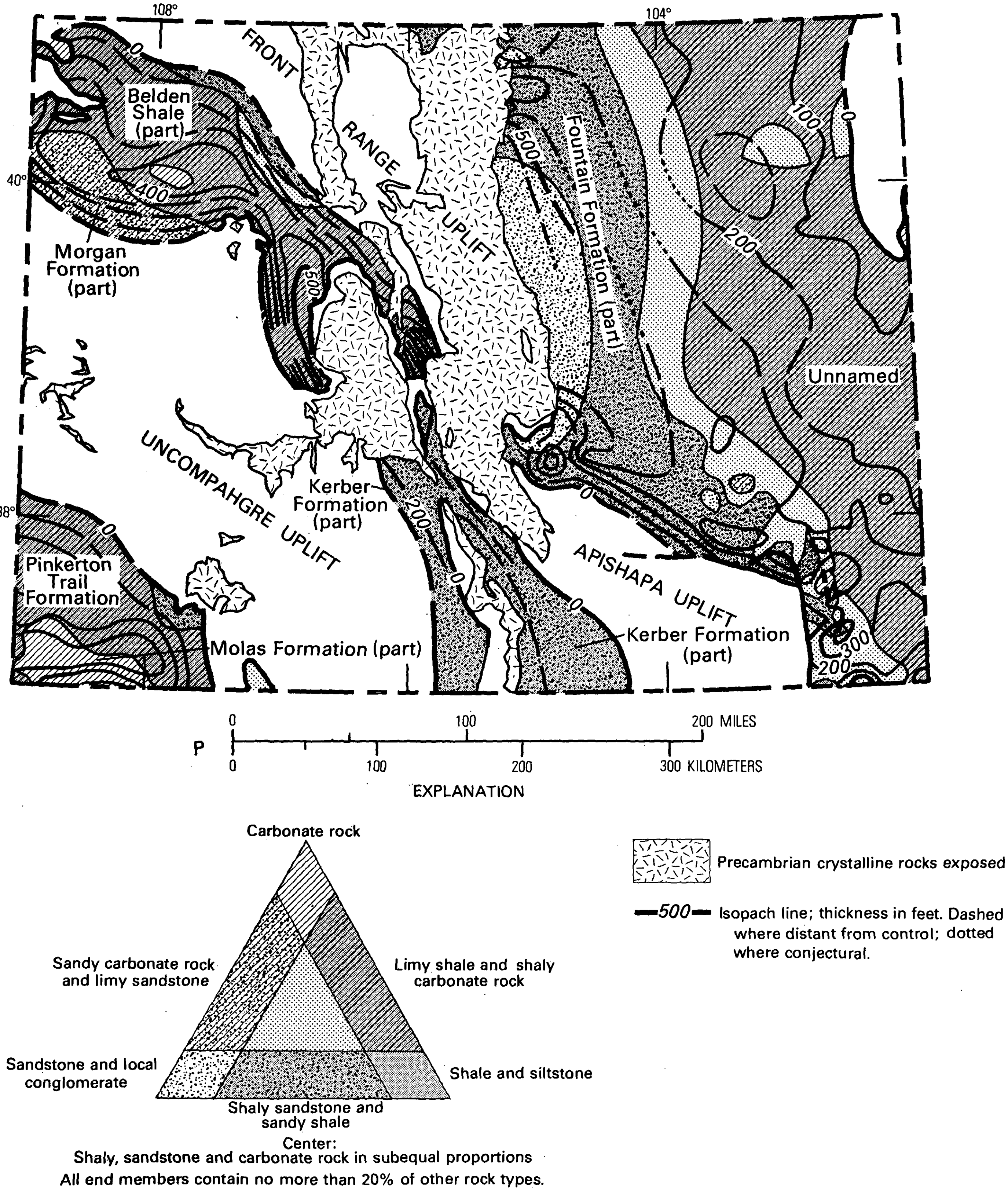


FIGURE 14.—Thickness and lithofacies of rocks mostly of Atoka age (modified from Mallory, 1972, p. 117).

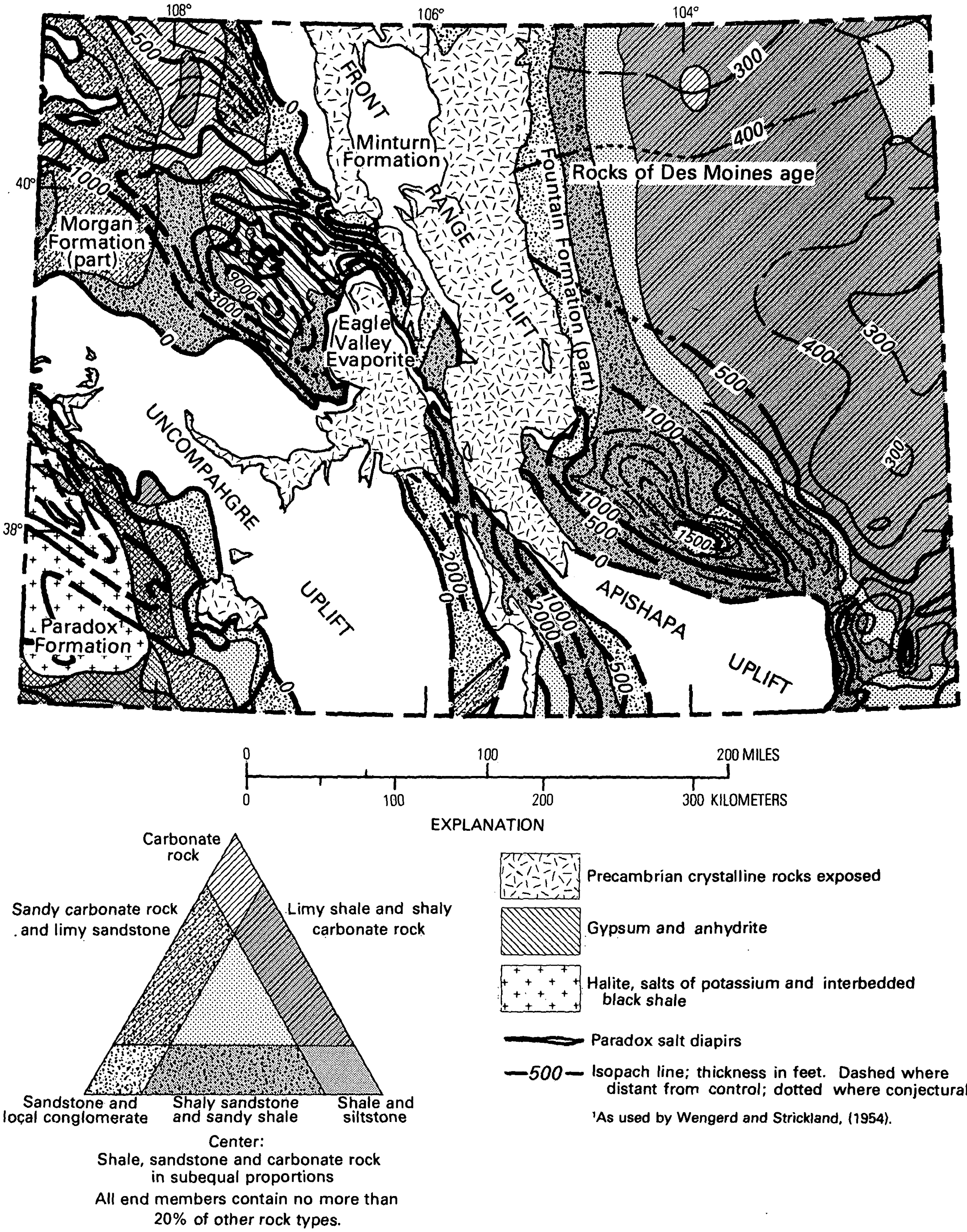


FIGURE 15.—Thickness and lithofacies of rocks mostly of Des Moines age (modified from Mallory, 1972, p. 119).

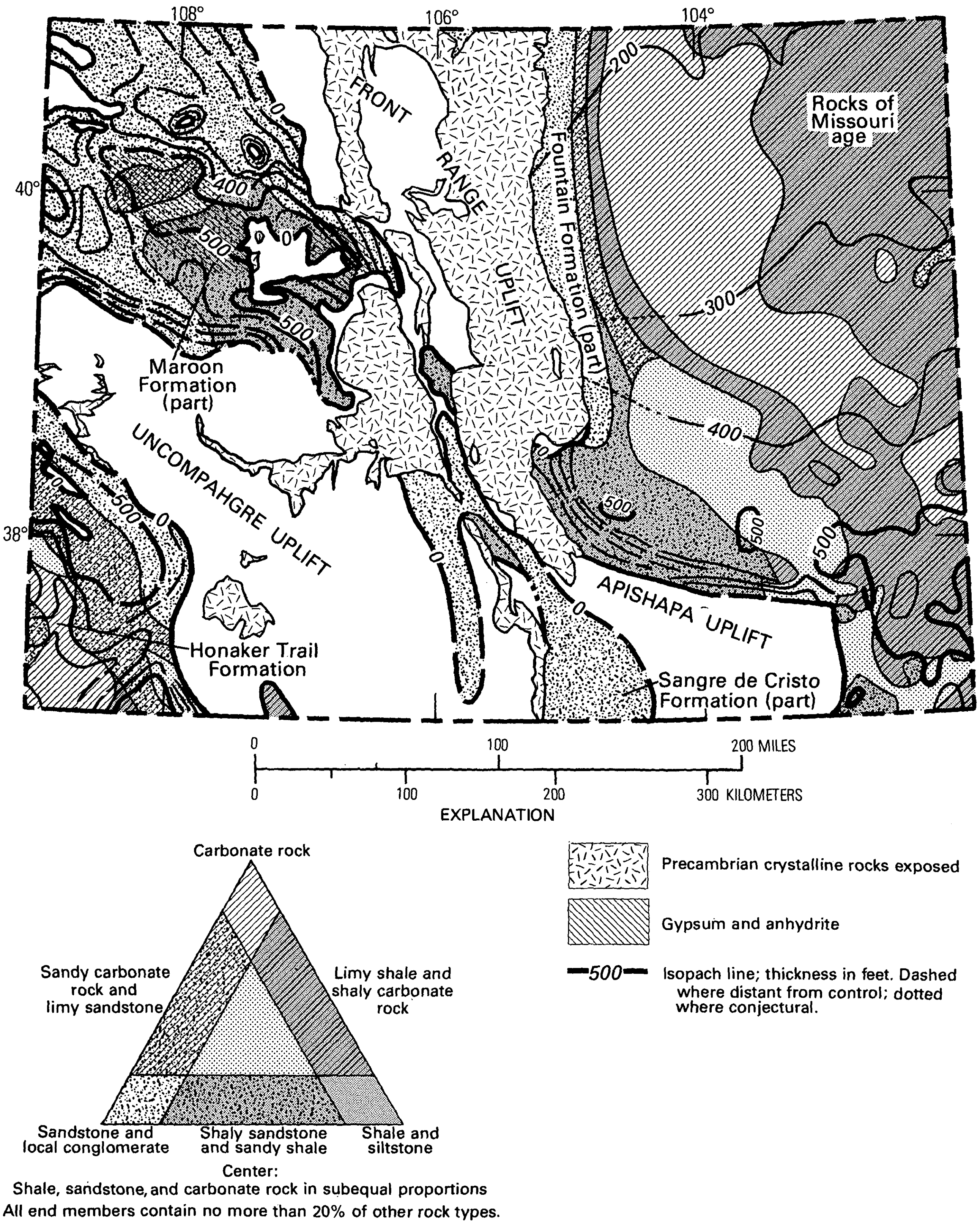


FIGURE 16.—Thickness and lithofacies of rocks mostly of Missouri age (modified from Mallory, 1972, p. 120).

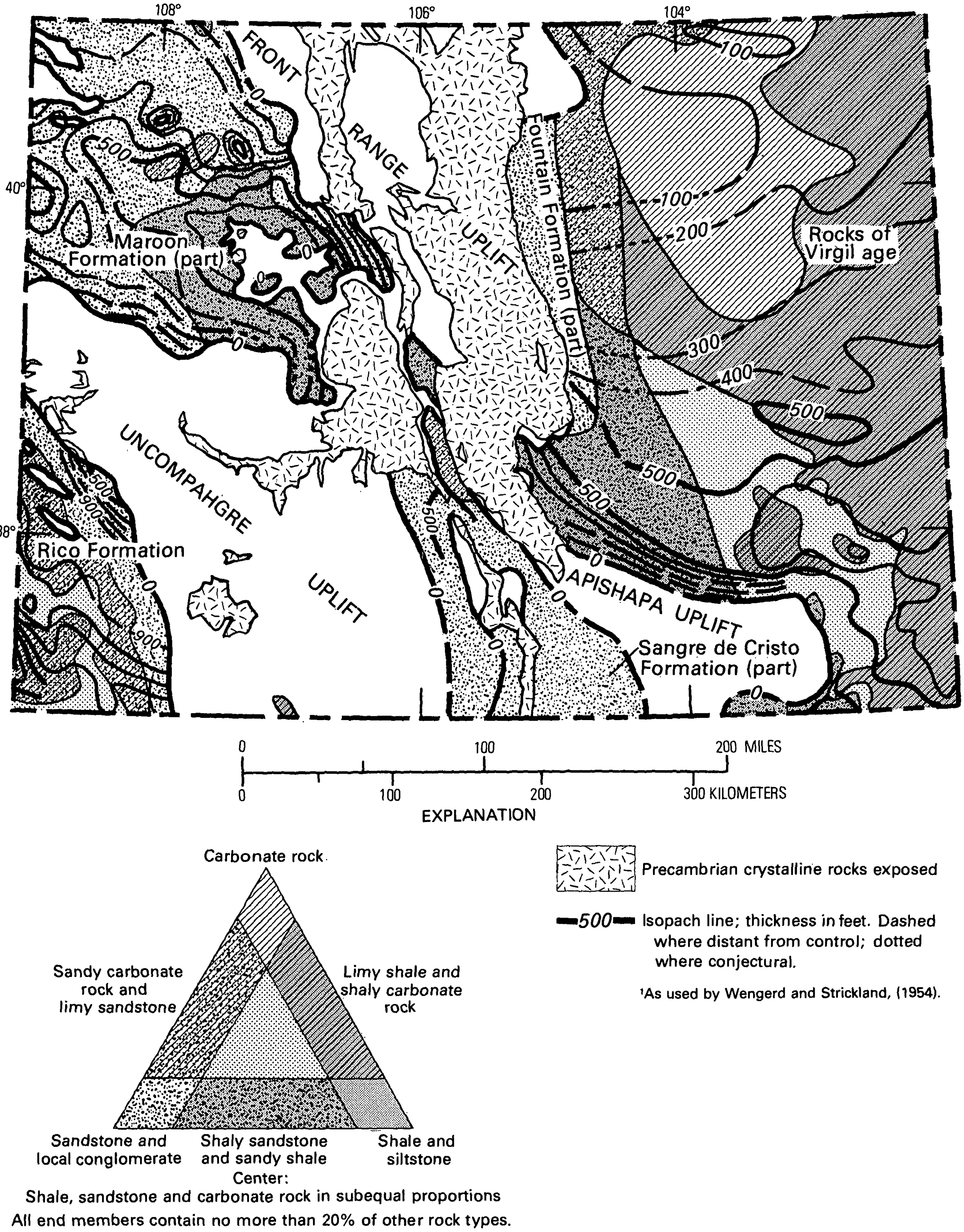


FIGURE 17.—Thickness and lithofacies of rocks mostly of Virgil age (modified from Mallory, 1972, p. 121).

although great lateral continuity exists in some marine beds. The Jacque Mountain Limestone of the Minturn Formation, a bed approximately 6 m (20 ft) thick, is present from the White River Plateau north of Glenwood Springs to the southern Sangre de Cristo Mountains, a distance of more than 300 km (about 200 miles) (fig. 11).

The depositional environment of the Mississippian in Colorado was that of an open unrestricted sea, which was receiving almost no clastic sediments. Present paleogeographic studies suggest that Colorado was at or near the Equator during Carboniferous time, so there can be little doubt that the entire State was in tropical latitudes. The lack of reefs and the rather limited marine fauna, in what are believed to have been tropical conditions, suggest that the depth of water was too great for reef and abundant, varied shallow-water faunas.

Much of the time in the Mississippian Period is not now represented by deposits (figs. 6, 7, 8, and 9). During the latter part of the period, which must have been warm and moist, an extensive karst formed, most of the earlier marine deposits were eroded away, and the residual Molas Formation was deposited.

The *Lepidodendron* forests of central and northwestern Colorado, which formed at about the boundary between Mississippian and Pennsylvanian time, suggest that a brief interval of tropical coal-swamp environment covered at least part of the State before the great upheaval of the Middle Pennsylvanian began.

Early Pennsylvanian environments were of transgressing seas from east and west over a land of low relief, where local uplifts were beginning. The dark-gray muddy deposits, the rather sparse faunas, and the lack of reefs suggest that the turbidity of the seas may have stifled marine growth; limited evaporite deposition at this time may also have prevented abundant life in some areas and certainly suggests that adjacent land areas were arid.

Middle Pennsylvanian environments, in contrast, varied from those of great aridity to those where broad evaporite basins were formed and flanked by massive reef growth. The climate must have been tropical and there appear to have been intermittent dry and moist periods, caused by simple climatic variation through time or by tectonic events, or both. In some localities, particularly the Sangre de Cristo Mountains, the Minturn area, and the Paradox basin, large reef masses formed at the edges of the marine basins. The Weber Sandstone of northwestern Colorado appears to have been deposited as dunes and beaches and suggests an arid climate for that area during Late Pennsylvanian time.

BIOSTRATIGRAPHY

Outcropping Mississippian strata in Colorado, although not very completely dated, are now considered to be mostly of Osage age, but deposits of Kinderhook, Meramec, and Chester age occur in the subsurface, mainly east of the mountains. At the east end of the Uinta Mountains in northwestern Colorado, Meremac and Chester deposits are present in facies unknown elsewhere in the State.

Pennsylvanian rocks whose age can be determined by fossils are mostly of Des Moines age, but the Atoka and Morrow are well documented in central Colorado outcrops, as well as in subsurface occurrences both east and west of the mountains. Missouri strata probably were deposited extensively over the State, but in all outcrops they are unfossiliferous. Documented strata of Missouri age are restricted to the subsurface of the Paradox basin and southeastern Colorado. Virgil rocks have a similar distribution, but marine facies are somewhat more widely distributed than those of Missouri, and occur in the subsurface in the Paradox basin and southeastern Colorado, as well as in strata at the northern end of the Front Range.

As figures 13–17 show, rocks of each of the series of the Pennsylvanian are probably extensively distributed over the State, but most are completely unfossiliferous, and their specific age assignments are based on inferences and probabilities.

The Mississippian faunal succession is very simple in outcrop: a lower, probably Osage brachiopod and coral fauna and a few gastropods and nautiloid cephalopods, mainly in central Colorado, and an upper, very restricted Meramec outcrop on the east side of the Wet Mountains, oolitic in facies and containing endothyrid foraminifera. Rocks of the other Mississippian series are poorly represented or known in the State because they are mostly in subsurface and have been drilled in only a few places for oil and gas. The Chester deposits of northwestern Colorado are dated from plant remains whose age may be in some doubt.

The Pennsylvanian faunal succession is complex in outcrop and in subsurface. The Morrow fauna is limited to a rather poorly defined zone that extends from the northwest corner of the State to near the southeast corner. *Millerella*, a primitive fusulinid, is abundant in most occurrences, along with other foraminifers, bryozoans, brachiopods, and ostracodes.

The Atoka and Des Moines faunas generally are easily recognized by their abundant fusulinids and by a much more diverse biota, especially algae, corals, echinoderms, and trilobites. Within the Des Moines, most of the fossiliferous strata are of the older Cherokee part, and few, if any, late Des Moines faunas are known. An excellent succession of fusulinids is known within Des Moines rocks, and these have been reasonably well documented in northwestern Colorado by Thompson (1945).

Missouri and Virgil faunas are known only from subsurface and have been recognized by their fusulinid content in well cuttings and cores.

A few vertebrates and plants have been found in Pennsylvanian rocks of Colorado, but they have not yet been thoroughly studied.

Girty (1903) included extensive descriptions of Carboniferous fossils from Colorado, differentiating 55 invertebrates from the Mississippian and 170 from the Pennsylvanian.

Of the Mississippian species, 7 were corals, 5 bryozoans, 23 brachiopods, 5 bivalves, 10 gastropods, 3 arthropods, and 2 echinoderms. These, Girty believed, came from two faunal realms, one in central and southwestern Colorado containing corals, bryozoans, gastropods, and productid brachiopods, and a second along the east edge of the Front Range containing bivalves, bryozoans, large ostracodes, and five distinctive brachiopods: *Orthotetes inequalis* Hall, *Spirifer centronatus* Winchell, *Punctospirifer solidirostris* (White), *Cranaena subelliptica* var. *hardingensis* Girty, and *Eumetria woosteri* (White). Girty suggested that, although they are distinct in nature, both faunas represented an early Mississippian (Osage) age for the enclosing Leadville Limestone.

Since Girty's time, a Meramec fauna has been recognized in the Hardscrabble and Beulah Limestones, mainly consisting of endothyrid foraminifera, productid brachiopods, and corals.

In the Pennsylvanian Period, faunas were much more diverse and abundant. Girty described 170 species of invertebrates, including only 2 foraminifers (both fusulinids), 2 sponges, 9 corals, 21 bryozoans, 41 brachiopods, 45 bivalves, a single scaphopod, 34 gastropods, 2 cephalopods, 6 arthropods, and 7 echinoderms from his scattered collections.

The earliest Pennsylvanian rock units in Colorado, the Kerber, Glen Eyrie, and Belden Formations, commonly have two biofacies. At the base, a coaly cyclic sequence of beds contains *Lepidodendron* and closely related plants; above, a marine, usually cyclic series of beds contain a varied assemblage of algae, foraminifera including *Millerella*, bryozoans, brachiopods, mollusks, arthropods including both trilobites and ostracodes, and echinoderms, mostly echinoid parts and crinoid stem segments. Small fish teeth and conodonts are evidence of vertebrate life at this time.

The Belden Formation also contains an Atokan fauna in many localities, particularly in northwestern Colorado, where several species of *Fusulinella* have been described. Other foraminifera, rhomboporoid bryozoa, productid brachiopods, ostracodes, and echinoid fragments are typical of the Atokan part of the Belden.

Minturn Formation faunas are very diverse, depending on sedimentary facies, and their ages are difficult to determine unless they contain fusulinids because most of the faunas are long ranging. Either with or without the fusulinids, which, in many places, are *Fusulina rockymontana* Roth and Skinner and a species of *Wedekindellina*, the faunas normally are composed of brachiopods, bryozoans, corals, echinoderms, and mollusks. In some localities, extremely varied faunas of gastropods, often diminutive, are present, closely resembling the profuse early Des Moines faunas of Central United States. Some beds of the Minturn in central Colorado are made of crinoid stem fragments or coarse monaxon sponge spicules, and nonfusulinid foraminifera are locally very abundant.

Post-Des Moines faunas, rare in Colorado, are mostly fusulinids. Ingleside Formation limestones containing fusulinids similar to *Triticites cullomensis* Dunbar and Condra are characteristic in the northern Front Range and in the subsurface of eastern Colorado.

Mississippian fossils are not common in Colorado, and those that are present are not easily collected. On Fossil Ridge, however, northeast of Gunnison, corals are well preserved and abundant, and near Meredith, east of Basalt, brachiopods, usually species of *Spirifer* or productid forms, and large horn corals may be easily collected.

Pennsylvanian fossils, by contrast, are very abundant and diverse and can be found weathered from shale beds at many places in the State. From the many localities, four may be selected as outstanding and easily accessible: (1) At Molas Pass, near Silverton, great numbers of corals and brachiopods are present in the limestone beds of the cyclothems. (2)

At McCoy in central Colorado, Roth and Skinner (1930) described *Fusulina rockymontana*, as well as many other microfossils. The limestones east and west of Rock Creek at this locality have yielded great numbers of these and other Des Moines fossils. (3) At Wellsville, southeast of Salida on the south side of the Arkansas River, are excellent outcrops of abundantly fossiliferous early Pennsylvanian faunas. (4) On Juniper Mountain, west of Craig, are many well-exposed and excellently preserved invertebrates and fusulinids.

IGNEOUS AND METAMORPHIC ROCKS

No igneous or metamorphic rocks of Carboniferous age are found in Colorado, although some of the Carboniferous strata were metamorphosed by Laramide mountain building.

ECONOMIC PRODUCTS

Although a few feeble attempts have been made to mine coaly beds of the Kerber and Belden Formations in south-central Colorado, no production of Carboniferous coal is known.

However, oil from Carboniferous beds in the State is big business, especially if production from the Weber Sandstone at Rangely is included. In 1976, about 900,000 barrels of oil were produced along the eastern border of the State; although some of this oil came from beds that may be either Mississippian or Pennsylvanian, most of the total amount was from Mississippian beds.

If we assume that the Weber Sandstone is at least partly Carboniferous, then Pennsylvanian production in 1976 was 21,619,918 barrels (State of Colorado Oil and Gas Conservation Commission, 1976). The Weber, however, produced 20,781,123 barrels of this, and a small part of the total may have come from the Lyons Sandstone, which is probably of Permian age.

Gas production is also of importance to the State; in 1976, probable Carboniferous production was about 13 billion cubic feet.

Excellent prospects for Carboniferous oil and gas production probably still exist in the eastern plains, in the Paradox basin, and in the northwestern part of the State.

A large part of Colorado's lead, zinc, gold, and silver has come from Laramide replacement of the Leadville Limestone. In 1976, more than $24 million in zinc was produced, and lead production was about $9 million. Production of gold and silver was $5,459,111 and $14,292,695, respectively; most of it was from Leadville and Gilman.

Many years ago, iron for the furnaces of the Colorado Fuel and Iron Co. was produced from replacement in the Leadville Limestone of the northern Sangre de Cristo Mountains, but the volume was limited and mining ceased long ago. Some copper has been produced from Carboniferous red beds at several localities, but none is now being mined.

Many localities have produced uranium, mostly in small amounts, from Carboniferous rocks of the Belden Formation in central Colorado. Most uranium from the Belden Formation is produced from the Pitch mine in the Monarch Pass-Marshall Pass district, where the uranium has been precipitated in and around carbonaceous beds of cyclothems. The Pitch mine is along a reverse fault on the southwest corner of the Sawatch Range; the Belden is faulted against the granite and is highly brecciated. Ore valued at several million dollars was produced in the early 1960's, after which the mine lay dormant until the recent revitalization of the uranium industry. The Homestake Mining Co. now owns the mine and has drilled its rocks extensively. They are planning production from an open pit and have about 20 years' production now on reserve.

The famous Yule Marble, used in the Lincoln Memorial and the Tomb of the Unknown Soldier, is Leadville Limestone metamorphosed by a Laramide intrusion in an area south of Glenwood Springs. About $7 million worth of this snow-white marble was produced between about 1900 and 1940, but the remoteness of the area has led to the closing of the quarry.

Many limestone quarries are in Carboniferous strata along the Front Range, but there is little production elsewhere at present, except east of Monarch Pass, where the Colorado Fuel and Iron Co. quarries Leadville Limestone as flux for its smelters in Pueblo.

Lyons Sandstone of the Front Range, at one time considered to be of Carboniferous age, is now thought to be Permian and so is not included in this discussion.

Gypsum is produced in considerable quantities from Pennsylvanian rocks, particularly at Gypsum in west-central Colorado and at Coaldale on the Arkansas River. Potash, from evaporitic beds of the Paradox basin, is produced just over the State border in Utah; it is not yet being mined in Colorado, although it has economic potential.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— New Mexico

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Historical review and summary of
areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in New Mexico

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—NEW MEXICO

By Augustus K. Armstrong, Frank E. Kottlowski,[1] Wendell J. Stewart,[2] Bernard L. Mamet,[3] Elmer H. Baltz, Jr., W. Terry Siemers,[4] and Sam Thompson III [5]

ABSTRACT

Initial Lower Mississippian deposits of New Mexico are Tournaisian (pre-Zone 7) in age and are unconformable on older rocks of Late Devonian or Precambrian age. These Mississippian rocks were laid down during transgression of the sea across an abraded surface of very low relief. Early and middle Tournaisian marine transgression began in the southwestern and south-central part of the State and deposited the Keating (207 m), Caballero (18 m), and Lake Valley (180 m) Formations on a carbonate platform. By the end of Tournaisian (Osagean) time, epicontinental seas had flooded the central and northern part of the State and had deposited the Kelly Limestone (35 m) and Espiritu Santo Formation. The latter is a sequence of subtidal to supratidal quartz sandstones and carbonate rocks. Leadville Limestone (50–100 m) in the San Juan Basin is a time-stratigraphic equivalent to the Espiritu Santo Formation (35 m) and is an eastern extension of part of the Redwall Limestone of Arizona. The Zuni Highlands and remnants of the transcontinental arch, the Pedernal Highlands, were two low islands. The end of Tournaisian (Osagean) time was marked by marine regression, regional uplift, and erosion of Tournaisian carbonate sedimentary deposits.

A major regional marine transgression took place in Viséan (Meramecian) time and is represented by the massive encrinites of the Hachita Formation (107 m) in southwestern New Mexico, the deeper water basin carbonate rocks of the lower part of the Rancheria Formation (46 m) in south-central New Mexico, and part of the Tererro Formation (18 m) in north-central New Mexico.

The Cowles Member (10 m) of the Tererro Formation indicates that sedimentation ceased in northern and central New Mexico in late Viséan (early Chesterian) time. In southwestern New Mexico, the Paradise Formation (134 m) represents shallow-marine sedimentation and ranges from Zone 15 (Meramecian) through Zone 19 (late Viséan and Namurian or end of Chesterian). The upper part of the Rancheria Formation (69 m) and the Helms Formation (50 m) of south-central New Mexico are a deeper water facies of the Paradise Formation.

Pennsylvanian sedimentary rocks in northern, central, and southern New Mexico truncate Mississippian sedimentary rocks of Namurian, Viséan, and Tournaisian age. The Pennsylvanian sequence is as much as 2,300 m thick in north-central New Mexico and is more than 600–900 m thick in the Delaware, Orogrande, and Pedregosa Basins areas. The most complete section is in the Big Hatchet Mountains, where a relatively continuous sequence is essentially conformable with underlying Chesterian strata and overlying Wolfcampian beds and contains Morrowan through Virgilian faunal equivalents.

Rocks of the Pennsylvanian System are present throughout New Mexico except where they were not deposited or where they have been removed by subsequent erosion. The major positive features during the Pennsylvanian were the Uncompahgre, Sierra Grande, Zuni, and Pedernal uplifts and the central-basin platform, a shallow-marine feature. Many of the sections consist of a lower clastic phase, a middle limestone unit, and an upper intertongued limestone and clastic-rock sequence, but facies change greatly from units on the margins of uplifts into dark carboniferous basinal sequences. These sequences indicate transgression of the sea in the Early Pennsylvanian, maximum inundations in the Middle Pennsylvanian and regression near the close of the period.

Age determinations are based mainly on fusulinid zones, but numerous other marine fossils are present. Pennsylvanian rocks have yielded significant quantities of oil and gas, contain some local coal lenses, are host rocks for base-metal and fluorite-barite-galena ores, and are quarried for road metal, flagstone, and material to make cement, brick, and tile.

MISSISSIPPIAN SYSTEM

PREVIOUS WORK

Fossils from Mississippian rocks of New Mexico (fig. 1) were first identified by White (1881) at Lake Valley, N. Mex. Cope (1882a, b) referred to the rocks at Lake Valley and proposed the name Lake Valley Formation.

Herrick (1904) used the name Kelly Limestone for the Mississippian rocks of the Magdalena

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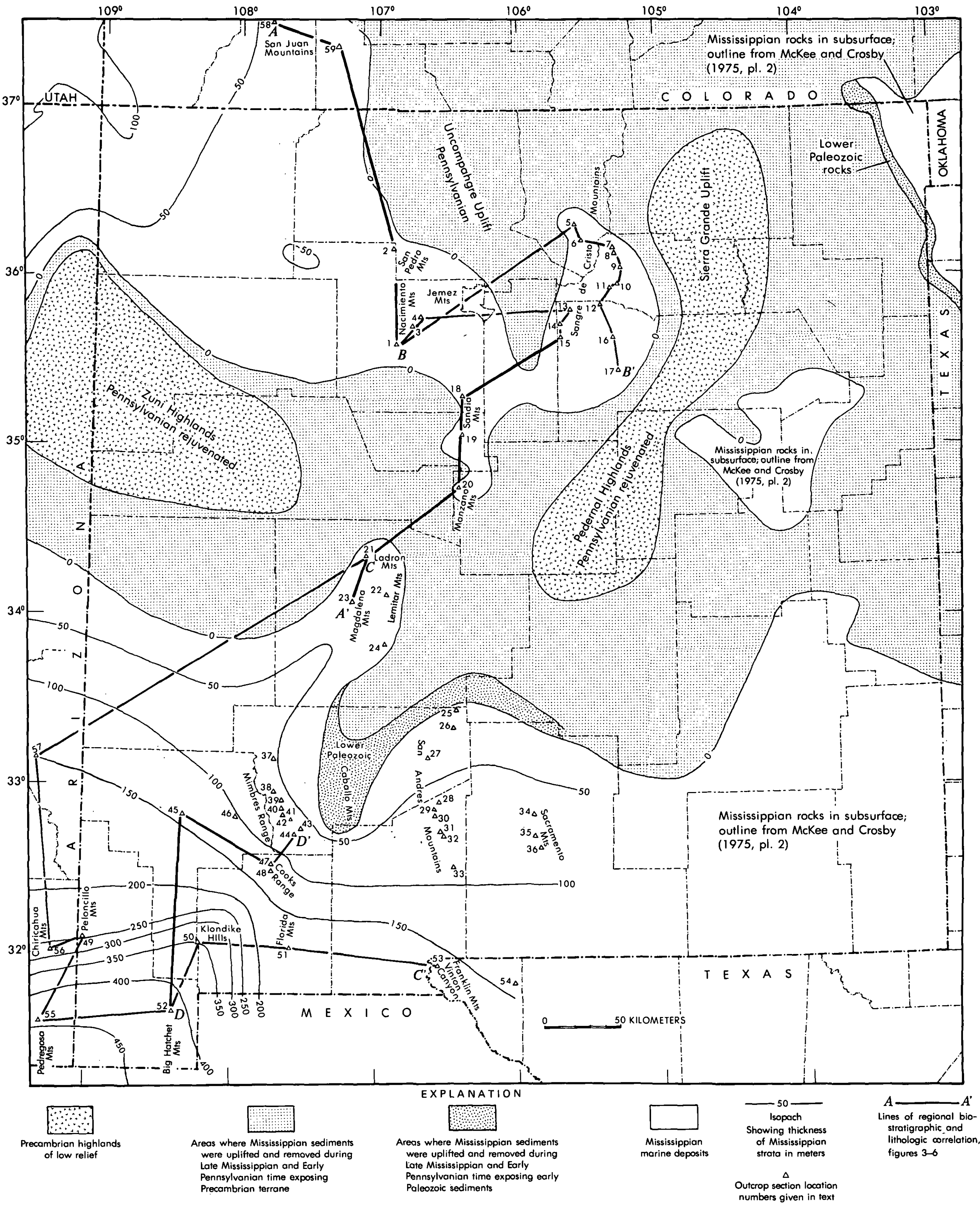


FIGURE 1.—Locations of Mississippian outcrop sections, pre-Pennsylvanian Mississippian isopachs and paleogeography, and locations of lines of biostratigraphic correlation charts (figs. 3–6) in New Mexico and nearby areas.

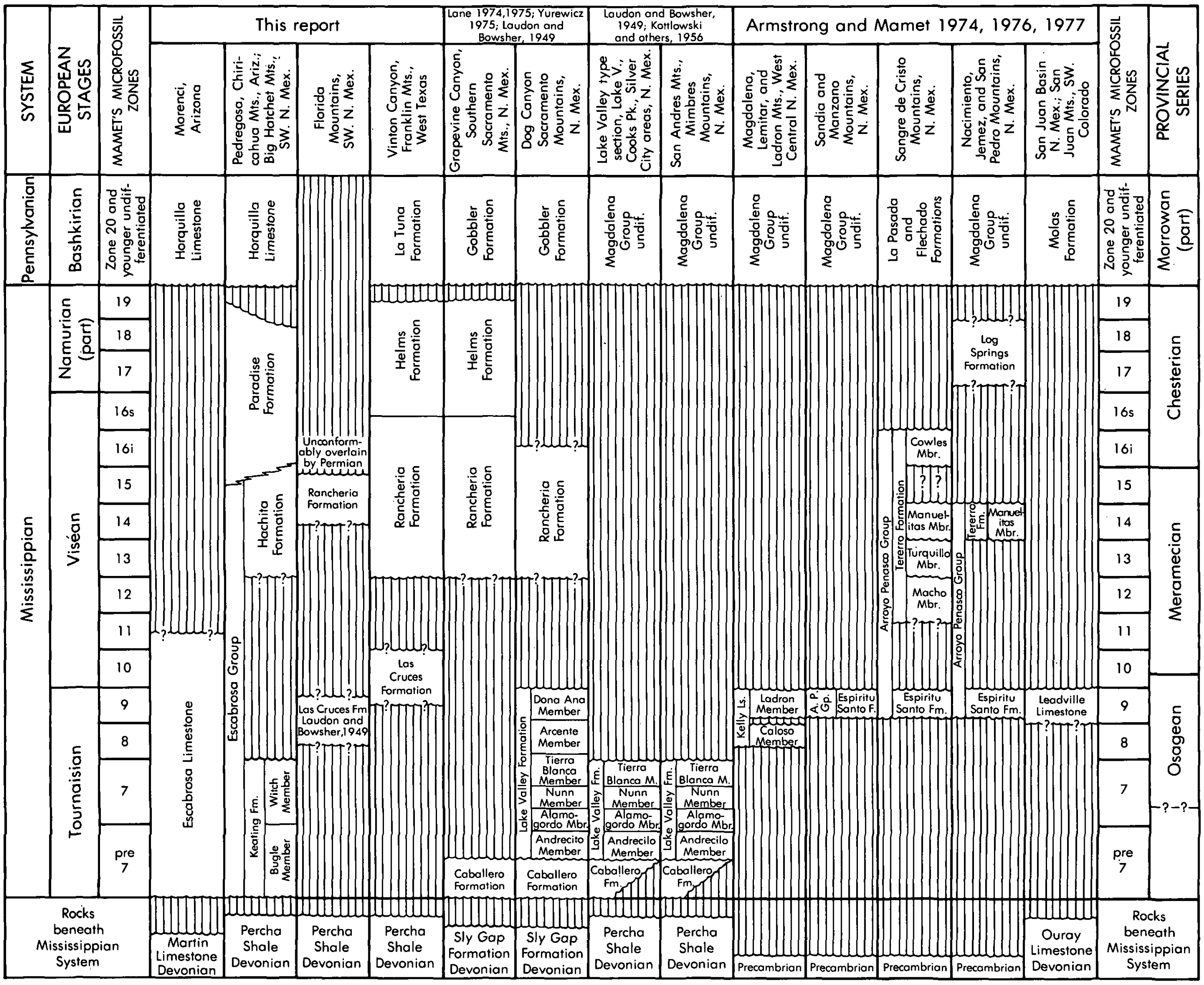


FIGURE 2.—Correlation chart of Mississippian rocks of New Mexico and nearby areas.

mining district. Laudon and Bowsher (1941, 1949) divided the Mississippian System and south-central New Mexico into five formations and divided the Lake Valley Formation into six members (fig. 2). They named the Caballero Formation for rocks of Kinderhookian age in the Sacramento and San Andres Mountains and at Lake Valley, N. Mex., and named the Las Cruces and Rancheria Formations for Meramecian rocks in the Franklin Mountains of western Texas and in the southern San Andres and Sacramento Mountains of New Mexico. They restricted Beede's (1920) designation of the Helms Formation to beds of Chesterian age. Laudon and Bowsher (1941, 1949) gave macrofaunal lists for the Lake Valley, the Rancheria, and the Helms Formation.

A section of pre-Pennsylvanian predominantly carbonate rocks of Paleozoic age in the San Pedro, Nacimiento, Jemez, Sandia, and Sangre de Cristo Mountains of northern New Mexico was studied by Read and others (1944), who first recognized the distinctiveness of these rocks. They mapped them as the lower limestone member of the Pennsylvanian Sandia Formation of the Magdalena Group. The lower gray limestone member was mapped and described by Wood and Northrop (1946) in the San Pedro and Nacimiento Mountains and by Northrop and others (1946) in the southeastern foothills of the Sangre de Cristo Mountains.

In 1955, Armstrong proposed the name Arroyo Penasco Formation for the lower gray limestone member of the Sandia Formation in the San Pedro, Nacimiento, Sandia, and Sangre de Cristo Mountains of north-central New Mexico.

Fitzsimmons, Armstrong, and Gordon (1956) listed a fauna of St. Louis (Meramecian) age that consisted of megafossils from the top of the Arroyo Penasco Formation in exposures on the northwestern side of the San Pedro Mountains and from its type section. Armstrong (1958a, 1967) described part of the Meramecian endothyrid fauna of the Arroyo Penasco Formation and demonstrated that at the type section and in the Sangre de Cristo Mountains, the rocks had the same lithologies and endothyrid species and were thus of the same age.

Because of the discovery of the *Spinoendothyra spinosa* (Chernysheva) microfauna in cherts of the basal carbonate rocks by Lee Holcomb of the Shell Oil Company, Armstrong (1963, p. 115; 1965, p. 133; 1967; Armstrong and Holcomb, 1967) determined the age of the Arroyo Penasco Formation to be late Osagean and Meramecian.

In 1960, Baltz and Read divided the pre-Pennsylvanian sandstone and carbonate rocks of the Sangre de Cristo Mountains into two newly named formations, the Espiritu Santo (Devonian(?)) and the Tererro (Kinderhookian to Meramecian). The Tererro Formation was divided into three members, in ascending order, the Macho, Manuelitas, and Cowles Members.

Armstrong (1967) considered Baltz and Read's (1960) Espiritu Santo and Tererro Formations of the Sangre de Cristo Mountains to be laterally equivalent parts of his (Armstrong, 1955) Arroyo Penasco. He recognized the Arroyo Penasco throughout northern New Mexico. Armstrong and Mamet (1974) raised the Arroyo Penasco to group rank. The age and nomenclature of these rocks are shown in figure 2.

The Escabrosa Limestone of Mississippian age was named by G. H. Girty (*in* Ransome, 1904) for the lower Carboniferous section in the Escabrosa Cliffs, west of Bisbee, Cochise County, southeastern Arizona. The Escabrosa Limestone in the Chiricahua Mountains of southeastern Arizona and southwestern New Mexico was elevated by Armstrong (1962, p. 5) to the Escabrosa Group, which he divided into two newly named formations—the Keating Formation, consisting of two members, A and B, and the overlying Hachita Formation (fig. 2). This nomenclature was extended into Luna, Hidalgo, and Grant Counties, southwestern New Mexico. The two informal members, A and B, of the Keating Formation were named the Bugle and Witch Members of the Keating Formation by Armstrong and Mamet (1978). The names are taken from the Bugle Ridge and Witch Well, published on the U.S. Geological Survey's 1:62,500 scale topographic map of the Big Hatchet Quadrangle. The type sections for the members are E½ sec. 30, T. 29 S., R. 15 W., northeast side of the Big Hatchet Mountains. Armstrong (1962) illustrated and described the brachiopod and coral faunas of the Escabrosa Group.

The Paradise Formation was named by Stoyanow (1926) for outcrops a few kilometers east of the old mining camp of Paradise, on the east side of the Chiricahua Mountains. The macrofauna of the Paradise Formation in the Chiricahua Mountains was studied and described by Hernon (1935). Zeller (1965) gave M. K. Elias' macrofossil lists of the Paradise Formation for the Big Hatchet Mountains outcrop.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the New Mexico Bureau of Mines and Mineral Resources.

GEOLOGIC SETTING

LOWER BOUNDARY OF THE MISSISSIPPIAN

The Leadville Limestone of the San Juan Mountains, Colo., and in the subsurface of San Juan County, N. Mex., has at its base a Tournaisian (Osagean) microfossil assemblage of Zone 9 (Armstrong and Mamet, 1976, 1977), and the underlying Ouray Limestone contains a well-defined fauna of Late Devonian brachiopods near its top (figs. 2, 3). In north-central New Mexico, Mississippian rocks of Zone 9 age unconformably overlie Precambrian metamorphic and igneous rocks (figs. 2–4). In west-central New Mexico, Zone 8 rocks overlie Precambrian rocks. In the Sacramento Mountains, pre-Zone 7 rocks overlie shale and limestone of Late Devonian age. In the northern San Andres Mountains, Tournaisian-age rocks rest unconformably on Upper Devonian shale and marl; in the southern part of the range, pre-Zone 7 beds unconformably overlie the Upper Devonian rocks. In the Mimbres Range and Silver City region, pre-Zone 7 rocks unconformably overlie the Upper Devonian Percha Shale. In the southwestern part of the State, pre-Zone 7 carbonate rocks unconformably overlie the Upper Devonian Percha Shale (figs. 5, 6).

UNITS OVERLYING THE MISSISSIPPIAN

Pennsylvanian rocks unconformably overlie the Mississippian at most places in New Mexico. In the mountains of north-central New Mexico, the Mississippian Arroyo Penasco Group and the Log Springs Formation are overlain by Pennsylvanian sedimentary rocks. The Lower Pennsylvanian Molas Formation overlies the Leadville Limestone in the subsurface of San Juan County. The hiatus probably represents active erosion of Mississippian rocks during Zones 17 to 20 (Chesterian to Morrowan) time.

In west-central New Mexico in the Lemitar, Ladron, and Magdalena Mountains, the Mississippian carbonate rocks are unconformably overlain by nearshore clastic rocks of the Pennsylvanian Sandia Formation. In the Coyote Hills southwest of Socorro, Tertiary volcanic rocks overlie and truncate the Mississippian outcrops.

Pennsylvanian sedimentary rocks in northern, central, and southern New Mexico truncate Mississippian sedimentary rocks of Namurian, Viséan, and Tournaisian (Chesterian to Osagean) age. In the Big Hatchet Mountains, however, the contact between the Paradise Formation and the Horquillo Limestone is at the boundary between Zones 19 and 20, and a hiatus, if present, must be minimal. In the Florida Mountains, the Mississippian is unconformably overlain by Wolfcampian (Permian) carbonate rocks.

STRUCTURAL EVENTS DURING THE MISSISSIPPIAN

An idealized illustration of the total thickness of Mississippian rocks is shown in figure 1. The map shows disconnected areas of Mississippian rock remnants of extensive sheets that were dissected and beveled in northern and central New Mexico in Namurian (Chesterian) time and throughout the entire State by erosion on structurally active features in Pennsylvanian and Permian time.

Marine flooding of the State began in early Tournaisian (pre-Zone 7) time in the southwestern and south-central part of the State and formed the Escabrosa carbonate platform. By the end of Tournaisian (Osagean) time, epicontinental seas had flooded the northern and central parts of the State. Two possible low islands may have existed, the Zuni Highlands and remnants of the transcontinental arch, the Pedernal Highlands (fig. 1). The Espiritu Santo Formation (figs. 2–4) is composed of carbonate tidal deposits in the Sangre de Cristo, Sandia, Nacimiento, and San Pedro Mountains of north-central New Mexico (Armstrong, 1967; Armstrong and Mamet, 1977; Vaughan, Eby, and Meyers, 1977). The Leadville Limestone is the time-stratigraphic equivalent in the San Juan Basin of northwestern New Mexico and is an eastern extension of the Redwall Limestone of Arizona. The end of Tournaisian (Osagean) time was marked by a marine regression, a regional uplift, and extensive erosion of the Tournaisian (Osagean) carbonate deposits (figs. 2–6).

The geographic and stratigraphic extent of this hiatus at the end of the Tournaisian (Osagean) is shown in figures 2–6. A major regional marine transgression took place in middle Viséan (Meramecian) and is represented by the massive encrinites of the Hachita Formation in the southwestern part of the State, the deeper water carbonate rocks of the Rancheria Formation in the southern San Andres and Sacramento Mountains, and the Turquillo Member (Meramecian) of the Tererro Formation in north-central New Mexico. Late Viséan carbonate rocks of Zone 16i (Chesterian) are also widely distributed in disjunct outcrops. These are the Cowles Member of the Tererro Formation, the upper part of the Rancheria Formation, and the lower part of the Paradise Formation.

Marine sedimentation ceased in northern and central New Mexico at the end of Zone 16i time. In southwestern New Mexico, marine sedimentation continued through Zone 19 time. The Paradise Formation (figs. 2, 5, 6) is a series of shallow-water shoaling to nearshore oolitic carbonate rocks to plant-fossil-bearing crossbedded sandstones and siltstones. The Helms Formation to the east appears to be a deeper water facies equivalent of the Paradise Formation (figs. 2, 5).

The Log Springs Formation of the San Pedro, Nacimiento, Jemez, and Sandia Mountains of north-central New Mexico unconformably overlies the Arroyo Penasco Group and in turn is truncated by limestones of Pennsylvanian (Zone 20) age. The Log Springs Formation is composed of terrigenous, red-brown iron-rich shale, siltstone, and lithic to arkosic conglomerate formed of angular cobbles of Mississippian and Precambrian rocks. It is interpreted as being post-Zone 16i and pre-Zone 20 in age (Namurian (Chesterian)) and represents, in part, a regolith and tectonically derived sediments washed into small basins adjacent to uplifted, faulted, and tectonically active highlands (figs. 2–4).

The biostratigraphy and facies relations of the Mississippian carbonate rocks in the San Andres

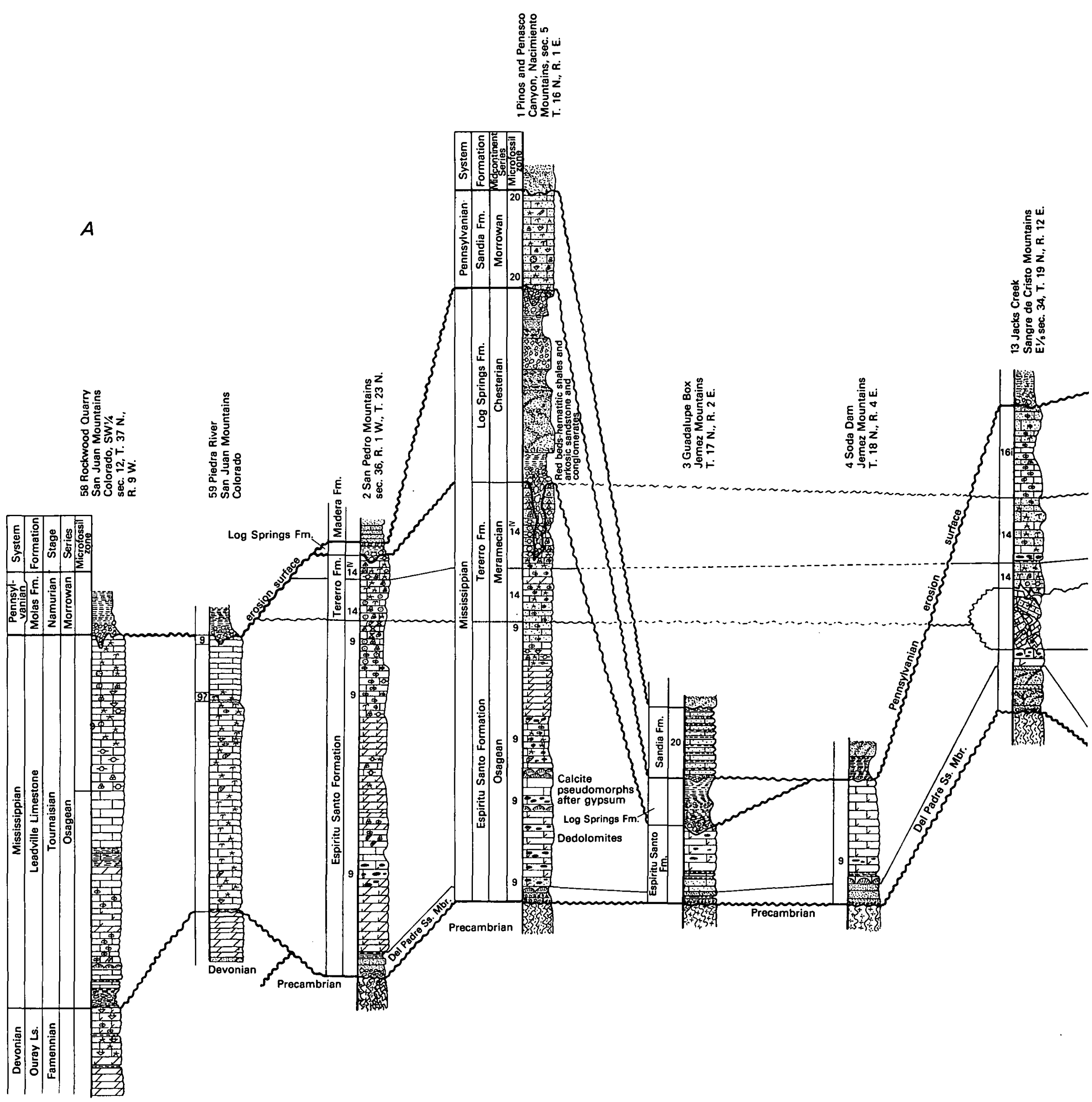


FIGURE 3.—Regional biostratigraphic and lithologic correlation of Mississippian strata along line A–A' from the San Juan Mountains of southwestern Colorado to the San Pedro, Nacimiento, Jemez, Sangre de Cristo, Sandia, Manzano, Ladron, and Magdalena Mountains of northern New Mexico. Line of section is shown in figure 1; symbols are explained in figure 5.

and Sacramento Mountains of south-central New Mexico are well described. The first modern studies of the Mississippian of the Sacramento and San Andres Mountains were by Laudon and Bowsher (1941, 1949) on the stratigraphy and megafaunas of the Mississippian Caballero and Lake Valley Formations and the bioherms of the Lake Valley. They described the wedge-on-wedge relations between these Lower Mississippian strata and the Upper Mississippian Rancheria and Helms Formations.

James Lee Wilson (1975, p. 125) stated the problem of the stratigraphic relationship of the Lake Valley Formation to the Rancheria and Helms Formations:

> The distribution of Mississippian beds in southern New Mexico is stratigraphically puzzling. Fine-grained and siliceous limestone and shale of the Late Mississippian Rancheria and Helms formations are apparently the only strata representing the system in the Franklin and Hueco Mountains around El Paso, whereas to the north and west in New Mexico thick crinoidal Osagean beds are present; only Early Mis-

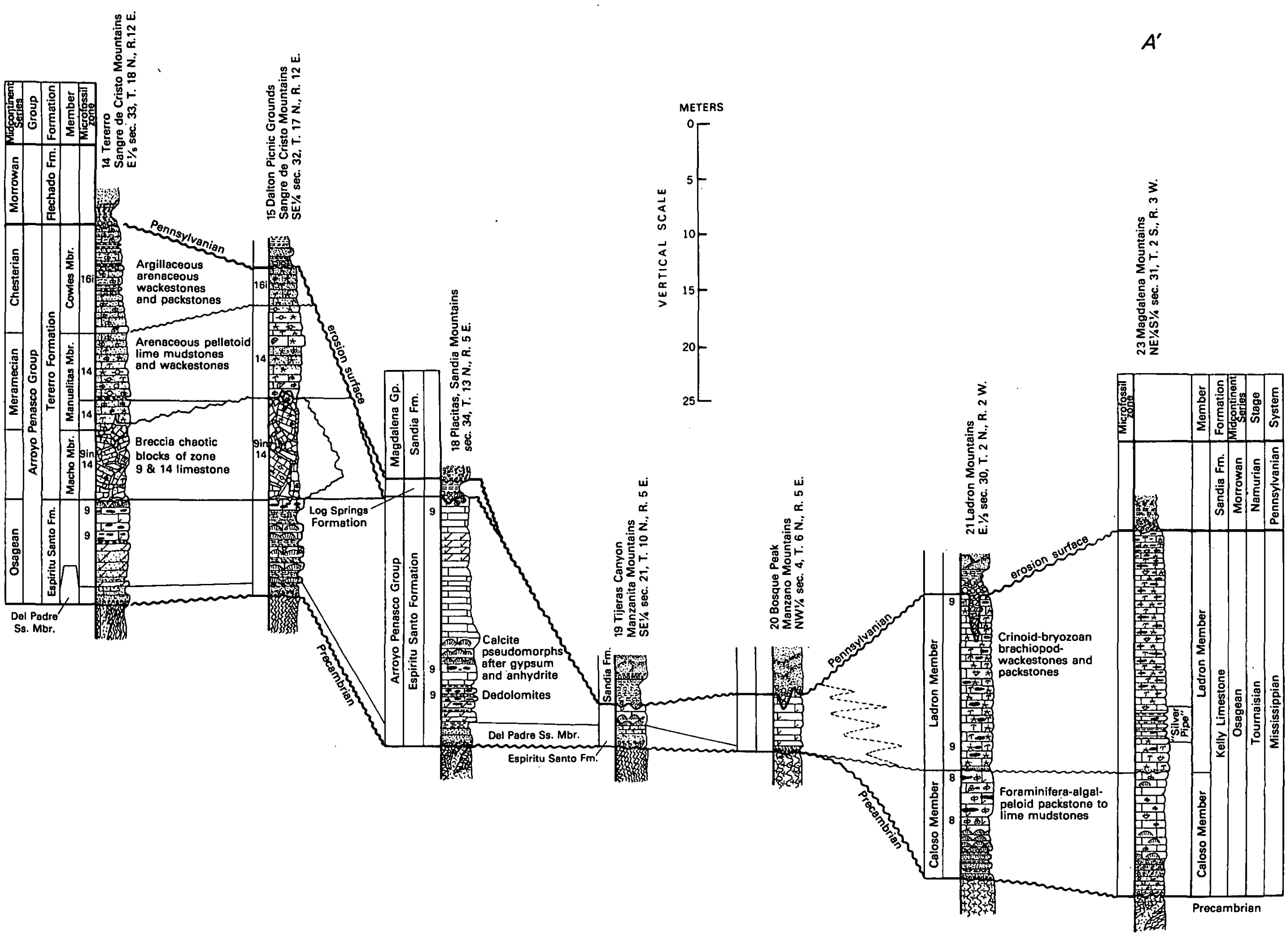


FIGURE 3.—Continued.

sissippian strata are known in the northern San Andres and Sacramento Mountains. Is this reciprocal distribution of major parts of the Mississippian best explained by differential tectonic uplift and erosion, or could it be in some way depositional in origin? Some data bearing on the problem are given below.

Regional Mississippian isopachs of southern New Mexico by A. K. Armstrong (1962) and F. E. Kottlowski (1970) indicate a considerable thickness (500 meters) of mostly Early Mississippian shelf encrinite in the southwestern part of the state and in adjacent Arizona. Control points for this occur in the Big Hatchet Mountains, Sierra de Palomas, and the Los Chinos oil test of Petroleos Mexicanos. These thick open marine shelf deposits grade eastward into thinner and more micritic limestone with some shale and with large scattered bioherms as displayed by the Sacramento Mountain outcrops. Pray (1961) and Meyers (1973) have demonstrated that the Sacramento Mountain strata represent deposition down a gentle southward slope. Correlation of detailed stratigraphic profiles along the north-south trending Sacramento scarp shows this clearly. Well-bedded shelf encrinites, with lens-shaped Waulsortian micritic bodies change southward to larger, equidimensional Waulsortian mud mounds surrounded by encrinites and dark micritic limestones. The strata thin unit by unit and together form a wedge tapering out in the Franklin and Hueco Mountains (Lane, 1974). The absence of Osagean Mississippian in the El Paso area and the presence there of only fine-grained and siliceous Meramecan [sic] and Chesteran strata were first pointed out by Laudon and Bowsher (1949). Late Mississippian strata wedge out northward, the uppermost Rancheria and Helms being present only in the southern Sacramento Mountains. Was an originally widespread sheet of Osagean limestone eroded from this southern area along a trend cutting transverse to the later Paleozoic structural grain, across both the Diablo platform and the Pedregosa basin? Or was the later Paleozoic depositional topography already set in Mississippian time? Could there have been an extended Oro Grande-Pedregosa basin in New Mexico and Chihuahua which was starved of sediment during Kinderhookian and Osagean time and filled with the basinal Rancheria during Meramecan [sic]? Armstrong (1962), the writer (Wilson, 1969, 1971 [1970]), and Kottlowski (1970) have all suggested the latter interpretation. The following lines of evidence indicate that this is

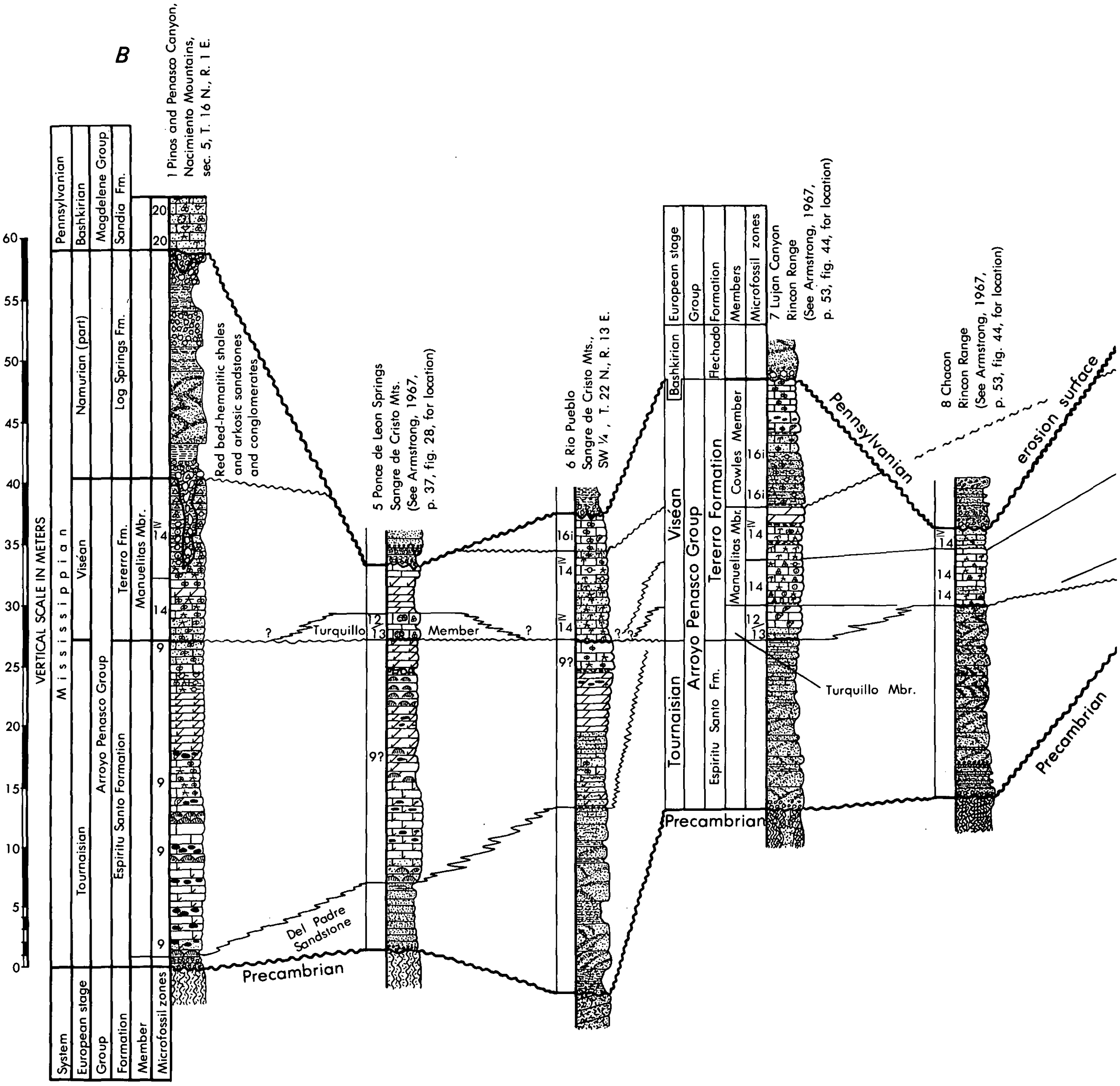


FIGURE 4.—Regional biostratigraphic and lithologic correlation of Mississippian strata along line *B–B′* from the Nacimiento and Sangre de Cristo Mountains of north-central New Mexico. Line of section is shown in figure 1; symbols are explained in figure 5.

reasonable but do not offer compelling proof. No paleontologic evidence of Early Mississippian beneath the Rancheria has yet been presented.

The facies of the southward tapering and prograding wedge of Early Mississippian strata in the Sacramento Mountains are best explained by deposition on a gentle slope south into a starved basin; e.g. the Waulsortian bioherms get larger and more equidimensional to the south before they disappear.

W. J. Meyers' (1974, 1975) reports on the Mississippian stratigraphy and diagenesis are based on petrographic and cathodo-luminescence studies of the carbonate sediments, cementation, and chert. He demonstrated that the nonferroan calcite cement zones in the Lake Valley Formation reflect ancient phreatic lenses established during pre-Viséan (pre-Meramecian) and pre-Bashkirian (pre-Morrowan) periods, when meteoric waters cemented these rocks below unconformities.

D. A. Yurewicz' (1975) investigation of the basin-margin sedimentary rocks of the Viséan and Namurian (Meramecian) Rancheria Formation in

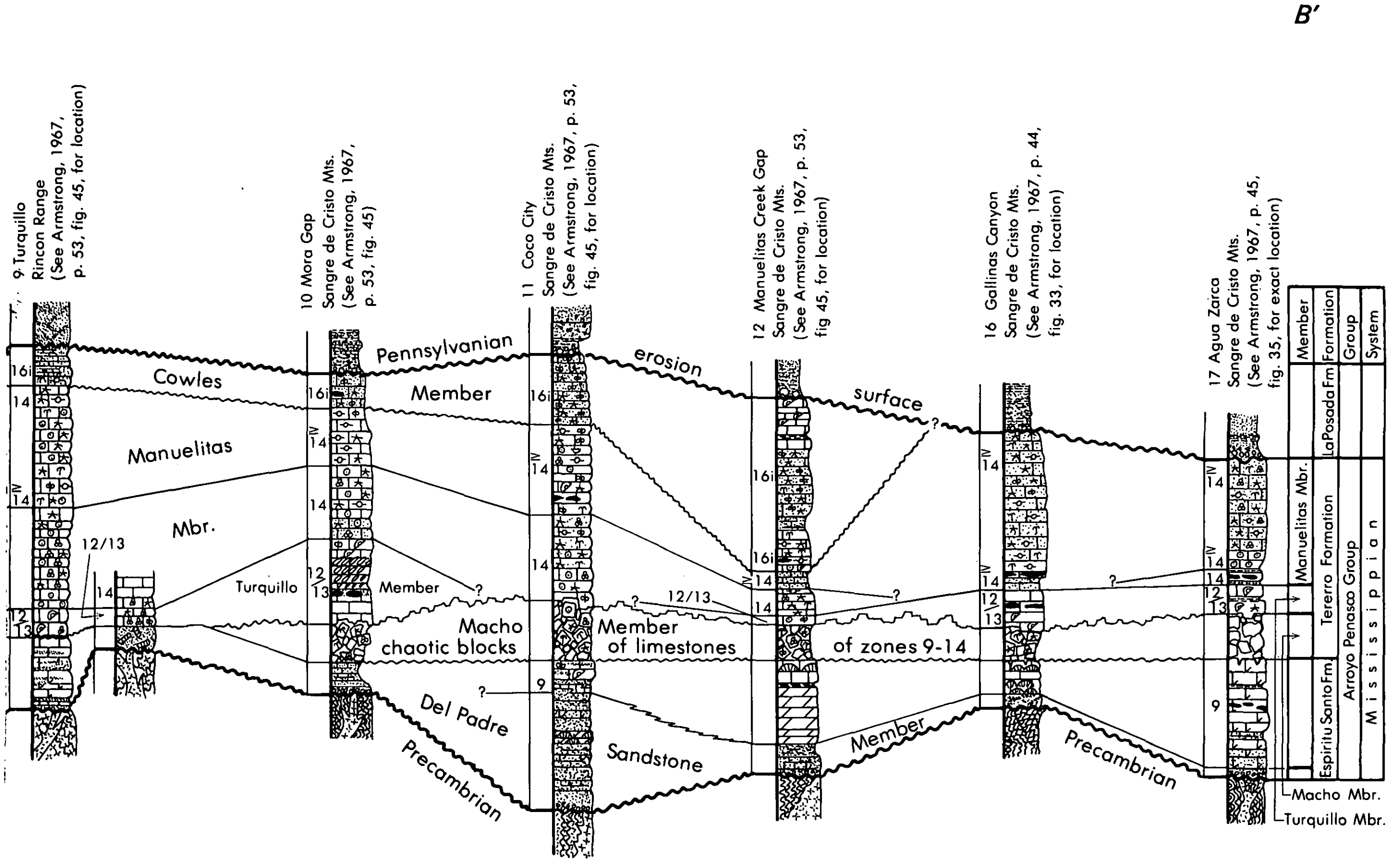


FIGURE 4.—Continued.

the Sacramento Mountains shows that the Rancheria Formation is younger than the Lake Valley Formation (Tournaisian (Osagean)) and is separated from it by an unconformity.

H. Richard Lane's (1974, fig. 4; 1975, fig. 3) study of the conodont faunas of the Lake Valley, Rancheria, and Helms Formations conclusively demonstrates that the Viséan and Namurian Rancheria Formation of the Franklin Mountains of east Texas and the southern San Andres and Sacramento Mountains of New Mexico has a wedge-on-wedge relation with the Tournaisian (Osagean) shelf carbonate rocks and bioherms of the Lake Valley Formation; the Rancheria and Lake Valley wedges are separated by an unconformity. The Viséan and Namurian (Zones 14–19, Meramecian–Chesterian) Rancheria and Helms Formations of western Texas and the Florida Mountains, N. Mex., also have a wedge-on-wedge relation with the Tournaisian Keating Formation of southwestern New Mexico (fig. 5) and are separated from the Keating by an unconformity.

LOCATIONS OF MISSISSIPPIAN OUTCROPS

Locations of Mississippian outcrop sections used in this report are shown on figure 1 and are described in the publications listed as follows, by section numbers:

1–20. Armstrong (1967), Armstrong and Mamet (1974), Baltz and Read (1960); The areas are the San Pedro, Nacimiento, Sangre de Cristo, Sandia, Manzano, and Jemez Mountains, N. Mex.

21–24. Armstrong (1958b); Ladron, Magdalena, Lemitar Mountains, and Coyote Hills, N. Mex.

25–33. Laudon and Bowsher (1949), Kottlowski and others (1956), Kottlowski (1975b); San Andres Mountains, N. Mex.

34–36. Laudon and Bowsher (1941, 1949), Pray (1958, 1961), Meyers (1973, 1974, 1975), Lane (1974, 1975), Yurewicz (1973, 1975); Sacramento Mountains, N. Mex.

37–48. Laudon and Bowsher (1949); Mimbres and Cooks Ranges, Silver City area, New Mexico.

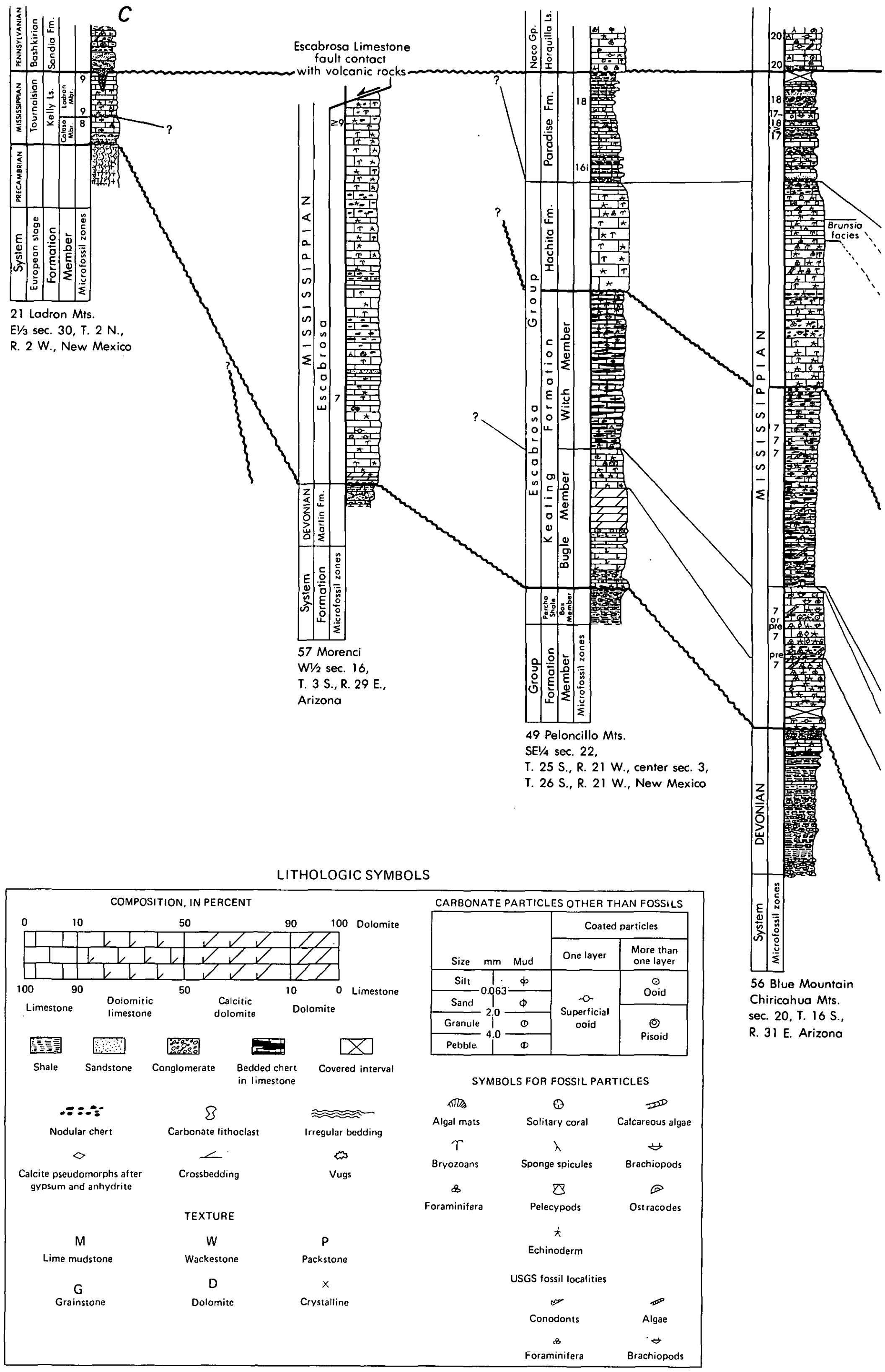
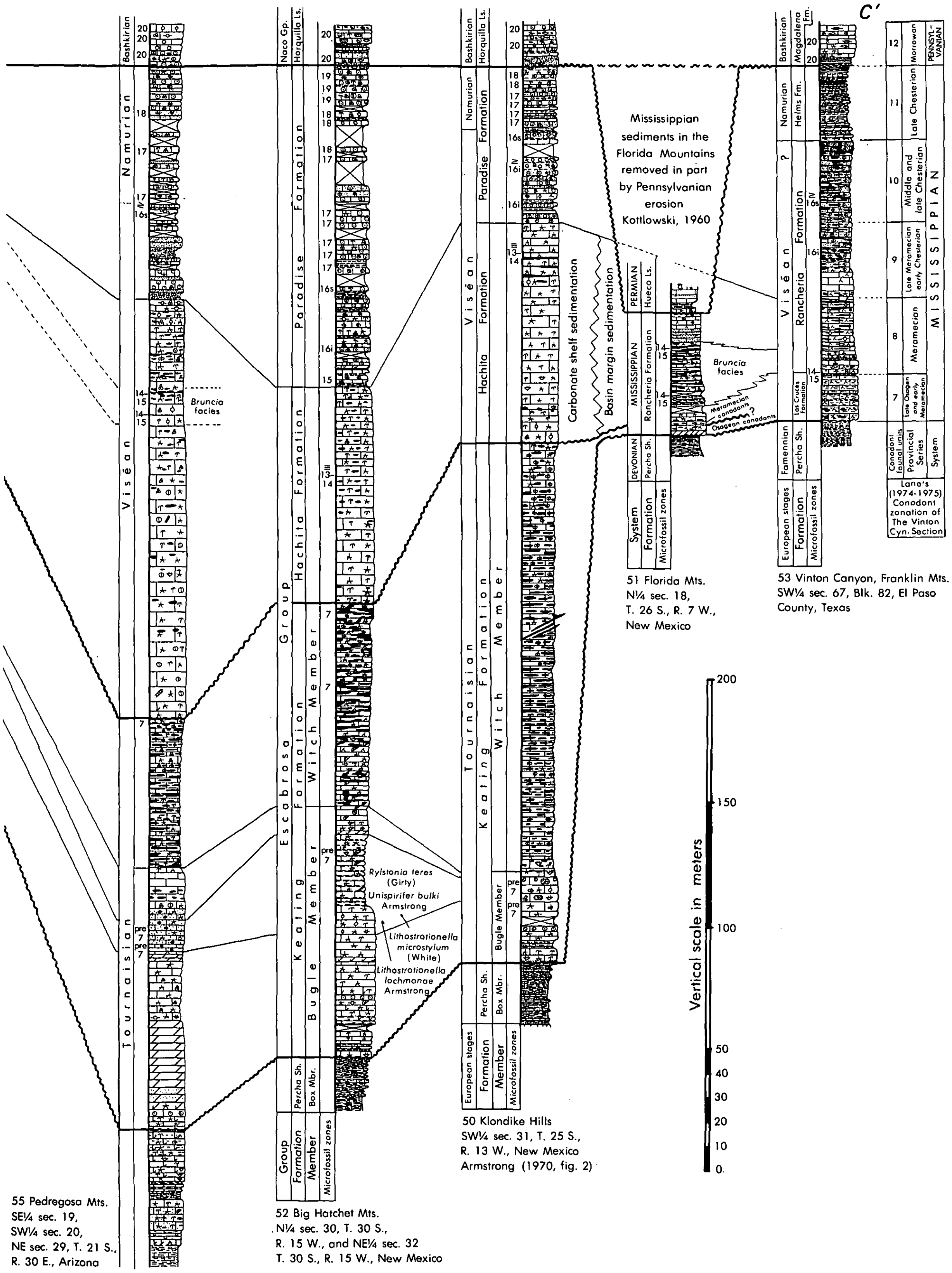


FIGURE 5.—Regional biostratigraphic and lithologic correlation for Mississippian strata along
tains, southeastern Arizona, the Peloncillo Mountains, New Mexico, the Pedregosa Moun-
ern New Mexico, to the Franklin Mountains and Vinton Canyon, western Texas. Line of



line C–C′ from the Ladron Mountains, west-central New Mexico, to Clifton, Arizona, the Chiricahua Mountains, southeastern Arizona, the Big Hatchet Mountains, Klondike Hills, and Florida Mountains, southwest- section is shown in figure 1.

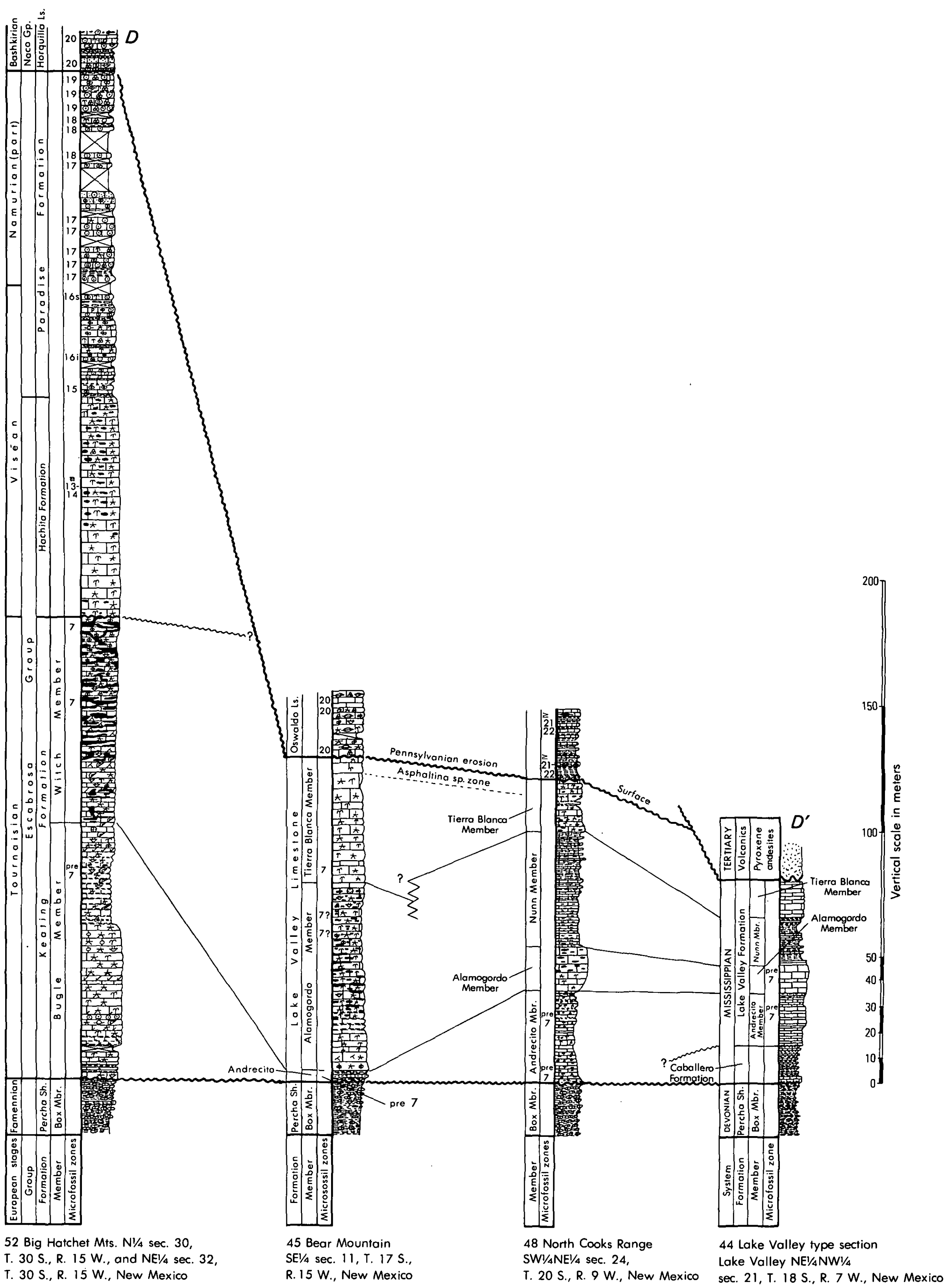


FIGURE 6.—Regional biostratigraphic and lithologic correlation of Mississippian strata along line *D–D′* from the Big Hatchet Mountains to Bear Mountain and the Mimbres Range, southwestern New Mexico. Line of section is shown in figure 1; symbols are explained in figure 5.

49–52. Armstrong (1962, 1970); Peloncillo, Big Hatchet, and Florida Mountains, Klondike Hills, N. Mex.
53, 54. Laudon and Bowsher (1949), Lane (1974, 1975); Vinton Canyon, Franklin Mountains, and Hueco Mountains, west Texas.
55. Epis (1956); Pedregosa Mountains, Ariz.
56. Sabins (1957); Chiricahua Mountains, Ariz.
57. Lindgren (1905); Clifton-Morenci district, Arizona.
58, 59. Armstrong and Mamet (1976); San Juan Mountains, Colo.

PENNSYLVANIAN SYSTEM

Pennsylvanian rocks in New Mexico represent a complete section in several areas, particularly in the southwestern and southeastern parts of the State, where a continuous section contains Morrowan to Virgilian equivalents. Rocks of this system are present throughout the State, except where they have been removed by erosion since the Pennsylvanian or where they were not deposited on the Pennsylvanian-age uplifts or preexisting highs. These major uplifts (fig. 7) were the Uncompahgre Uplift, in the north-central part of the State; the Sierra Grande, in the northeastern area; the Pedernal Uplift, extending through the central part of the State; the Zuni Uplift, in the west-central area; the Matador Arch, in the east-central area; and the Florida Uplift, in the southwestern part of the State.

The major outcrops of Pennsylvanian rocks are (1) in the Sangre de Cristo Mountains and Nacimiento Mountains, in the north-central part of the State; (2) in the Sandia, Manzano, Ladron, and Los Pinos Mountains and Lucero Mesa area, in central New Mexico; (3) in the Oscura, San Andres, Sacramento, Caballo, Fra Cristobol, and Robledo Mountains, of south-central New Mexico; and (4) in the Black Range, on Cookes Peak and Lone Mountain, in Silver City, and in the Peloncillo, Animas, and Big Hatchet Mountains, of southwestern New Mexico. The outcrops in southwestern New Mexico are in basin-and-range fault blocks.

Thick subsurface sections of Pennsylvanian rocks are present in the San Juan Basin of northwestern New Mexico, in many graben valleys of the southwestern and south-central part of the State, and in the subsurface of the western part of the Permian Basin in southeastern New Mexico and western Texas in the areas of the Delaware Basin and the northwest shelf. Sections of Pennsylvanian rocks are thickest (1) in the Sangre de Cristo Mountains, where they probably exceed 2,300 m in thickness (Baltz, 1972; Read and Wood, 1947; Sutherland, 1963), (2) in the Big Hatchet Mountains area of southwestern New Mexico and in the flanking Pedregosa Basin, where thicknesses are greater than 750 m, (3) in the Orogrande Basin of south-central New Mexico, where they are thicker than 900 m, and (4) in the Delaware Basin of southeastern New Mexico and western Texas, where more than 600 m of Pennsylvanian rocks are present.

Uplifting of mountain ranges in Cenozoic time, particularly those bordering the Rio Grande rift and those in the basin-and-range country of south-central and southwestern New Mexico, resulted in exposure of many spectacular sections of Pennsylvanian rocks, particularly in the Sangre de Cristo, Sandia-Manzano, San Andres, Sacramento, Caballo, and Big Hatchet Mountains. Owing to the semiarid climate of the southern part of the State, Pennsylvanian limestones are well preserved in prominent peaks such as the Big Hatchet Mountains in southwestern New Mexico and the Oscura, Caballo, and San Andres Mountains of the south-central region.

HISTORY

Jules Marcou (1856), J. S. Newberry (1876), and others recognized Carboniferous rocks in early reconnaissance surveys. Stratigraphic sections and lists of Carboniferous fossils from the north-central areas were published by J. J. Stevenson (1881). Keyes (1906) summarized the Carboniferous sections in south-central areas. Gordon (1907) named the Magdalena Formation from outcrops in the Magdalena Mountains, and Lee (1909) and Darton (1928) briefly described Pennsylvanian rocks in various areas of New Mexico. The Carboniferous rocks in the Silver City mining area were described by Paige (1916) and Spencer and Paige (1935). Many of the early descriptions of Carboniferous rocks were from the mining districts in the southwestern, central, and north-central parts of the State.

Exploration for oil and gas, spurred by World War II, led to many reports in the 1940's and essentially began modern studies of the Pennsylvanian in the State (table 1). Expansion of the work by the New Mexico Bureau of Mines and Mineral Resources and by the U.S. Geological Survey and many projects by university professors and their students, many of them supported by the New Mexico Bureau of Mines and Mineral Resources, led to a rapid

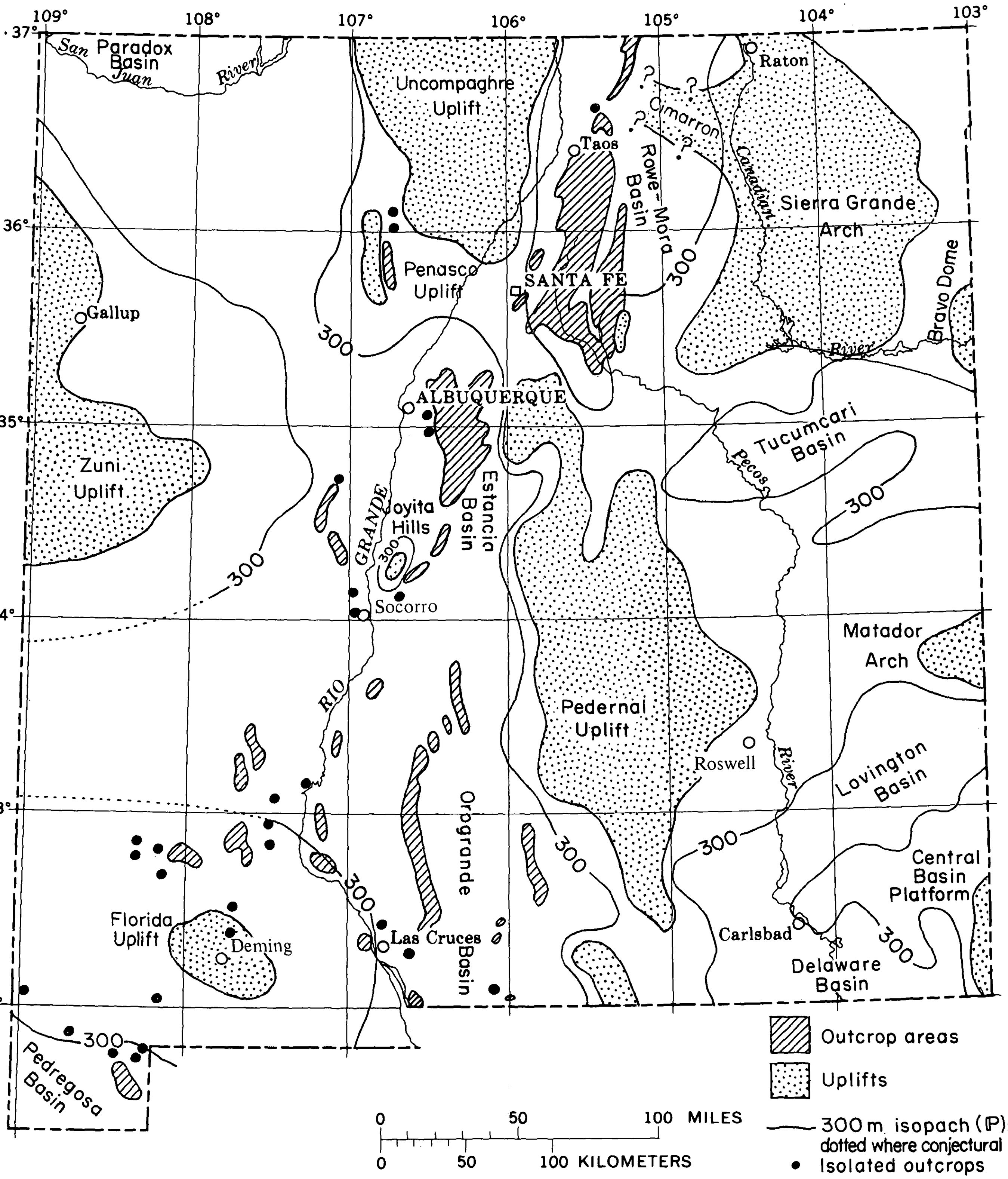


FIGURE 7.—Map of New Mexico showing areas where Mississippian and Pennsylvanian rocks (1) crop out, (2) are more than 300 m thick, and (3) are absent. Positive areas are labeled.

TABLE 1.—*Reports on the Pennsylvanian deposits of New Mexico, 1937–1978*

Year	Author(s)	Area or subject
	1930's	
1937	Needham	Fusulinids.
	1940's	
1940	Needham	Fusulinids.
1942	Thompson	Statewide.
1942	Loughlin and Koschmann.	Type Magdalena Formation.
1944	Henbest and Read	Sierra Nacimiento.
1944	Read	North-central New Mexico.
1946	Northrop	Southeastern Sangre de Cristo Mountains.
1946	Henbest	Central New Mexico.
1946	Wilpolt, MacAlpine, Bates, and Vorbe.	Do.
1946	Kelley and Wood	Lucero uplift.
1946	Stark and Dapples	Los Pinos Mountains.
1947	Read and Wood	Northern New Mexico.
1948	Thompson	Southern New Mexico.
1949	Lloyd	Southeastern New Mexico.
	1950's	
1950	Bradish and Mills	San Juan Basin.
1951	Wilpolt and Wanek	Central New Mexico.
1952	Brill	North-central New Mexico.
1952	Borden	San Juan basin.
1952	Kelley and Silver	Caballo Mountains.
1953	Kottlowski	Central New Mexico.
1953	Bachman	Mora County.
1953	Plumley and Graves	Virgilian reefs.
1954	Pray	Sacramento Mountains.
1954	Kuellmer	Black Range.
1955	Thompson and Kottlowski.	South-central New Mexico.
1956	Baltz and Bachman	Southeastern Sangre de Cristo Mountains.
1956	Kottlowski, Flower, Thompson, and Foster.	San Andres Mountains.
1957	Herman and Barkell	Paradox basin.
1958	Wengerd and Matheny.	Four Corners region.
1958	Kottlowski	Southwestern New Mexico.
1958	Galley	Southeastern New Mexico.
1958	Hardie	Northern Hueco Mountains.
1958	Gillerman	Peloncillo Mountains.
1958	Gehrig	Brachiopods.
1958	Bachman and Hayes	Sacramento Mountains.
1959	Cline	Do.
1959	Kottlowski	West-central New Mexico.
1959	Oppel	Sacramento Mountains.
1959	Otte	Do.
1959	Wengerd	Northwestern New Mexico.
	1960's	
1960	Kottlowski	Southwestern New Mexico.
1961	Pray	Sacramento Mountains.
1961	Northrop	Paleontology.
1961	Kottlowski	North-central New Mexico.
1962	Adams	Eastern New Mexico.
1962	Wray	Algal banks.
1962	Kottlowski	Southwestern New Mexico.
1962	Wengerd	Northwestern New Mexico.
1963	Sutherland	South Sangre de Cristo Mountains.
1963a	Kottlowski	South-central New Mexico.
1963b	Kottlowski	Socorro County.
1963c	Kottlowski	Sante Fe area.
1965	Baltz	Raton Basin.
1965	Kottlowski	Southwestern New Mexico.
1965	Zeller	Big Hatchet Mountains.
1966	Meyer	Southeastern New Mexico.
1967	Sutherland and Harlow.	Brachiopods.
1967	Wilson	Sacramento Mountains.
1969	Kottlowski	South-central New Mexico.
1969	Wilson	Southwestern New Mexico.
1969	Wilson, Madrid-Soils, and Malpica-Cruz.	Do.
	1970's	
1970	Kottlowski and Stewart.	Joyita area; fusulinids.
1971	Martin	Lucero Mesa.
1972	Baltz	Gallinas Creek area.
1972	Clark and Read	Eagle Nest area.
1972	Foster, Frentress, and Riese.	East-central New Mexico.
1972	Harbour	Franklin Mountains.
1972	Hills	Southeastern New Mexico.
1972	Wilson	Sacramento Mountains.
1972	Sutherland	Sangre de Cristo Mountains.
1973	King	Fusulinids.
1973	Myers	Manzano Mountains.
1973	Siemers	Socorro County.
1974	DuChene	North-central New Mexico.
1974	LeMone, King, and Cunningham.	Silver City Range.
1974	Northrop	Paleontology.
1975	Bachman	Entire State.
1975	Bachman and Myers	South-central New Mexico.
1975a, b	Kottlowski	San Andres Mountains.
1975	Thompson and Bieberman.	Doña Ana County.
1975	Zidek	Fossil fish.

TABLE 1.—*Reports on the Pennsylvanian deposits of New Mexico, 1937–1978*—Continued

Year	Author(s)	Area or subject
	1970's	
1976	Roberts, Barnes, and Wacker.	Northeastern New Mexico.
1977	Greenwood, Kottlowski, and Thompson.	Pedregosa Basin.
1978	Siemers	West-central New Mexico.

increase in the number of reports in the 1950's and 1960's (table 1). In the 1970's some notable regional papers and many detailed geologic maps showing areas of Pennsylvanian outcrops have been published. Some of these are listed in table 1.

The early regional papers were by Stevenson (1881), Gordon (1907), Lee (1909), and Darton (1928). More recent regional reports were those of Meyer (1966) and Galley (1958) on southeastern New Mexico, Wengerd (1962) on the San Juan basin area, Kottlowski (1960, 1962) on south-central and southwestern New Mexico, Read and Wood (1947) on north-central New Mexico, Baltz (1965) on the Raton Basin area, and Bachman (1975) as a summary of the Pennsylvanian in the State.

GEOLOGIC SETTING

UNDERLYING ROCKS

Mississippian rocks immediately underlie Pennsylvanian strata throughout New Mexico except along the flanks and on the crest of uplifts present in Pennsylvanian time. Along the flanks of the uplifts, particularly in south-central New Mexico, Pennsylvanian rocks lap northward onto Devonian, Silurian, Ordovician, or Cambrian units. The system rests on Precambrian rocks on the Sierra Grande Arch, Uncompahgre Uplift, Peñasco Uplift, Zuni Uplift, Joyita Hills, Pedernal Uplift, Florida Uplift area, Central Basin Platform, Matador Arch, and Bravo Dome.

NATURE OF CONTACT WITH UNDERLYING ROCKS

In many parts of southern New Mexico, oldest Pennsylvanian (Morrowan) rocks rest upon youngest Mississippian (Chesterian), and no hiatus is evident, although an erosion surface of low relief is seen locally. In the central and northern regions, the Morrowan laps out so that Atokan strata rest unconformably on older rocks. Moreover, the Chesterian is absent, adding to the missing section.

The erosional unconformity is evidenced by local channeling and slight reworking of Mississippian residual chert fragments into the lowest clastic strata of the Pennsylvanian. In some areas, such as the San Andres Mountains, Pennsylvanian beds were deposited on a pronounced erosional surface; in places, this surface cuts through the entire Mississippian section and the basal channel fills of the Pennsylvanian are chiefly chert-pebble conglomerates. In most areas where detailed mapping has been done, the basal clastic phase of the Pennsylvanian varies greatly in thickness from place to place and is almost entirely absent above low-relief hills of pre-Pennsylvanian rocks.

OVERLYING ROCKS

Pennsylvanian rocks are overlain by Permian (Wolfcampian) strata throughout the State except along the crest of some of the Laramide and Cenozoic uplifts where post-Paleozoic erosion has stripped off the overlying Permian deposits, and Tertiary volcanic rocks rest unconformably on the Pennsylvanian.

NATURE OF CONTACT WITH OVERLYING ROCKS

The Pennsylvanian-Permian contact in many areas appears to be conformable. In these areas, the boundary is somewhat arbitrarily drawn within a few tens of meters between the highest Virgilian and the lowest Wolfcampian fossil control. Permian rocks at some localities, as well as Cretaceous and Tertiary rocks in other places, are erosionally unconformable on the Pennsylvanian strata.

Along the edges of the uplifts that persisted into Permian time, such as the Pedernal and Uncompahgre Uplifts, Pennsylvanian beds were overlapped by Permian strata or were removed by erosion prior to Permian deposition. Upper Pennsylvanian rocks probably never were deposited on the crests of these uplifts. Pennsylvanian rocks may never have been deposited in parts of the Florida Uplift in southwestern New Mexico. Erosion during early and middle Mesozoic time stripped pre-Mesozoic rocks from the crest of the Burro Uplift in that area; on the northeast flank, Cretaceous beds rest with erosional unconformity on Pennsylvanian strata. Likewise, during the Laramide deformation in southwestern New Mexico, erosion stripped the crests of the highs so that Tertiary volcanic rocks rest unconformably on upper Paleozoic rocks in some localities.

In the north-central part of the State in the Sangre de Cristo Mountains area, Pennsylvanian rocks are overlain by the Sangre de Cristo Formation of Late Pennsylvanian and early Wolfcampian age. The nature of the contact is problematic at places because nonmarine arkoses and red beds of the Sangre de Cristo Formation transgress the time boundary and intertongue with the underlying fossiliferous Upper Pennsylvanian mixed-marine-nonmarine sedimentary rocks. Throughout most of the northern half of New Mexico, the Wolfcampian Abo red beds or time-equivalent continental facies overlie Pennsylvanian strata. In some central areas, a gradational sequence of interbedded limestones and red beds at the base of the Permian rests conformably on Virgilian rocks. In south-central and southeastern New Mexico, marine shelf carbonate rocks of the Hueco Formation overlie Pennsylvanian rocks conformably and in some places with an abrupt unconformity, as in the Hueco Mountains area. In southwestern New Mexico in the Big Hatchet Mountains area, the systemic boundary is within the upper part of the thick Horquilla Formation, a monotonous marine limestone sequence containing some shale.

STRUCTURAL EVENTS DURING PENNSYLVANIAN TIME

Positive and negative elements active in pre-Pennsylvanian time (fig. 1) were modified during the Pennsylvanian. The regional depositional surface in early Pennsylvanian time was tilted down to the south, and a general northward thinning resulted from the overlap of that surface by Pennsylvanian rocks. More localized elements that formed during the Pennsylvanian appear, in contrast, to trend roughly north, for example, the Pedernal, Sierra Grande, Uncompahgre, Joyita, and Penasco Uplifts (fig. 7). However, the Bravo Dome and Matador Arch on the east side of the State are east-trending uplifted features, whereas Central Basin Platform is essentially aligned north-south. The Zuni, Florida, and Cimarron Uplifts had northwest trends. In the Sangre de Cristo Mountains area, more than 2,300 m of Pennsylvanian clastic sediments were deposited in a deep north-trending structural basin called the Rowe-Mora Basin by Read and Wood (1947) and the Taos Trough by Sutherland (1963). A southeastern extension of the Paradox Basin into northwestern New Mexico during Pennsylvanian time was connected southeastward with a chain of small basins that connected southward with the Orogrande Basin; the Orogrande Basin occupied an area along the west side of the Pedernal Uplift and extended southward into Texas. The Delaware Basin in the southeast, the Pedregosa Basin in the southwest, and the Orogrande Basin in south-central New Mexico were areas of thick deposition throughout Pennsylvanian time.

STRUCTURAL EVENTS FOLLOWING PENNSYLVANIAN DEPOSITION

In the Rowe-Mora Basin area of north-central New Mexico, a thick sequence of nonmarine red beds and arkoses of the Sangre de Cristo Formation was deposited in late Pennsylvanian and Wolfcampian time indicating uplift and erosion of the bounding uplifts. Similarly, as part of the Rocky Mountain orogeny, in south-central New Mexico, the Pedernal Uplift was active in early Permian; it remained as a highland in local areas until middle Permian time, as shown by areas such as Pedernal Mountain and Pajarito Mountain, where Precambrian rocks are overlapped by Abo red beds, the Yeso Formation, and the San Andres Limestone of Leonardian and early Guadalupian age. Laramide structural deformation greatly affected the State; overthrust belts formed in southwestern New Mexico. Present distribution of Pennsylvanian rocks has been greatly affected by basin-and-range faulting during Miocene to Holocene time. Much of this late Cretaceous-early Tertiary and middle and late Cenozoic structural deformation appears to have been controlled somewhat by the features of Pennsylvanian and early Permian age. For example, the Rio Grande rift is subparallel to the string of central New Mexico Pennsylvanian-age depositional basins that ran essentially from the southeast corner of the San Juan Basin area southward into south-central New Mexico.

LITHOSTRATIGRAPHY

LITHOSTRATIGRAPHIC SUBDIVISIONS

Gordon (1907) proposed the term Magdalena Group for all the sedimentary rocks in central New Mexico above the Mississippian and below the Abo red beds, which are the basal Permian. The unit in places is synonymous with Pennsylvanian, but in other areas, it has been used to include Mississippian rocks at the base and (or) Wolfcampian rocks at the top.

Gordon divided the Magdalena Group into a lower clastic phase, the Sandia Formation, and an upper carbonate phase, the Madera Limestone. In many areas, the Madera was subdivided into a lower limestone member and an upper arkosic limestone member. In addition, at the top of the Magdalena Group in central New Mexico, is the Bursum Formation, which in most areas is entirely of Wolfcampian age.

In the Silver City mining district of southwestern New Mexico, Spencer and Paige (1935) used the term Magdalena Group; however, they divided the group into a lower Oswaldo Formation and an upper Syrena Formation, which were roughly comparable to the Sandia and Madera units. Thompson (1942, 1948) divided the Pennsylvanian into correlatives of the Des Moines, Missouri, and Virgil Series and introduced a new series, Derry, for the pre-Des Moines strata in the southern part of the State. He subdivided each series into two groups and then into thin formations on the basis of rock types and fusulinid zonation.

Geologic mapping of isolated mountain ranges in New Mexico has led to a nomenclature being used in each range that is somewhat distinct from the general classification. The terms "Sandia" and "Madera" have been used in north-central New Mexico in the Sangre de Cristo Mountains, although in the vicinity of Pecos, Sutherland (1963) combined rocks equivalent to the Sandia and the lower member of the Madera into the La Pasada Formation. Because to the north, facies of this unit are different, he applied the name Flechado Formation to the rocks generally equivalent to the La Pasada to the south. The upper member of the Madera was called the Alamitos Formation by Sutherland, and the uppermost Pennsylvanian rocks are in the lower part of the Sangre de Cristo Formation in this area. In the San Juan Basin region, where essentially all the Pennsylvanian units are in the subsurface in New Mexico but crop out north of Durango, Colo., on the edge of the San Juan Mountains, the Atokan part of the Pennsylvanian is the Molas Formation and is overlain by the Hermosa Group (fig. 8). The Hermosa Group consists of, in ascending order, the Pinkerton Trail Formation, the Paradox Formation, and the Honaker Trail Formation.

In the Lucero Mesa area southwest of Albuquerque, Kelley and Wood (1946) mapped the Sandia Formation and Madera Limestone and divided the

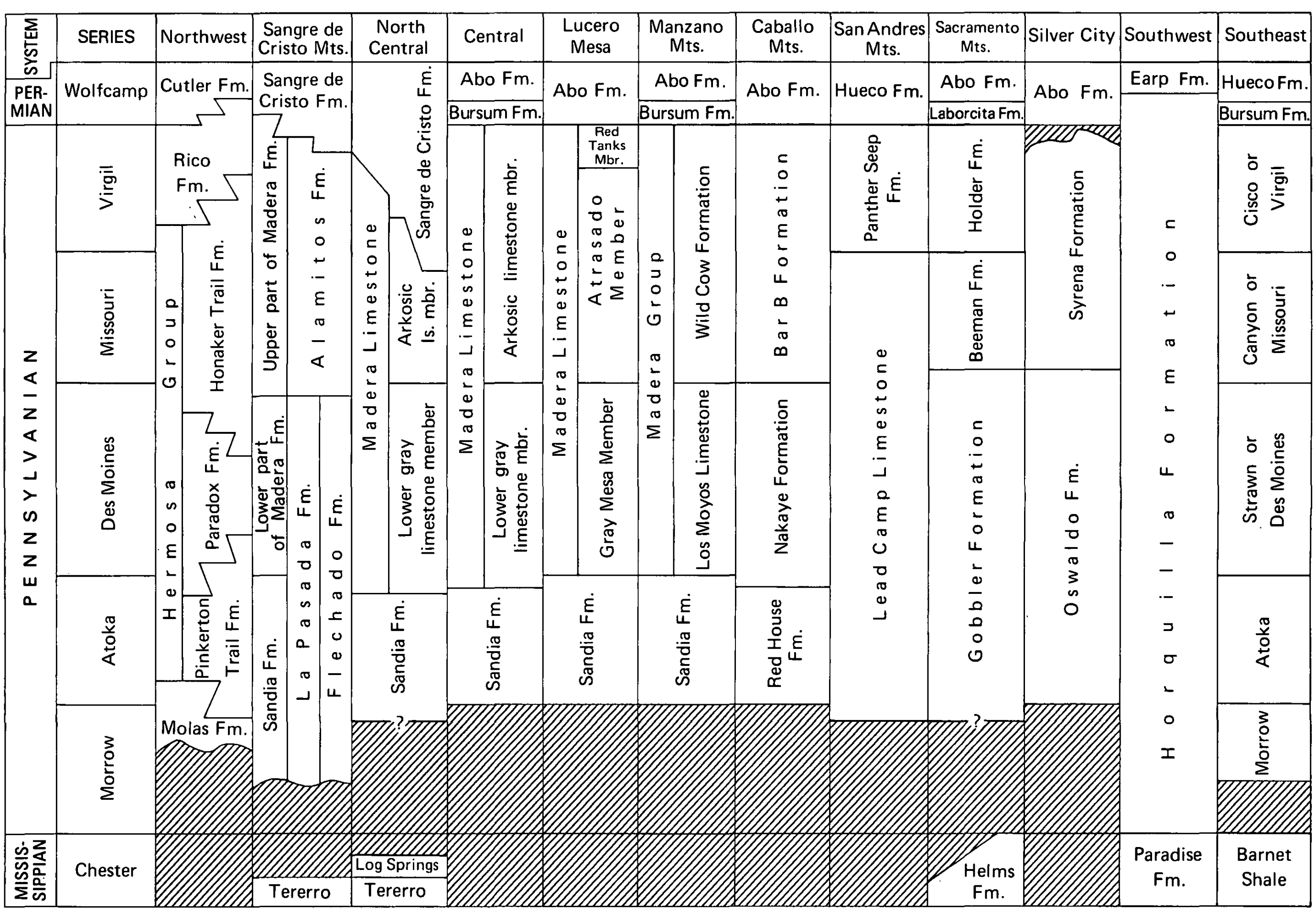


FIGURE 8.—Correlation chart of Pennsylvanian rocks of New Mexico.

Madera into the Gray Mesa Member at the base, the middle Atrasado Member, and the upper Red Tanks Member. Myers (1973) recently mapped in the Manzano Mountains area, southeast of Albuquerque; he retained the use of the Sandia Formation and raised the Madera to a group that includes at its base the Los Moyos Limestone and at its top, the Wild Cow Formation. He divided the Wild Cow into three members. Kelley and Silver (1952), in their mapping of the Caballo Mountains in south-central New Mexico, used essentially the threefold division of Sandia, Madera Limestone, and upper clastic units of the Madera. However, they named their map units, in ascending order, the Red House, Nakaye, and Bar B Formations. In the Sacramento Mountains area of south-central New Mexico, Pray (1954, 1961) divided the Pennsylvanian into the Gobbler, Beeman, and Holder Formations, in ascending order. In general, the Gobbler Formation is correlative with the Sandia and the lower part of the Madera Limestone, and the Beeman and Holder Formations are correlative with the upper arkosic limestone member of the Madera.

In the subsurface in the Delaware Basin area and adjoining marine shelves, lithic subdivisions have not generally been used; the strata have been referred mainly to series by utilizing the Midcontinent (or Texas) series names and thus have been labeled as rocks of Morrowan, Atokan (or Bend), Des Moinesian (or Strawn), Missourian (or Canyon), and Virgilian (or Cisco) age. In southwesternmost New Mexico, in the Big Hatchet, Animas, and Peloncillo Mountains, the Pennsylvanian has been referred to the Horquilla Formation, which in that area also includes Wolfcampian beds in its upper part. In the San Andres Mountains area of south-central New Mexico, the lower part of the Pennsylvanian has been called the Lead Camp Limestone of Bachman and Myers (1975), which includes equivalents of the Sandia Formation and the lower part of the Madera Limestone; the upper units are mapped in the Panther Seep Formation.

PRINCIPAL ROCK TYPES

Generally speaking, the Pennsylvanian sequence consists of a lower clastic unit roughly 150 m thick, a middle limestone unit 200–300 m thick, and an upper 150-m-thick unit of interbedded limestone and shale containing red beds near the top of the Pennsylvanian. In the Sangre de Cristo Mountains, the basal Sandia Formation consists mainly of shale, siltstone, and fine-grained to very coarse grained sandstone in varying proportions. In most of the area, shale beds form the greatest part of the Sandia, and most of these shales, even those containing marine fossils, are carbonaceous. Some coal beds and thin marine limestones are minor constituents; in the eastern part of the area, coarse arkosic sandstones are common in the Sandia Formation. The lower member of the Madera Formation in the southern and southeastern part of the mountains is characterized by light-gray marine limestones, many of them biostromal or biohermal, and interbedded calcareous shale and some thin sandstones. Northward in the Sangre de Cristo Mountains, limestones become subordinate, and thick arkosic sandstones and thick calcareous shales become major constituents. The upper part of the Madera Formation and the Alamitos Formation consist of varying proportions of red, greenish-gray, and gray shale, coarse-grained arkosic sandstone, green limestone, and nodular limestone.

Outcrops in the central part of the State in the Sandia, Manzano, and Los Pinos Mountains, Lucero Mesa, and Magdalena Mountains show a typical sequence of a lower clastic unit, middle limestone unit, and upper interbedded limestone and clastic unit. A similar sequence is seen in the southern and south-central part of the State, in the Caballo, San Andres, Sacramento, and Robledo Mountains as well as in the northern Franklin Mountains. However, the upper clastic unit, the Panther Seep Formation in the San Andres Mountains and the Beeman and Holder Formations in the Sacramento Mountains, is as much as 750 m thick. This upper clastic unit was deposited in or near the Orogrande Basin on the west side of the Pedernal Uplift and is lithologically distinct from beds typical of this part of the section in surrounding parts of New Mexico. Deposits are deltaic to brackish-water clastic rocks and precipitates, all deposited in relatively shallow waters; they include silty brownish shales, dark carbonaceous shales, dark-gray argillaceous limestones, laminated calcilutites, silty calcarenites, silty calcareous sandstone, thick lenses of massive biostromal limestone, and numerous biohermal reefs. Two thick gypsum beds are near the top of this sequence in south-central New Mexico.

In the southwestern panhandle of New Mexico, especially in the Big Hatchet Mountains area, the Pennsylvanian rocks are included with an overlying conformable unit of lower Wolfcampian age in the Horquilla Limestone and are conformable on Chesterian strata. This formation is about 700 m thick; the shelf facies consists dominantly of limestone; some interbedded siltstones are in the lower part. Along the margin of the Alamo Hueco-Pedregosa Basin (Zeller, 1965) to the southwest, porous dolostones are interlayered with limestones. In the deep-marine-basin facies, argillaceous limestone and mudstone are dominant.

FACIES CHANGES

The Pennsylvanian sections of central New Mexico were the first to be studied by geologists because they are near the populous Rio Grande Valley and also are near some of the early mining districts. In this limited area, the Pennsylvanian appeared to be relatively uniform. However, when Read and Wood (1947) and Thompson (1948) traversed from Pennsylvanian uplifts into the large basins, abrupt facies changes became obvious.

In the Sangre de Cristo Mountains, the Sandia Formation is a suite of mixed marine and nonmarine sedimentary rocks. In the southern part of the area, the Sandia is relatively thin but thickens greatly northward into the Rowe-Mora Basin; the proportion of black shale increases markedly from shelf to basin. In the southern part of the Sangre de Cristo Mountains, the lower member of the Madera Formation is mainly biohermal and biostromal limestone and contains lesser amounts of gray shale and thin sandstone, indicating a shallow-marine environment. Northward, the equivalent rocks grade into a facies in which dark-gray shale predominates and in which arkoses become conspicuous constituents. In the upper member of the Madera and the equivalent Alamitos Formation are mixed-marine and nonmarine deposits; marine limestone and shale are the predominant constituents in the south, but thick arkoses and red beds are predominant in the north.

In the southwestern panhandle, outcrops in the Big Hatchet Mountains are predominantly shallow-marine shelf limestone. Along the shelf margin, porous dolostone is interbedded with the limestone. To the south in the Alamo Hueco Basin, the facies changes into deep-marine basinal limestone and mudstone.

In south-central New Mexico, the lower part of the Pennsylvanian sequence in the Sacramento Mountains, on the west flank of the Pedernal Uplift, is similar to that in the San Andres Mountains 65 km to the west on the west side of the Orogrande Basin. However, the overlying strata of Missourian and Virgilian age in the Sacramento Mountains are feldspathic sandstone, limestone, and shale, and discontinuous algae reefs and upper reddish marl, nodular limestone, and whitish massive limestone. This facies contrasts greatly with the Orogrande Basin deposits in the Panther Seep Formation of the San Andres Mountains. These deposits include many types of fine-grained shales, siltstones, and sandstones as described above. On the edge of the Pedernal Uplift in the northern Sacramento Mountains, Otte (1959) determined the order of facies, from west to east (toward the landmass), to be: (1) massive marine limestone, (2) nodular argillaceous fusulinid-bearing limestone, (3) silty limestone containing shallow-marine fossils such as mollusks and brachiopods, (4) dolomitic limestone, (5) green calcareous shale, and (6) marine to nonmarine red shale and other terrigenous clastic rocks.

Similar facies changes exist on the east side of the Pedernal Mountains from the shelf into the Delaware Basin area. Pennsylvanian rocks do not crop out in southeastern New Mexico but have been explored by thousands of oil tests in that area.

ENVIRONMENTS OF DEPOSITION

Throughout New Mexico, depositional environments during Pennsylvanian time were strongly influenced by their tectonic positions relative to the subsiding basins and rising uplifts. The major subsiding elements were the Paradox Basin, the Rowe-Mora Basin, a central New Mexico basinal area, the Orogrande Basin, the Pedregosa Basin (Greenwood, Kottlowski, and Thompson, 1977), and the Delaware Basin. The depositional areas between the basins and the uplifts ranged from wide, tectonically stable shelves to narrow unstable belts where terrigenous deposits predominated and intertongued basinward with marine limestone and dark carboniferous basinal deposits.

In such a sedimentary framework, a wide variety of terrigenous and nonterrigenous sediments was deposited. Outcrops and subsurface data show examples of nearshore deposits including: (1) swamp and marsh deposits consisting of dark, organic-rich, carbonaceous, bioturbated claystone containing a few silt laminae and a little well-preserved plant debris; (2) tidal-flat sand and mud; (3) mud and sand indicative of overbank and channel deposition associated with small, nearshore deltaic complexes; and (4) bioturbated, muddy sediments containing sandy layers indicative of lagoonal conditions. Shelf environments are shown by carbonate wackestone and mudstone and terrigenous sandstone and mudstone. Dark carbonate mudstone, bedded with dark clay-shale admixed with carbonate deposits or forming thin distinct layers, also is present at some localities and reflects deeper basin sedimentation.

BIOSTRATIGRAPHY

AGE OF ROCKS

In all quadrants of the State, the standard Pennsylvanian series, Morrow, Atoka, Des Moines, Missouri, and Virgil, are present. However, in places, particularly along the flanks of uplifts, the entire Morrowan or its lower part is absent. In other areas, the uppermost part of the Virgilian is missing. The most complete section is in the Big Hatchet Mountains area of southwestern New Mexico, where all the Pennsylvanian series are present in a dominantly carbonate-rock sequence and where contacts with the underlying Chesterian and overlying Wolfcampian units are conformable.

FAUNAL SUCCESSION AND ASSIGNMENT OF SERIES

Marine rocks are dominant in the Pennsylvanian sequences in New Mexico, and marine invertebrate fossils are abundant throughout most of the sections. Fusulinids provide the main basis for the assignment of series boundaries and correlation of biostratigraphic units. In some areas, the brachiopod faunas have been described in detail, particularly in the Sangre de Cristo Mountains by Sutherland (1963) and Sutherland and Harlow (1973).

Morrowan rocks contain a distinctive fauna of *Eostaffella* and *Millerella* which differ from the same genera in the lower part of the Atokan. The Atokan series is divided into a lower zone on the range of the genus *Profusulinella* and into an upper zone on the range of *Fusulinella.* Des Moinesian rocks are marked by *Beedeina* and *Wedekindellina.* Missourian beds are marked by more primitive forms of the genus *Triticites* and by *Eowaeringella.* The biozone of *Eowaeringella* is one of the best defined and most restrictive fusulinid zones (Stewart, 1968, 1970) and is present near the base of the Missourian series. Virgilian rocks are distinguished by the ranges of certain species *Triticites,* by *Dunbarinella,* and by certain species of *Pseudofusulinella. Triticites* ranges up into the lower part of the overlying basal Permian (Wolfcampian) series; *Schwagerina, Pseudofusulina,* and *Leptotriticites* occur in the lower part of the Wolfcampian and extend up into younger Permian but do not occur in Virgilian rocks. Wilde (1975) has systematically studied the fusulinids in the Big Hatchet Mountains section and places the Pennsylvanian-Permian boundary on the basis of the evolutionary development of various species of *Triticites* and at the first appearance of *Leptotriticites* and (or) *Schwagerina.*

IGNEOUS AND METAMORPHIC ROCKS

No igneous rocks of Pennsylvanian age are known in New Mexico, although some of the Missourian shales in the area east of Socorro are bentonitic, indicating their possible derivation from volcanic ash. In many mining districts, Pennsylvanian rocks have been metamorphosed by Laramide- and Cenozoic-age intrusions and, along with the Mississippian rocks, are hosts for some of the base-metal ore deposits.

ECONOMIC PRODUCTS

COAL

Scattered lenses of coal are present in the lower part of the Pennsylvanian in the Sangre de Cristo Mountains. In general, these coal lenses are in the shoreline deposits on the east and west sides of the Rowe-Mora Basin and are in beds ranging in age from Atokan to Des Moinesian. Some of the coal lenses are as much as 2 m thick but are very local in extent. This coal was used only in homes and in small limestone kilns. Thin coal beds and coal laminae are present in Virgilian strata in the Sandia and Manzano Mountains; they are intercalated with red beds and marine limestone in shoreline deposits on the west side of the Pedernal Uplift.

Lenses of coal have been found in oil tests on the northwest shelf of the Delaware Basin amid the deltas and along the irregular coastlines of the shallow seas that bordered the Pedernal Uplift on the east and southeast during the Pennsylvanian. These coal lenses are associated with barrier-bar and point-bar stream-channel sandstones and with clay and silty flood-basin deposits. Most of these thin coal lenses (detected only by drilling) are Morrowan or Atokan in age. During early Pennsylvanian time, this area may have been dominated by large deltas similar to those of the Illinois Basin.

PETROLEUM RESOURCES

Prolific production of oil has been obtained from Pennsylvanian reservoirs in the Delaware Basin area of southeastern New Mexico and in the San Juan-Paradox Basin of northwestern New Mexico. Annual production from Pennsylvanian rocks reached its height of about 27 million barrels per year in 1969 and has dropped off since then to a little more than 7 million barrels in 1976. Cumulative production from the Pennsylvanian reservoirs in the State (fig. 9) is about 260 million barrels.

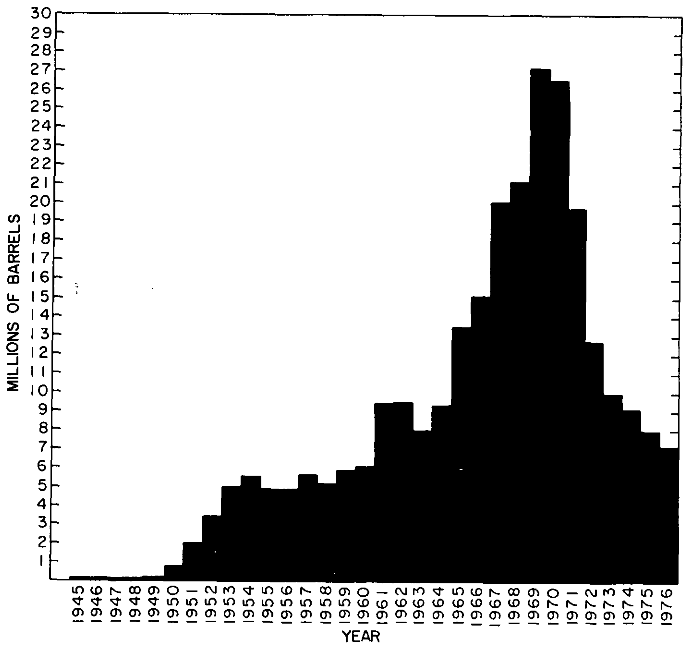


FIGURE 9.—Amount of crude oil produced from Pennsylvanian reservoirs in New Mexico, 1945–1976. Data from Roy W. Foster (written commun., 1978).

Production from Mississippian rocks is restricted mainly to southeastern New Mexico and the Delaware Basin area; peak production was in 1960 when 80,000 barrels was produced from Mississippian rocks. The last production of any significance from Mississippian strata was in 1972 when about 20,000 barrels was produced. The cumulative production from the Mississippian for New Mexico has been about 450,000 barrels.

Pennsylvanian rocks are a major source of natural gas in southeastern New Mexico. In 1976, production from these reservoirs was about 300 billion cubic feet. However, an accurate estimate of total natural gas produced from Pennsylvanian reservoirs is difficult to obtain because many wells are multiple completions in Pennylvanian as well as in older and younger rocks. Most of the natural gas from the Middle and Upper Pennsylvanian is casinghead gas (produced with crude oil). Most of the gas produced from lenticular sandstone bodies in the deltaic Morrowan sequences of the Delaware Basin is dry gas.

METALLIC ORES

Some of the red beds in Upper Pennsylvanian units, particularly in the north-central part of the State, contain scattered deposits of copper sulfides and carbonates. These are present mostly as nodules and as disseminated grains associated with small amounts of uranium and vanadium in arkose, green and gray shale, and nodular limestone (Zeller and Baltz, 1954). In the base-metal mining districts of central and southwestern New Mexico, Pennsylvanian limestones are host for some replacement deposits. They are also fractured and faulted and contain vein and stocklike metallic deposits. In most areas, however, the purer crinoidal limestones of the Mississippian are more favorable horizons for replacement deposits than the Pennsylvanian. The Carboniferous limestone sequence is also host to vein and breccia deposits of fluorite-barite-galena in south-central and central New Mexico where locally massive crystalline limestones have been selectively replaced.

LIMESTONE AND OTHER NONMETALLIC MINERALS

Throughout the State where Carboniferous limestone crops out, it is used locally for manufacture of cement (as at Tijeras east of Albuquerque), road metal, flagstone, lime, and dimension stone. Production of road metal is by far the largest use of Mississippian and Pennsylvanian limestones. Fossiliferous flagstones of Pennsylvanian limestone were placed in the courtyard of the Palace of Governors in Santa Fe almost four centuries ago.

Clay and shale of the Pennsylvanian have been used locally to make brick and tile, although for the most part, the argillaceous rocks are too highly calcareous. In south-central New Mexico, in the San Andres, Organ, and northern Franklin Mountains, gypsum beds in the upper part of the Pennsylvanian have been used locally in the manufacture of cement and as a soil conditioner. Some of the more even-bedded Pennsylvanian sandstones have been used in minor amounts for flagstones and for building some of the ancient dwellings.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— Montana

By DONALD L. SMITH *and* ERNEST H. GILMOUR

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Colorado

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—MONTANA

By DONALD L. SMITH[1] and ERNEST H. GILMOUR[2]

ABSTRACT

Carboniferous strata underlie all but the northwestern corner of Montana and are well exposed on the flanks of Tertiary uplifts throughout the State. The Carboniferous rock package attains a maximum thickness of 1,000 m along the Big Snowy trough, an east-trending paleostructural feature in central Montana; it thins to 300 m and 450 m in northern and southern Montana, respectively, on the flanks of the trough.

The contact of the Carboniferous rocks with underlying strata is unconformable, the rocks beneath the unconformity ranging in age from Ordovician to latest Devonian and generally increasing in age toward the southern part of the State. The contact of Carboniferous strata with overlying rocks is also unconformable, the overlying strata ranging from Permian and Triassic at the Montana-Wyoming border to Middle Jurassic in the northern part of the State.

Carboniferous rocks of Montana are divided into four lithologic units, each deposited under a different set of tectonic and environmental conditions. These units are the Madison Group, the Big Snowy Group, the Amsden Group, and the Quadrant Formation. In the Carboniferous section in Montana, a general upward decrease in clean limestone and an increase in both fine and coarse detrital components reflects the increasing epeirogenic-orogenic tempo of the later Carboniferous.

Major groups of Carboniferous strata were named or recognized in the late 1800's; since then, detailed studies have refined the stratigraphic nomenclature to its present complexity. The Madison Group of Kinderhookian, Osagean, and Meramecian age consists of the Lodgepole, Mission Canyon, and Charles Formations, in ascending order. The Lodgepole is divided into the Cottonwood Canyon, Paine, and Woodhurst Members. The Big Snowy Group is Chesterian in age and incorporates the Kibbey, Otter, and Heath Formations. The Amsden Group is latest Mississippian (Springerian), Morrowan, and Atokan(?) in age and includes three formations—the Tyler, Alaska Bench, and Devils Pocket. The Tyler is divided into the unnamed lower member and the Cameron Creek Member. The Desmoinesian-age Quadrant Formation is the fourth package, completing the Carboniferous section in Montana.

Carboniferous strata in Montana were deposited predominantly on the western edge of the North American craton, but in the extreme western part of the State, Carboniferous sediments accumulated in the Cordilleran miogeosyncline. During the latest Devonian and earliest Mississippian, the craton was divided into four shallow marine basins, all separated by low-lying arches that, through the erosion of rocks as old as Cambrian, provided a source of fine-grained sediment. Throughout the remainder of Madison and Big Snowy deposition, the Montana part of the craton was characterized by stable shelves to the north and south, separated by the elongate Big Snowy trough that extended from the Cordilleran miogeosyncline on the west to the Williston basin in the extreme northeastern corner of the State. During deposition of the Amsden Group, uplift in northern Montana on the site of the former northern stable shelf provided clastic sediment to the Big Snowy trough, which continued to subside. Deposition of the Quadrant Formation brought the Carboniferous to a close; the coarse clastic sediments were provided from large western uplifts as well as from the eastern craton.

Deposition of Carboniferous rocks began with a complex interplay of sea-level change and epeirogenic warping of the craton. In shallow basins between low-lying arches, black shale and siltstone of the late Devonian Bakken, Exshaw, and Englewood Formations and the Sappington Member of the Three Forks Formation were deposited. This latest Devonian transgression was short-lived and the sea partly regressed from Montana as epeirogenic movements continued to block out arches and basins on the Montana craton, providing coarser clastic sediment to the intervening basins. The earliest Mississippian transgression from the Cordilleran miogeosyncline is recorded in the black shale and siltstone of the Cottonwood Canyon Member of the Lodgepole Formation and the upper black shale of the Bakken Formation. The environments in which these fine-grained clastic rocks accumulated quickly gave way to higher energy environments in which the bioclastic facies at the base of the Paine Member of the Lodgepole Formation were deposited.

After the deposition of the bioclastic facies of the Paine Member, the rate of transgression and/or subsidence of the Big Snowy trough outstripped the rate of carbonate sediment production, creating a deeper water environment in central Montana while shallow-water carbonate sediments contemporaneously accumulated on the stable shelves to the north and south. A decrease in the rate of sea-level rise and/or rate of subsidence of the Big Snowy trough and concomitant increased production of carbonate sediment led to the progradation of high-energy, shallow-water carbonate sediments from the north and south into the Big Snowy

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trough, leading to deposition of the cyclic neritic deposits of the Woodhurst Member of the Lodgepole Formation.

The Mission Canyon Limestone was deposited under similar but more stable conditions than was the Woodhurst Member of the Lodgepole Limestone. The limestone of the Mission Canyon records shallow-water high-energy conditions that eventually gave rise to restricted environments in which extensive evaporite deposits accumulated, creating the evaporite-solution-breccia couplets of the surface and subsurface of Montana. The Charles Formation is probably the evaporite-rich subsurface equivalent of the upper brecciated part of the Mission Canyon Limestone. After deposition of the Mission Canyon-Charles evaporites, the Madison sea retreated to the Cordilleran miogeosyncline from the craton, and the ensuing exposure and erosion produced a karst surface of regional extent.

In Chesterian time, the Big Snowy sea transgressed across this surface, beginning in southwestern Montana and proceeding across the Big Snowy trough to the Williston basin. At the leading edge of this sea, the Kibbey Sandstone was deposited in beach and nearshore environments. Synchronously, in deeper, quieter water eastward of the coarse clastic zone, shale and limestone of the Otter Formation were deposited. In the deep and quiet water of the trough axis, dark shale and limestone of the Heath Formation accumulated. After deposition of the Big Snowy Group, the sea withdrew from most of the Big Snowy trough, creating an unconformity between Big Snowy strata and overlying deposits.

During the next major transgressive phase of the Carboniferous, strata of the Amsden Group were deposited. The Tyler Formation was deposited in either stagnant marine or nonmarine environments in central and eastern Montana along the axis of the Big Snowy trough. A marine limestone tongue, the Bear Gulch Limestone Member, near the top of the Tyler in central Montana, bears a marine fauna identified as latest Mississippian. Thus, the Mississippian-Pennsylvanian systemic boundary appears to be within the Tyler Formation rather than at the unconformity at the Big Snowy-Tyler contact.

After deposition of the Tyler Formation, limestone, dolomite, and mudstone of the Alaska Bench Formation accumulated in the Big Snowy trough in marine environments that ranged from supratidal to subtidal.

Dolomite, limestone, sandstone, and shale of the Devils Pocket Formation were deposited unconformably over the top of the Alaska Bench Formation in the Big Snowy trough. These sediments, like those of the underlying Alaska Bench Formation, were deposited in shallow marine environments.

After deposition of the Amsden Group in Montana, the Quadrant Formation accumulated in Desmoinesian time. The contact with the underlying Devils Pocket is gradational. The Quadrant Formation is probably of shallow marine origin. Uplifts to the west and the craton to the east provided abundant quartz sand.

INTRODUCTION

Carboniferous rocks of Montana have been the subject of man's curiosity and exploitation from approximately 2,000 years before the present, when stone-age miners quarried Mississippian chert from bluffs overlooking the Madison River in southwestern Montana (Davis, 1976), to the last several decades, when Carboniferous strata have produced petroleum, water, and aggregate and have provided abundant intellectual stimulus and challenge to students of Mississippian and Pennsylvanian rocks. Geologic investigation of Carboniferous strata from Peale's (1893) early work to the present has revealed a complex and detailed geologic history of transgressions and regressions, depositional environments, and epeirogenic activity; however, in the process of providing answers, this investigation has supplied a great number of unanswered questions. As a consequence, this paper is a compilation of work on diverse topics by many geologists having diverse backgrounds and orientations and should be read accordingly. The manuscript is a joint effort: Smith is responsible for the Mississippian parts; Gilmour, for the Pennsylvanian. We have made little attempt to resolve current stratigraphic, paleontologic, and sedimentologic conflicts and have chosen to present both arguments where two interpretations exist.

Carboniferous strata underlie all of Montana except the northwest corner, where they have been removed by Tertiary erosion (fig. 1). They are well exposed in the Cenozoic uplifts of central Montana, in the overthrust belt of western Montana, around large Laramide intrusive bodies, and on the flanks of basement block uplifts. In these exposures, the Madison Group forms a series of strong cliffs and ridges above the relatively nonresistant Devonian Three Forks Formation, the Big Snowy and Amsden Groups form a swale with low ridges, and the Quadrant Formation crops out in a series of cliffs and ridges.

Carboniferous rocks attain a maximum thickness of 1,000 m in an east-trending belt in central Montana, thinning to less than 300 m to the north and 450 m to the south. These thickness trends reflect cratonic paleotectonic elements that controlled Carboniferous sediment accumulation as well as Carboniferous and post-Carboniferous erosional events.

The contact of the Carboniferous rocks with the underlying rocks is generally unconformable, the age of the rocks under the unconformity generally increasing southward into Wyoming. However, in central and eastern Montana, the time value of this unconformity increases along the Bearpaw and Cedar Creek anticlines and the central Montana uplift, reflecting the positive nature of these structures during latest Devonian and earliest Mississippian

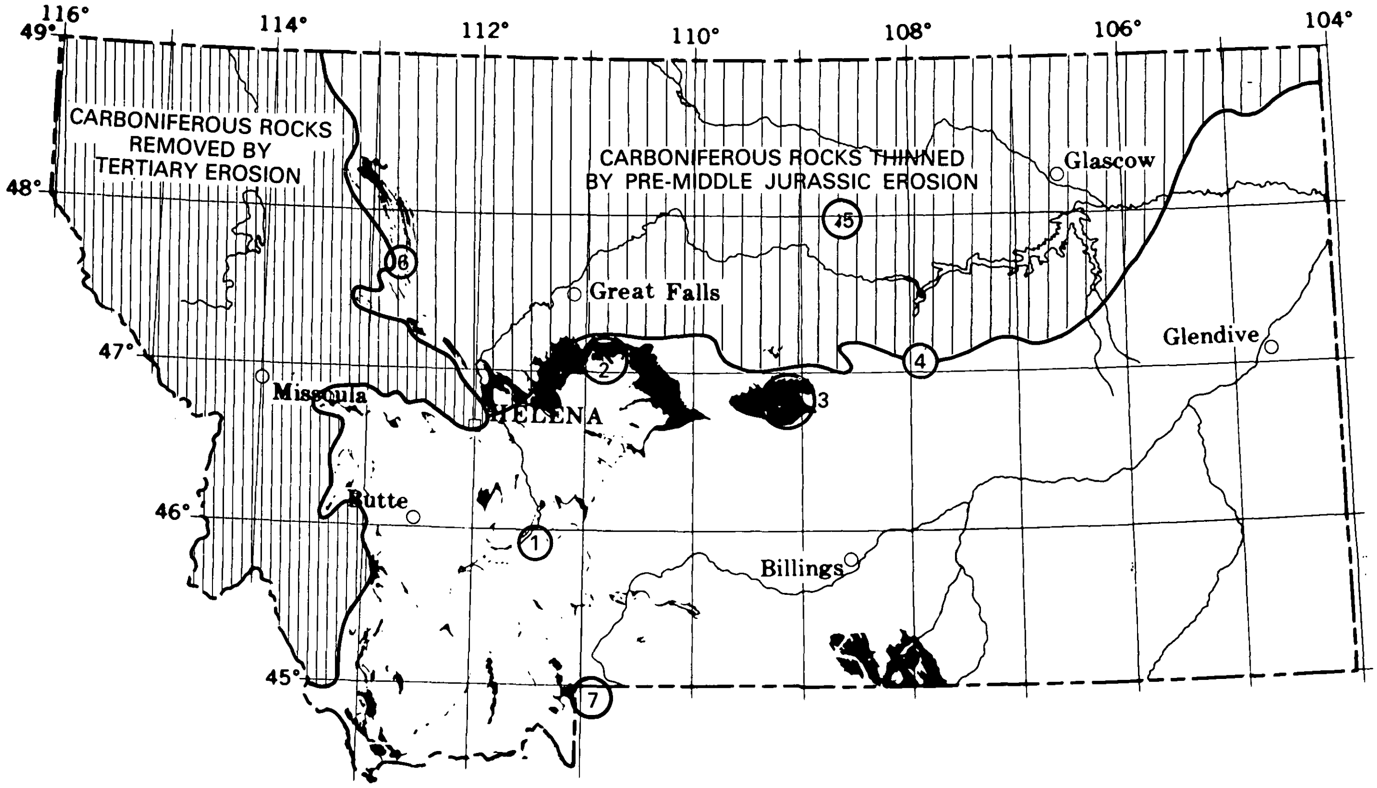


FIGURE 1.—Index map of major exposures of Carboniferous rocks in Montana, showing areas modified by pre-Jurassic and Tertiary erosion. Numbers refer to Carboniferous type and reference sections. (1) Logan-Three Forks area-type section of the Madison Group. (2) Northern Little Belt Mountains-type sections of the Paine and Woodhurst Members of the Lodgepole Limestone and reference section for the Mission Canyon Limestone. (3) Big Snowy Mountains-type sections of the Kibbey, Otter, and Heath Formations of the Big Snowy Group, the Bear Gulch, Cameron Creek, and Stonehouse Canyon Members of the Tyler Formation, and the Devils Pocket and Alaska Bench Formations. (4) Arro Oil and Refining Company and California Company No. 4 well-type well of the Charles Formation. (5) Little Rocky Mountains-type sections of the Lodgepole and Mission Canyon Formations. (6) Sawtooth Range-type sections of Allan Mountain Limestone, Castle Reef Dolomite, and Sun River Dolomite. (7) Quadrant Mountain, Gallatin Range, Wyoming-type section of the Quadrant Formation. (Map modified from Ross, Andrews, and Witkind, 1955, and Sando, 1976).

time. This deformation may be related to the Antler orogeny in the Cordilleran geosyncline to the west.

The contact of Carboniferous rocks with overlying strata is also unconformable and reflects the tectono-sedimentary history of the area from the beginning of the Permian through the Middle Jurassic. Beneath this unconformity, rocks as old as Osagean are unconformably overlain by the Middle Jurassic Ellis Group in northern Montana, whereas, to the south, Permo-Triassic rocks truncate the Middle and Upper Pennsylvanian Quadrant Formation. This erosional interval produced a northern zero edge on the Big Snowy, Amsden, and Quadrant Formations and significantly thinned the upper part of the Madison Group. This truncation suggests late Paleozoic-early Mesozoic uplift along the "Milk River uplift" of Maughan and Roberts (1967).

Carboniferous strata of Montana consist of four depositional packages of rock, each deposited under unique paleotectonic conditions. These are the dominantly carbonate Madison Group, the clastic and carbonate Big Snowy and Amsden Groups, and the clastic Quadrant Formation. The upward increase in detrital components in these Carboniferous units reflects increasing epeirogenic tempo on the craton and the influence of the Antler orogeny in the Cordilleran geosyncline to the west.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Montana Bureau of Mines and Geology.

HISTORY OF NOMENCLATURE

Carboniferous nomenclature in Montana has evolved from Peale's (1893) naming of the Madison limestone and overlying Quadrant Formation to the present Madison, Big Snowy, and Amsden Groups and the Quadrant Formation. The history of this evolution has been reviewed and clarified for the Madison Group by Sando and Dutro (1974) and has been discussed for the Big Snowy and Amsden Groups and the Quadrant Formation by Maughan and Roberts (1967). The following discussion has been derived from original sources, with the guidance of these papers. Important developmental stages are charted in figure 2.

MADISON GROUP NOMENCLATURE

Madison Group nomenclature was initially derived from the Three Forks area of southwestern Montana, but formation names were soon added from the Little Belt Mountains and, decades later, from the Little Rocky Mountains, the subsurface of central Montana, and outcrops in northwest Montana (fig. 1).

Prominent lower Carboniferous outcrops in the Three Forks area were given the name Madison Limestone by Peale (1893). In this area, these rocks are sandwiched between the Devonian Three Forks Formation and the upper Carboniferous Quadrant Formation. Although Peale did not designate a type section, and there has been much subsequent debate about whether the type section is along the Madison River or in the Madison Range, Sloss and Hamblin (1942) proposed and Sando and Dutro (1974) have concurred that the type section is along the Madison River at Logan (fig. 1). In this area, Peale (1893) divided the Madison into the Laminated limestones, the Massive limestones, and the Jaspery limestones, in ascending order. Several years later, Weed (1899, 1900) recognized three similar lithologic units in the Little Belt Mountains, which he termed Paine Shale, Woodhurst Limestone, and Castle Limestone. However, here, as in the Three Forks area, type sections were not designated, and Sando and Dutro (1974) proposed type sections for the Paine and Woodhurst and a reference section for the Mission Canyon Formation (Jaspery limestones) in the northern Little Belt Mountains (fig. 1).

Collier and Cathcart (1922) identified similar Madison rocks in the Little Rocky Mountains (fig. 1) and applied the local physiographic names Lodgepole and Mission Canyon to formations within the Madison that they recognized as a formal group (fig. 2). However, the threefold division of Peale (1893) and Weed (1899) is recognizable in the Little Rocky Mountains, and the Lodgepole also contains the lower two of the three stratigraphic units (Sando and Dutro, 1974).

In their synthesis of Madison nomenclature, Sloss and Hamblin (1942) gathered these lower Carboniferous names into a single classification scheme (fig. 2) that included Weed's (1899) Paine Shale and Woodhurst Limestone as members of Collier and Cathcart's (1922) Lodgepole Formation and that equated Peale's (1983) Jaspery limestones, Weed's (1899) Castle Limestone, and Collier and Cathcart's (1922) Mission Canyon Limestone, retaining the last name as a second formation at the top of the Madison group.

The name Charles was proposed by Seager (1942) for an interbedded sequence of limestone, dolomite, and evaporite above the Madison and below the Kibbey Formation in the subsurface of central Montana (fig. 1). Seager designated the Charles Formation as the basal unit of the Big Snowy Group, but a decade later, Sloss (1952) suggested that the Charles should be considered the uppermost unit of the Madison Group (fig. 2). Seager (1942) designated a type well for the Charles, for which a graphic log was prepared by Perry and Sloss (1943). The Charles is presently in a state of stratigraphic limbo, as indicated by Balster (1971, p. 220):

> There is some doubt that the Charles Formation persists to the outcrop. Stratigraphers disagree: some believe that the breccia zones in the uppermost part of the Mississippian represent the Charles with evaporites removed by selective solution; others believe that the Charles is restricted to the subsurface. Obviously, more study is necessary.

Sando and Dutro (1974) suggested that until a definitive study is conducted, the Charles should be considered the equivalent of the upper part of the Mission Canyon and that the use of the name should be restricted to the subsurface of the Williston basin.

The name Little Chief Canyon Member was proposed by Knechtel, Smedley, and Ross (1954) for a thin black shale at the base of the Lodgepole Limestone at its type section in the Little Rocky Mountains. This black shale is in a similar stratigraphic position in many parts of the State and is variously included in the Bakken Formation and the Cottonwood Canyon Member of the Lodgepole Formation (Sandberg and Klapper, 1967; Macqueen and Sandberg, 1970). The name Cottonwood Canyon Member was proposed by Sandberg and Klapper (1967) from exposures of black shale and siltstone at the base of the Madison Formation in the northern Big Horn Mountains of Wyoming, and Sando and Dutro (1974) recommended the use of this term and the abandonment of the name Little Chief Canyon Member.

The names Silvertip Conglomerate, Saypo Limestone, Dean Lake Chert, Rooney Chert, and Monitor Mountain Limestone were all proposed by Deiss (1933) for five members of the Madison Limestone in northwestern Montana. Later, Deiss (1941, 1943) included these members in the Hannan Limestone, replacing the term Madison. Since that time, these terms have been only locally used, and the Montana Geological Society (Balster, 1971) recommended that "more current terminology" be applied to these rocks.

Additional Madison terms for northwestern Montana (figs. 1, 2) were proposed by Mudge, Sando, and Dutro (1962, p. 2004) because ". . . formational boundaries recognized in the type locality cannot be consistently followed in this area." These names are Allan Mountain Limestone and Castle Reef Dolomite, which are chronologically and lithologically similar to the Lodgepole Limestone and Mission Canyon Limestone, respectively (Mudge, 1972). The Allan Mountain Limestone is divided into three unnamed members, and the Castle Reef Dolomite is composed of an unnamed lower member and the overlying Sun River Dolomite (Mudge, Sando, and Dutro, 1962; Mudge, 1972). The Montana Geological Society (Balster, 1971) suggested that the names Allan Mountain and Castle Reef should be abandoned in favor of the more widely used names Lodgepole and Mission Canyon.

BIG SNOWY GROUP NOMENCLATURE

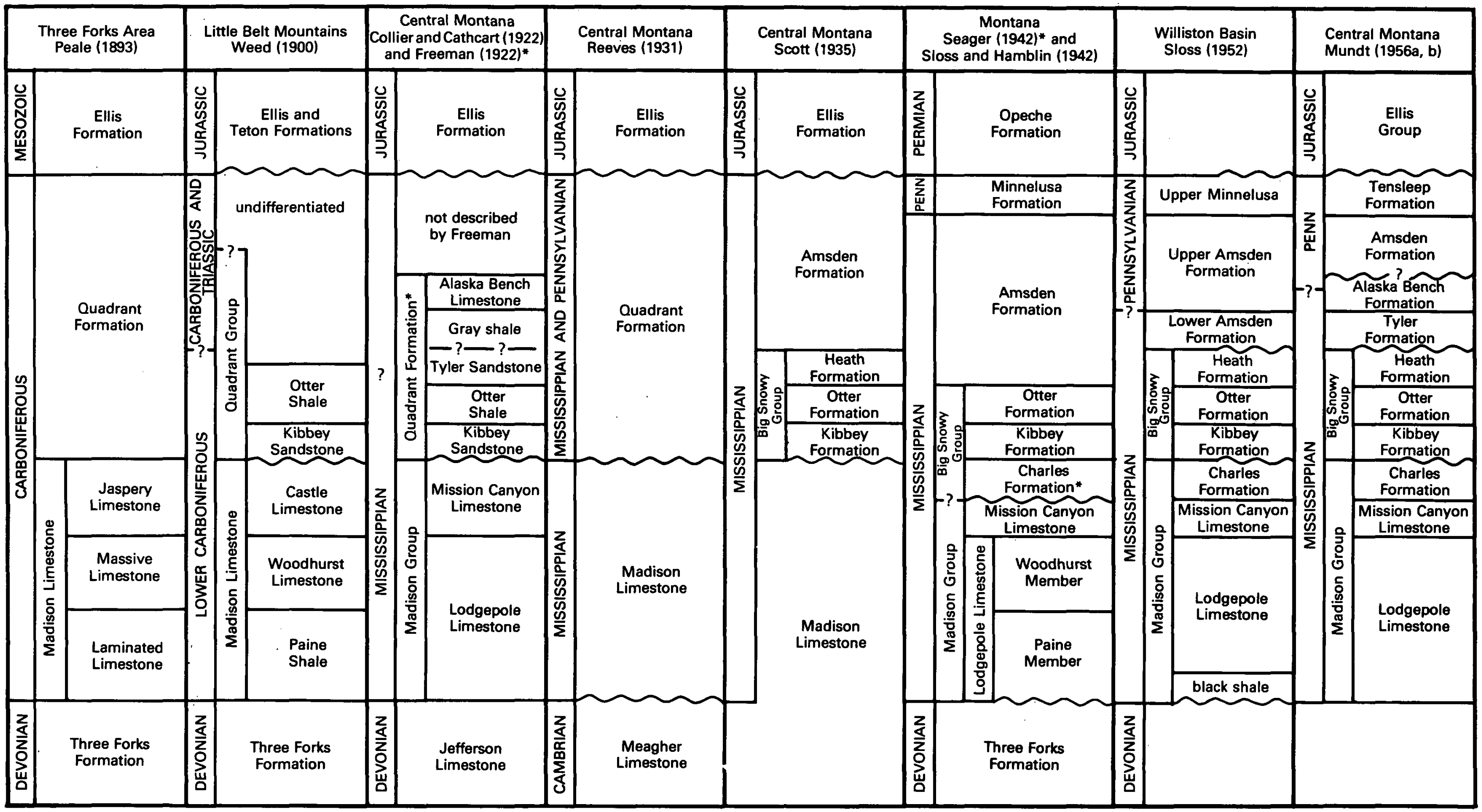
The Big Snowy Group was named by Scott (1935) from exposures in the Little Belt and Big Snowy mountains of central Montana (fig. 1). In the northern Little Belt Mountains, he included the Kibbey and Otter Formations that were previously described as part of the Quadrant Group by Weed (1899, 1900). The third formation of the Big Snowy Group—the Heath—is absent in the Kibbey and Otter type area, but is present in the eastern Little Belt Mountains and in the Big Snowy Mountains, where it was described by Scott (1935). With few exceptions, Scott's (1935) terminology has been used in subsequent syntheses; the marked exception was Gardner (1959), who expanded the Big Snowy Group to include rocks that Scott (1935) and subsequent investigators included in the Amsden Group (fig. 2).

AMSDEN GROUP NOMENCLATURE

Originally, rocks now considered part of the Amsden Group were included in the Quadrant Formation as used by Peale (1893), Weed (1896), Freeman (1922), and Reeves (1931). Freeman (1922) divided the Quadrant Formation into the Kibbey Sandstone, Otter Shale, Tyler Sandstone, and Alaska Bench Limestone (fig. 2).

Scott (1935) established the Big Snowy Group, which included the Kibbey Sandstone, the Otter Shale, and the overlying Heath Shale. Rocks above the Heath Shale were placed in the Amsden Formation, named by Darton (1904) for similar rocks exposed in northern Bighorn Mountains, Wyoming. Mundt (1956a) referred to sandstone and shale beds between the Heath Shale and the overlying limestone as the Tyler Formation and the overlying limestone as the Alaska Bench Formation (fig. 2). Rocks overlying the Alaska Bench Limestone were called the Amsden Formation by Mundt (1956a). Gardner (1959) and Easton (1962) placed the contact between the Heath Shale and the overlying sandstone and red beds at the lithologic break between the black shale and red beds. Gardner named this sandstone and red-bed sequence the Cameron Creek Formation and used the Alaska Bench Formation for the overlying limestone. He substituted the name Devils Pocket Formation for the overlying dolomite and sandstone referred to as Amsden Formation in earlier studies. Willis (1959) used the Tyler Formation to include the lower dark shale interstratified with sandstone above the black shale of the Heath Formation, which he called the lower member. The red shale and interbedded limestone sequence above the lower member was named the Cameron Creek Member. The carbonate rocks were grouped into the Amsden Formation (restricted) and divided into two members: the Alaska Bench Member and the dolomite member. Rocks above the dolomite member were referred to as the Tensleep Sandstone.

Easton (1962) and Gilmour (1967, 1969) used essentially the nomenclature suggested by Gardner. Gilmour (1969, p. 181) used Tyler Sandstone and Bear Gulch Limestone as member designations for parts of the Cameron Creek Formation.


Three Forks Area
Peale (1893)
MESOZOIC
Ellis Formation
CARBONIFEROUS
Quadrant Formation
Madison Limestone
Jaspery Limestone
Massive Limestone
Laminated Limestone
DEVONIAN
Three Forks Formation
Little Belt Mountains
Weed (1900)
JURASSIC
Ellis and Teton Formations
CARBONIFEROUS AND TRIASSIC
undifferentiated
Quadrant Group
Otter Shale
Kibbey Sandstone
LOWER CARBONIFEROUS
Madison Limestone
Castle Limestone
Woodhurst Limestone
Paine Shale
DEVONIAN
Three Forks Formation
Central Montana
Collier and Cathcart (1922)
and Freeman (1922)*
JURASSIC
Ellis Formation
not described by Freeman
Quadrant Formation*
Alaska Bench Limestone
Gray shale
Tyler Sandstone
Otter Shale
Kibbey Sandstone
MISSISSIPPIAN
Madison Group
Mission Canyon Limestone
Lodgepole Limestone
DEVONIAN
Jefferson Limestone
Central Montana
Reeves (1931)
JURASSIC
Ellis Formation
MISSISSIPPIAN AND PENNSYLVANIAN
Quadrant Formation
MISSISSIPPIAN
Madison Limestone
CAMBRIAN
Meagher Limestone
Central Montana
Scott (1935)
JURASSIC
Ellis Formation
Amsden Formation
MISSISSIPPIAN
Big Snowy Group
Heath Formation
Otter Formation
Kibbey Formation
Madison Limestone
Montana
Seager (1942)* and
Sloss and Hamblin (1942)
PERMIAN
Opeche Formation
PENN
Minnelusa Formation
Amsden Formation
MISSISSIPPIAN
Big Snowy Group
Otter Formation
Kibbey Formation
Charles Formation*
Mission Canyon Limestone
Madison Group
Lodgepole Limestone
Woodhurst Member
Paine Member
DEVONIAN
Three Forks Formation
Williston Basin
Sloss (1952)
JURASSIC
PENNSYLVANIAN
Upper Minnelusa
Upper Amsden Formation
Lower Amsden Formation
MISSISSIPPIAN
Big Snowy Group
Heath Formation
Otter Formation
Kibbey Formation
Madison Group
Charles Formation
Mission Canyon Limestone
Lodgepole Limestone
black shale
DEVONIAN
Central Montana
Mundt (1956a, b)
JURASSIC
Ellis Group
PENN
Tensleep Formation
Amsden Formation
Alaska Bench Formation
Tyler Formation
MISSISSIPPIAN
Big Snowy Group
Heath Formation
Otter Formation
Kibbey Formation
Madison Group
Charles Formation
Mission Canyon Limestone
Lodgepole Limestone

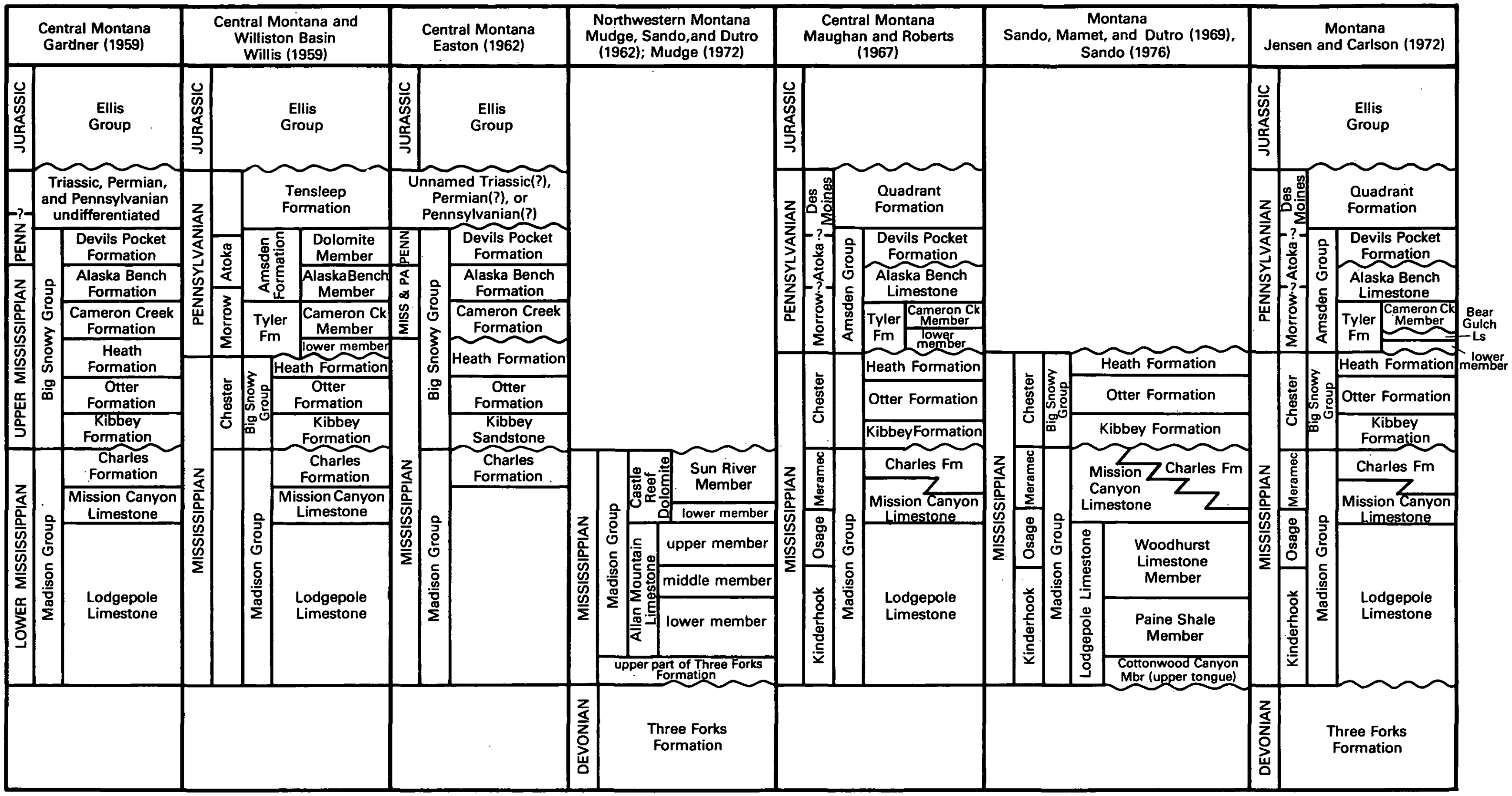


FIGURE 2.—History of Carboniferous nomenclature in Montana, 1893–1976. (Compiled from original sources with guidance from Maughan and Roberts, 1967, and Sando and Dutro, 1974.)

Maughan and Roberts (1967) and Maughan (1975) included all the rocks above the Heath Formation and below the uppermost sandstone beds (Quadrant Formation) as the Amsden Group, consisting of the Tyler Formation, Alaska Bench Limestone, and Devils Pocket Limestone. The Tyler Formation was divided into the Stonehouse Canyon Member and the overlying Cameron Creek Member. These units are overlain by the Alaska Bench Limestone, which, in turn, is overlain unconformably by the Devils Pocket Limestone. Maughan and Roberts (1967, p. B19) also recommended that this nomenclature be used in the subsurface throughout eastern Montana to the State boundary where the Minnelusa Formation is used in North Dakota, South Dakota, and eastern Wyoming. They also suggested that the strata found in southwestern Montana and western Wyoming, approximately equivalent to the Amsden Group in central Montana, be referred to as the Amsden Formation.

For this report, the nomenclature suggested by Maughan and Roberts (1967) is used. However, the use of formations and members should be based strictly on lithology and not on questionable unconformities in black-shale sequences, or between partially dolomitized limestones and totally dolomitized limestones.

QUADRANT FORMATION NOMENCLATURE

The Quadrant Formation was named by Peale (1893). Weed (1896) designated the type locality as Quadrant Mountain in the Gallatin Range, Wyoming (fig. 1), which included 32 m of limestone and shale now referred to as the Amsden Formation. Maughan and Roberts (1967, p. B6) used the name Quadrant Formation for the quartzite or sandstone sequence overlying the Devils Pocket Formation in central Montana and the Amsden Formation in southwestern Montana. Mallory (1972) also used Quadrant "Sandstone" for the same rocks in his synthesis of the Pennsylvanian System. Other workers (Scott, 1935; Mundt, 1956a; and Willis, 1959) assigned these rocks to the Tensleep Formation, which was named by Darton (1904) from the lower canyon of Tensleep Creek in the Big Horn Mountains, north-central Wyoming.

The term Quadrant Formation is used in this report in accord with workers in the U.S. Geological Survey. However, the authors recognize the general use of Tensleep Sandstone by workers in the petroleum industry in central and southern Montana.

GEOLOGIC SETTING

REGIONAL PALEOTECTONIC SETTING

Carboniferous sedimentation patterns are closely tied to the latest Devonian and Carboniferous tectonic framework of the northern Rocky Mountains. In eastern and central Montana, this framework was dominated by the North American craton. Western Montana was the site of the Cordilleran miogeosyncline. Because most of the Carboniferous strata were removed from western Montana by Tertiary erosion, the Carboniferous history of the Cordilleran miogeosyncline is incompletely known there. However, Huh (1967) documented the craton-miogeosynclinal margin in extreme southwestern Montana and adjacent Idaho, where he recognized a "transition zone" between these two major tectonic elements. Other students of the northern Rockies Carboniferous place the craton-miogeosyncline boundary in the vicinity of Huh's craton-transition line (figs. 3, 4).

MISSISSIPPIAN PALEOTECTONIC ELEMENTS

During the latest Devonian and earliest Mississippian, the craton in Montana was the site of parts of four shallow marine basins (fig. 4), the Bakken and Exshaw basins in northern Montana and the Sappington-Cottonwood Canyon and Englewood basins in the southern part of the State. These basins were separated by linear uplifts that may have been a cratonic manifestation of the Antler orogeny. The largest of these positive features was the central Montana uplift, an anticline that separated the two northern basins from the two southern basins and that probably provided sediment to both sides from the weathering and erosion of Cambrian through Devonian strata. In eastern Montana, the Cedar Creek anticline was similarly active, and as much as 230 m of Cambrian through Devonian rocks was eroded along a high-angle fault on its western flank (Sandberg and Mapel, 1967).

This uppermost Devonian and lowermost Mississippian pattern of shallow cratonic basins and uplifts changed markedly in Early Mississippian time. East of the craton-miogeosyncline hingeline, the Mississippian cratonic setting of Montana featured three major tectonic elements: the Cordilleran platform (Sando, Gordon, and Dutro, 1975) in northern and southern Montana, the unstable shelf or Big Snowy trough in central Montana, and the Williston basin in easternmost Montana and adjacent western North Dakota (fig. 3). The Cordilleran platform is, in turn, divided into a northern Alberta shelf and

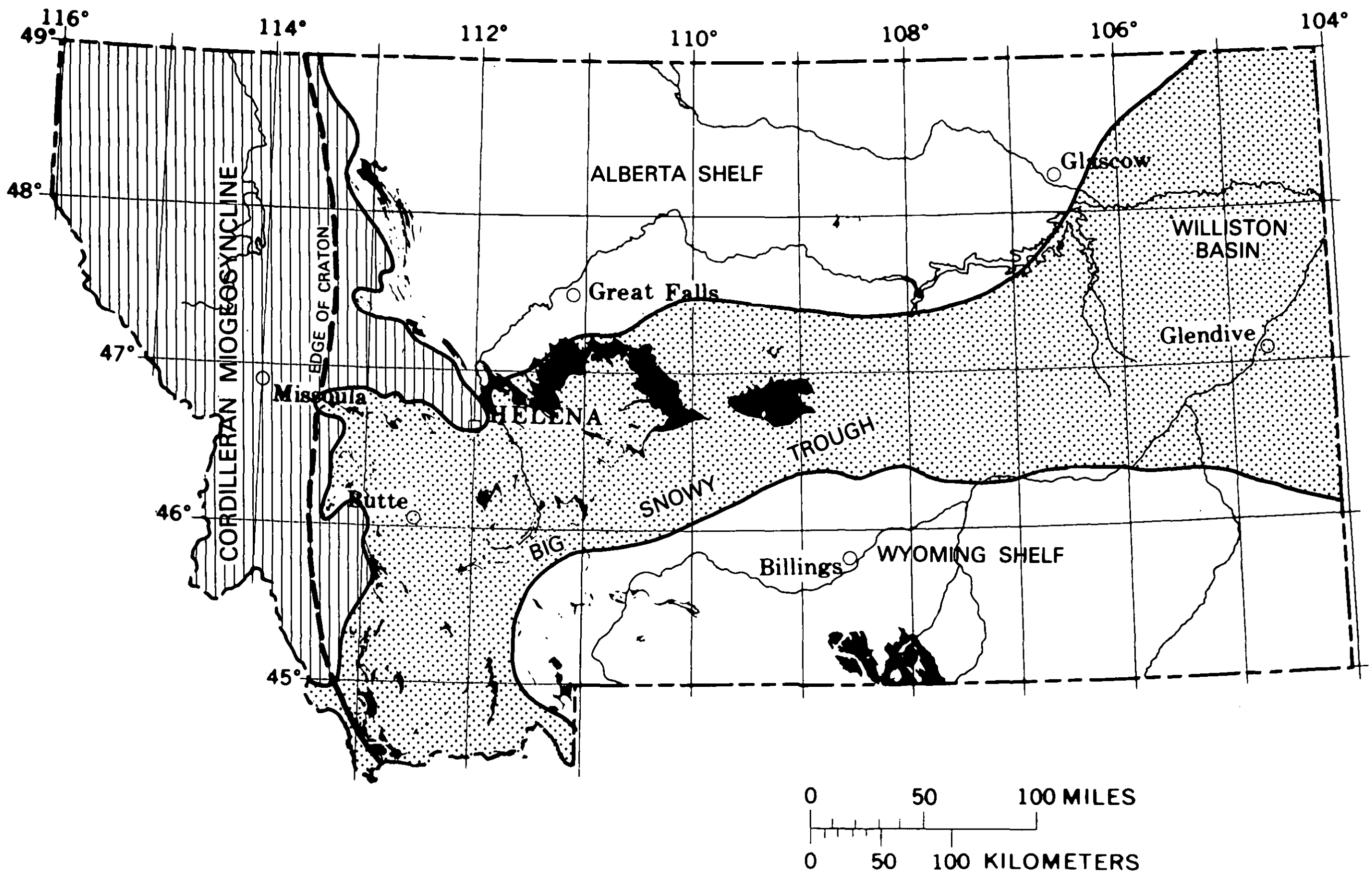


FIGURE 3.—Mississippian paleostructural features in Montana. The shelf-trough boundaries are drawn on Craig's 450-m isopach for the Mississippian. (Map modified from Sando, 1967; Rose, 1976; Sando, 1976; and Craig, 1972).

a southern Wyoming shelf, both having stable or relatively neutral Mississippian histories. These structural elements remained intact during deposition of the Madison Group but began to be modified at the onset of Big Snowy deposition (latest Meramecian-earliest Chesterian). During accumulation of the diverse rocks of the Big Snowy Group, the Big Snowy trough was inundated and the Wyoming shelf formed an east-trending peninsular feature—the southern Montana arch (Sando, 1976)—that separated the Wyoming basin to the south from the Big Snowy trough on the north. This uplift influenced sedimentation until the latest Chesterian when parts of the Big Snowy trough were uplifted, providing sediment to the Amsden Formation of the northward-expanding Wyoming basin (Sando, Gordon, and Dutro, 1975; Sando, 1976).

PENNSYLVANIAN PALEOTECTONIC ELEMENTS

At the beginning of Pennsylvanian time, the sea occupying the Big Snowy trough was severely restricted. Uplift to the north exposed land areas that provided much of the terrigenous sediment for the Tyler, Alaska Bench, and Devils Pocket Formations. The Big Snowy trough continued to subside through Morrowan and Atokan time, and deposition was more or less continuous in local areas. The Montana uplift (Sando, Gordon, and Dutro, 1975) produced a large island area in south-central and southeast Montana in late Morrowan time. During Atokan and Desmoinesian time, large uplifts in the west provided sand-sized clastic sediment for the Quadrant and equivalent formations.

STRATIGRAPHY

Carboniferous strata of Montana are divisible into four depositional packages, each representing a transgressive-regressive event. These packages are, in ascending order, the Madison Group, Big Snowy Group, Amsden Group, and Quadrant Formation. The time value of the hiatus separating these packages is variable, but, in general, is greater on the Cordilleran platform than in central Montana and the Williston basin.

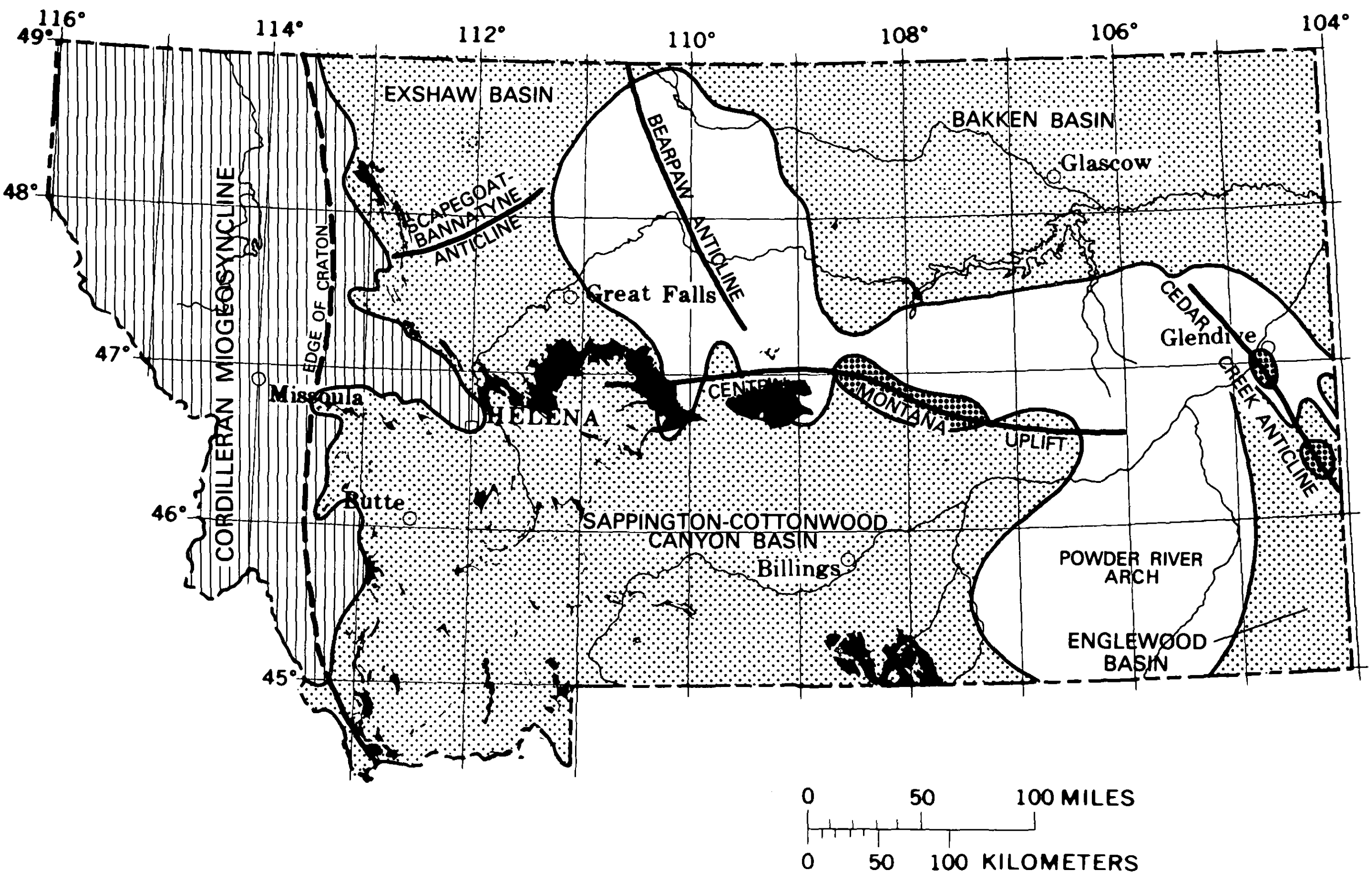


FIGURE 4.—Map of uppermost Devonian and lowermost Mississippian rocks and structural features. Light-shaded areas are basins in which rocks of the Bakken, Exshaw, and Englewood Formations and Cottonwood Canyon and Sappington Members accumulated. Unshaded areas in central and eastern Montana are uplifts that separated these basins. Dark-shaded areas along the axes of the central Montana uplift and the Cedar Creek anticline are areas where uppermost Devonian and lowermost Mississippian erosion cut down into Cambrian, Ordovician, and Silurian strata. (Map modified from Sandberg and Klapper, 1967; Sandberg and Mapel, 1967; and Baars, 1972).

UPPER DEVONIAN AND LOWER MISSISSIPPIAN STRATA

Black shale and dolomitic siltstone below and at the base of the Madison Group unconformably overlie earlier Devonian and other lower Paleozoic strata and record the complex history of the Devonian-Mississippian transition in Montana. Rocks included in this sequence are the Bakken and Exshaw Formations, the Sappington Member of the Three Forks Formation, and the Englewood Formation.

The Bakken Formation is restricted to the Bakken basin in northeastern Montana (fig. 4). Here, it attains a maximum thickness of 42 m in the North Dakota part of the Williston basin, and it thins from 21 m along the international boundary to zero on the northern flank of the central Montana uplift. The Bakken is composed of two black radioactive carbonaceous shale beds that sandwich a medial dolomitic siltstone and sandstone (Coleville Sandstone Member). To the west, the upper black shale has been removed by earliest Mississippian erosion on the Bearpaw anticline, and an arbitrary western limit has been drawn along this axis (Macqueen and Sandberg, 1970). Conodonts from the Bakken Formation indicate a Late Devonian age (lower and upper *to* V), and a spore flora indicates a position near the Devonian-Mississippian boundary (Macqueen and Sandberg, 1970). This boundary may be within the Coleville Sandstone Member (fig. 5).

In northwestern Montana, the Exshaw Formation is a southern extension of that formation from Alberta and, according to Macqueen and Sandberg (1970), represents a western continuation of the basal black shale and medial siltstone of the Bakken Formation (fig. 5). The Exshaw Formation is restricted to the Exshaw basin that is bounded on the east by the Bearpaw anticline, on the south by the Scapegoat-Bannatyne anticline, and on the west by the Cordilleran miogeosyncline (fig. 4). The formation attains its maximum thickness in Alberta and is only approximately 40 m thick at the international boundary. It thins southward onto the Scapegoat-Bannatyne anticline, where it is less than 6 m thick and where the siltstone was removed by erosion in early Mississippian time. The basal black carbonaceous shale is fissile and has at its base a thin phosphatic sandstone bed. This bed has been interpreted as a basal lag deposit, a product of the erosional interval represented by the Three Forks-Exshaw unconformity (Macqueen and Sandberg, 1970). The upper siltstone is calcareous, grading in places to a silty limestone. Its contact with the underlying black shale is gradational. South of the Scapegoat-Bannatyne anticline, the basal black shale of the Exshaw is equated with a similar shale at the base of the Sappington Member of the Three Forks Formation, and the medial Exshaw siltstone is considered to be equivalent to limestone, siltstone, and shale (units 2 through 5) of the Sappington Member (fig. 5). Conodonts collected from the Exshaw Formation north of the international boundary indicate a very late Devonian and early Mississippian age for the unit, the contact being somewhere within the black-shale unit (Macqueen and Sandberg, 1970).

The Sappington Member of the Three Forks Formation occupies the western end of the incipient Big Snowy trough, sandwiched between the Scapegoat-Bannatyne anticline and the Wyoming shelf (figs. 3, 4). It attains a maximum thickness of approximately 30 m in the center of the "Sappington basin" (Gutschick, McLane, and Rodriquez, 1976) and thins northward onto the Scapegoat-Bannatyne anticline and southward and eastward onto the Wyoming shelf. The Sappington Member was divided into five units by Sandberg (1962, 1965). These are: (1) a basal black carbonaceous shale, similar to the basal shale of the Exshaw, with which it is equated: (2) an alga-sponge biostromal limestone; (3) a lower siltstone; (4) a middle olive-gray shale; and (5) an upper calcareous siltstone, which is a cliff former and which is occasionally cross stratified and ripple marked. Analysis of conodont faunas and spore floras suggests that units 1 through 4 are all latest Devonian but that the upper part of unit 4 and all of unit 5 are earliest Mississippian (fig. 5).

The Cottonwood Canyon Member of the Lodgepole Formation occupies that part of the Cordilleran platform called the Wyoming basin. The member is in a linear belt from west-central Wyoming to the flanks of the Big Snowy uplift in central Montana (fig. 4). Along the axis of this belt, thicknesses are as great as 18 m, but the unit thins abruptly both east and west and is less than 3 m thick over half its extent (Sandberg and Klapper, 1967). In Montana, the Cottonwood Canyon Member unconformably overlies the Sappington and Trident Members of the Three Forks Formation and is overlain by and intertongues in its upper part with the Paine Member of Lodgepole Limestone. The Cottonwood Canyon Member is composed of two tongues (fig. 5), each of which is divisible into a western shale and siltstone facies and an eastern dolomitic facies (Sandberg and Klapper, 1967). The base of each tongue is characterized by a basal conglomeratic sandstone that contains phosphatic nodules, coprolites, conodonts, fish fossils, and glauconite grains. Sandberg and Klapper (1967) interpreted this rock as a lag deposit, the product of erosion and reworking during marine transgression. This basal lag deposit is similar in character and origin to those of the Sappington, Exshaw, and possibly the Bakken black-shale beds. The lower tongue of the Cottonwood Canyon Member is restricted for the most part to the Wyoming basin, bears latest Devonian conodonts, and intertongues with the basal beds of the Madison Formation. The upper tongue is more extensive than its lower counterpart, interfingers with the Madison Limestone, and is entirely of Mississippian age (fig. 5), according to Sandberg and Klapper's (1967) analysis of conodonts.

The Englewood Formation unconformably underlies the Madison-equivalent Pahasapa Limestone in the Black Hills and subsurface of southeastern Montana. Here, it accumulated in an embayment that extended from southeastern Wyoming to central North Dakota (Sandberg and Mapel, 1967). In Montana, the Englewood is 0 to 6 m thick and unconformably overlies the Devonian Three Forks and Jefferson Formations and the Ordovician Bighorn Dolomite (Baars, 1972). The Englewood is characterized by a basal shale that is overlain by silty dolomite, dolomitic limestone, and limestone containing beds of sandstone, shale, and siltstone. Conodont faunas from the Englewood were sampled and analyzed by Klapper and Furnish (1962), Sandberg (1963), and Klapper (1966) and determined to be of Late Devonian-Early Mississippian age.

A complex depositional history was proposed for these uppermost Devonian and lowermost Mississippian rocks by Sandberg and Klapper (1967), a history that features shallow marine basins isolated

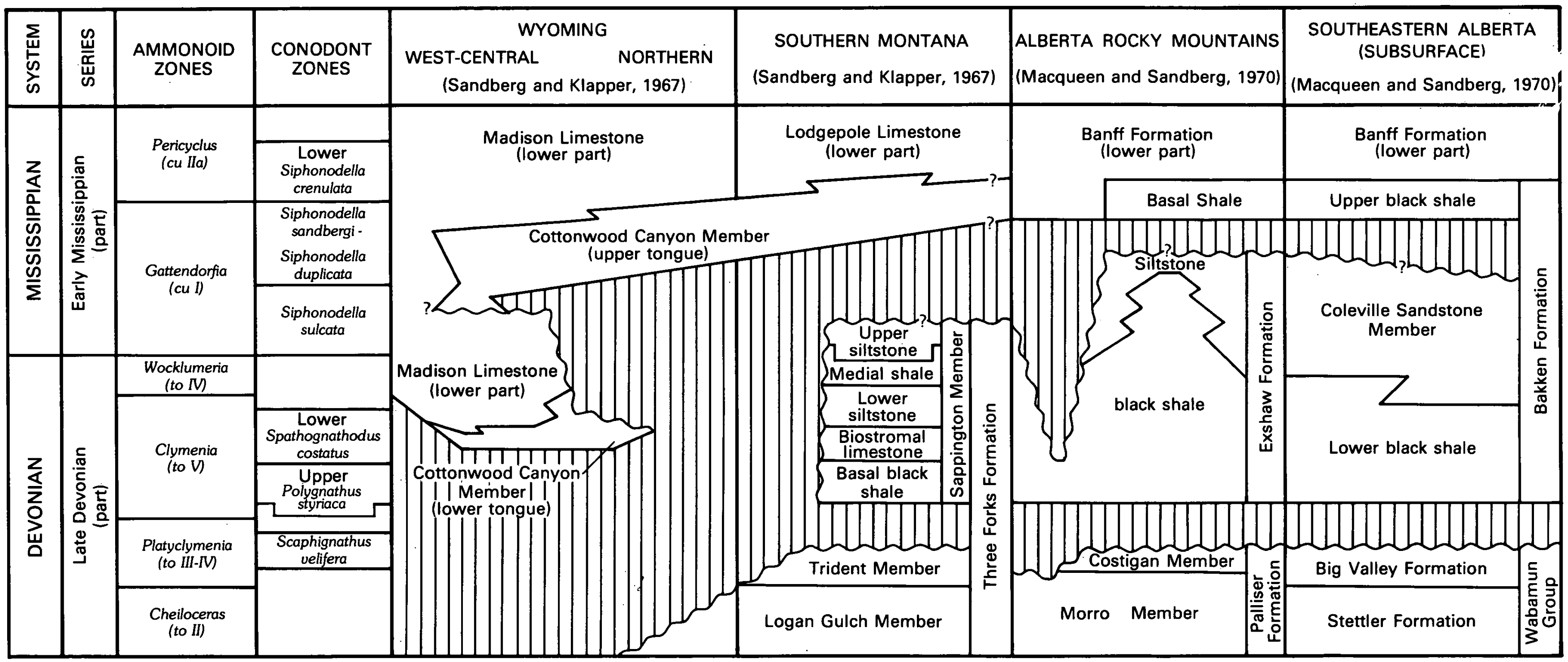


Figure 5.—Stratigraphic relationships of uppermost Devonian and lowermost Mississippian strata in northern Wyoming, southern Montana, and southern Alberta. (Modified from Sandberg and Klapper, 1967; Sandberg, 1965; Gutschick, Suttner, and Switek, 1962; and Macqueen and Sandberg, 1970).

by Antler orogeny-related cratonic uplifts and by minor transgressions and regressions.

In the latest Devonian (Lower *to* V), a marine transgression took place that inundated the shallow marine basins of the Montana craton (fig. 4). This transgression is recorded by the basal sandstone and black shale of the Sappington, Exshaw, and Bakken. The sandstone in each of these units represents lag accumulations on the weathered and eroded surface of the Three Forks Formation. Fine-grained clastic debris was probably derived from weathering and erosion of low-lying, basin-separating uplifts (fig. 4) and was deposited in shallow stagnant marine water.

Accompanying this brief transgression were more intense movements of the uppermost Devonian and lowermost Mississippian arches and uplifts, most notably the central Montana uplift and the Cedar Creek anticline (fig. 4). Erosion of these and other positive elements provided coarser sediment to the regressing latest Devonian sea, resulting in the accumulation of the medial siltstone of the Bakken Formation, the upper siltstone of the Exshaw, and units 2 through 5 (Sandberg, 1965) (units E through H of Gutschick, Suttner, and Switek, 1962) of the Sappington Member of the Three Forks Formation. These sediments accumulated in a variety of intertidal and subtidal depositional environments in the Bakken, Exshaw, and Sappington basins.

At the same time that the sea was regressing from the Montana basins and these coarser clastic rocks were accumulating there, the lower tongue of the Cottonwood Canyon Member of the Lodgepole Limestone was deposited in a shallow trangressive sea in the Cottonwood Canyon or Wyoming basin in western Wyoming. Here, deposition of the basal black shale and sandstone of the lower tongue took place in an environment similar to that of the earlier Montana basins. This stagnant marine environment gave way in time and space to less restricted environments in which the lower part of the Madison Limestone was deposited during latest Devonian time (Upper V and *to* IV). This initial transgression of the Madison sea was short lived, ending in a brief regression during earliest Mississippian time (Lower *cu* I). At approximately this same time, the sea may have regressed from the Sappington basin, but shallow seas continued to occupy the Exshaw and Bakken basins, and black shale and siltstone continued to accumulate there. However, prior to the invasion of the craton by the second and major advance of the Madison sea, these shallow seas regressed from all the Montana basins, with the exception of the western part of the Sappington basin (Sandberg and Klapper, 1967), producing the interregional unconformity shown in figure 5.

The second advance of the Madison sea was not restricted to isolated basins and embayments as was the first transgression, but was an eastward transgression along the entire length of the Cordilleran miogeosyncline (Sandberg and Klapper, 1967). During this relatively rapid transgression, the basal lag sandstone and black shale of the upper tongue of the Cottonwood Canyon Member accumulated in shallow stagnant environments, as did the contemporaneous upper black shale of the Bakken Formation (fig. 5). As transgression continued, these stagnant conditions gave way to the more agitated and oxygenated environments in which the shale and siltstone facies and dolomite facies of the upper tongue of the Cottonwood Canyon Member were deposited in intertidal and shallow subtidal environments. As the transgressing Madison sea rapidly engulfed and effectively mantled sources of detrital sediment, the bioclastic facies at the base of the Paine Member of the Lodgepole Limestone was deposited in a series of shoals. This coarse-grained, glauconitic, bioclastic limestone spread rapidly eastward across the stable and unstable shelves of Montana in response to the rapid expansion of the shallow detritus-free Madison sea.

MADISON GROUP

Excellent syntheses of sedimentation, stratigraphy, paleontology, and depositional history of the Madison Group in Montana and adjacent areas have been published by Craig (1972), Sando (1976), Rose (1976), and Gutschick, McLane, and Rodriguez (1976), and comprehensive lists of references are found in these papers.

The Madison Limestone covers all of Montana except for the centers of Tertiary uplifts and the northwestern part of the State, where it was removed by Tertiary erosion. It is a blanket of limestone and dolomite that reaches maximum thicknesses of 550 to 660 m in the Big Snowy trough and Williston basin, respectively, 600 m in the miogeosyncline of extreme southwestern Montana, and that ranges in thickness from 200 to 450 m on the Wyoming and Alberta shelves (fig. 6). The northward thinning is due to pre-Jurassic erosion; the southward thinning is primarily depositional, although original thicknesses may have been modified here by the formation of a late Mississippian karst across much of the area (Henbest, 1958; Roberts, 1966; Sando, 1974).

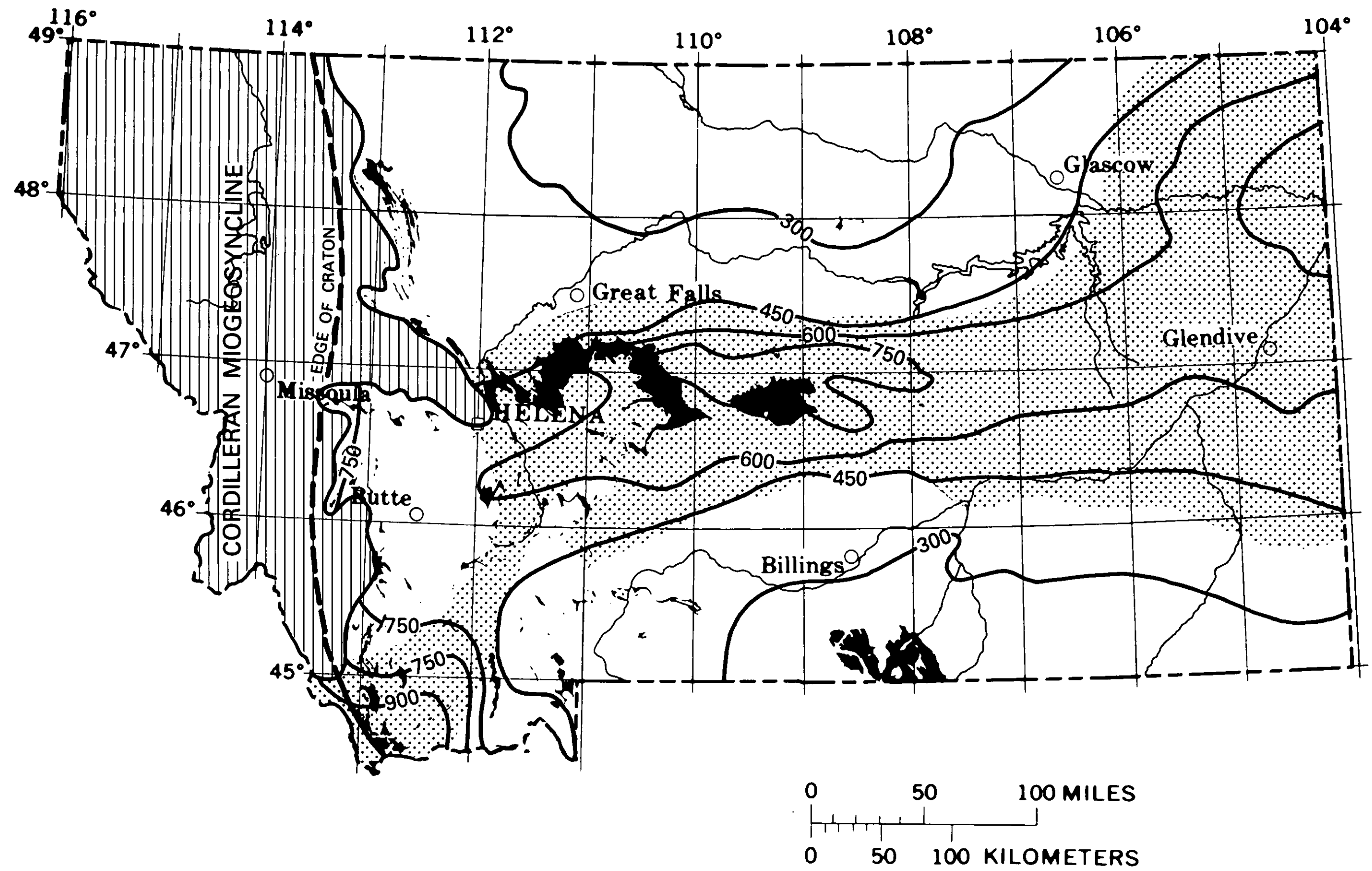


FIGURE 6.—Total thickness of Mississippian rocks in Montana. Light-shaded area indicates extent of Big Snowy Group. Isopach values are given in meters, and isopachs are modified from those of Sando (1976) and McMannis (1965).

In Montana, the Madison Group is composed of the Lodgepole, Mission Canyon, and Charles Formations. The Lodgepole is divided into the Cottonwood Canyon, Paine, and Woodhurst Members. These rocks are products of the first major transgression of the Madison sea onto the North American craton in Montana. The Lodgepole records (1) the initial rapid inundation of the Montana stable and unstable shelves, beginning with the accumulation of shallow-water sediments of the Cottonwood Canyon Member; (2) accumulation of deeper water deposits in the Big Snowy trough and expansion of the Madison sea on the Cordilleran platform as transgression continued; and (3) progradation of shallow-water carbonate sediment from the stable shelves across the Big Snowy trough, bringing to a close the deeper water phase of Lodgepole deposition. The Mission Canyon and Charles Formations accumulated as the Madison sea began to retreat prior to the late Mississippian emergence of the craton.

The biostratigraphic studies of conodonts, foraminifers, and corals and brachiopods summarized by Sando, Mamet, and Dutro (1969) indicate that the Lodgepole is Kinderhookian and early Osagean and that the Mission Canyon and Charles are late Osagean and early Meramecian.

LODGEPOLE LIMESTONE

The Lodgepole Limestone is a slightly lenticular deposit 240 m thick in the Big Snowy trough and Williston basin; it thins northward to 170 m in the Little Rocky Mountains on the Alberta shelf and southward to 130 to 150 m in the Beartooth Mountains on the Wyoming shelf. In the miogeosyncline of extreme southwestern Montana, the Lodgepole is more than 300 m thick. Lodgepole strata unconformably overlie Upper Devonian rocks throughout most of Montana, except along the central Montana uplift and the Cedar Creek anticline, where they overlie rocks as old as Cambrian and Ordovician (fig. 4).

Throughout most of Montana, the Lodgepole Limestone is divided into two members, the Paine and the Woodhurst. However, in southwestern Montana, a third member is present where the upper tongue of the Cottonwood Canyon Member intertongues with and is overlain by the widespread bioclastic and glauconitic limestone at the base of the Paine Member. A similar stratigraphic configuration exists in north-central and northeastern Montana, where the upper black shale of the Bakken Formation unconformably overlies the medial Bakken siltstone but conformably underlies the Lodgepole Limestone (fig. 5). In northwestern Montana, temporal and lithologically equivalent strata are called the Allan Mountain Limestone, which is divided into three unnamed members (fig. 2).

The Paine Member of the Lodgepole Limestone is most extensive and thickest in the Big Snowy trough in central Montana, where it is 75 to 90 m thick. From this slight thickening along the axis of the trough, the Paine Member thins north and south onto the Alberta and Wyoming shelves. On the Wyoming shelf, the Paine Member thins to 45 m before it becomes indistinguishable by intertonguing with the lower dolomite member of the Madison Formation (Sando, 1972).

At the base of the Paine Member, Upper Devonian strata or the siltstone of the Cottonwood Canyon Member are overlain by a widespread but discontinuous sheet of light-gray bioclastic and glauconitic limestone 0 to 17 m thick. Rocks of this unit are both micritic and sparry and are composed of coarse fragments of crinoids, brachiopods, bryozoans, and corals. This basal unit is abruptly overlain by 30 to 70 m of uniformly thin-bedded dark argillaceous lime mudstone and calcareous shale. Alternation of these rocks produced the distinctive rhythmic outcrop pattern of the Paine Member. These dark lime mudstone beds are characterized by a sparse fauna of crinoids, corals, bryozoans, brachiopods, and spicules of unknown origin, in addition to the trace fossils *Cosmoraphe* and *Scalarituba* (Gutschick, McLane, and Rodriguez, 1976). Large "interformational truncation surfaces" and smaller scale soft-sediment deformation features are also found in this dark lime mudstone (Wilson, 1969; Smith, 1977).

The dark lime mudstone of the Paine Member enclosed lime mud bioherms in the Bridger Range and the Big Snowy Mountains, as well as in several other localities in central Montana and along the Cedar Creek anticline. These are "Waulsortian" bioherms, comparable with those of similar age in North America and Europe (Cotter, 1965, 1966; Stone, 1972). The cores of these bioherms are characterized by alternating bioclast-rich and bioclast-poor layers, inclined at as much as 35° to the enclosing dark lime mudstone bedding and traceable vertically for as much as 50 m. Fossil components of these layers are more diverse than those of the enclosing dark lime mudstone and include crinoids, fenestrate bryozoans, brachiopods, coelenterates, and mollusks (Merriam, 1958). Between bioherm cores is a flank facies, composed of large crinoid fragments in a lime-mud matrix.

The upper few meters of the Paine Member is gradational from dark lime mudstone to thicker bedded, lighter colored, pellet and bioclastic limestone beds. The top of the member is placed ". . . at the base of the lowest crinoidal limestone bed, which marks the beginning of cyclical alternation of crinoidal limestone (commonly oolitic) and shaly, predominantly fine grained limestone characteristic of the Woodhurst Member" (Sando and Dutro, 1974, p. 4).

The Woodhurst Member of the Lodgepole Limestone is easily distinguished from the underlying Paine Member by its distinctive outcrops of medium to thick resistant beds that are cyclically interspersed with recessive units of thinner beds. This member is more widespread than the Paine Member and occurs not only in central Montana but also on the Cordilleran platform in southern and northern Montana. The Woodhurst Member is nearly 180 m thick in the Big Snowy trough in central Montana, thinning to less than 90 m to the north and south on the Alberta and Wyoming shelves.

Woodhurst lithologies are arranged in cyclic packages, each package consisting of a thin-bedded, fine-grained, nonresistant lower part that is capped by more thickly bedded, coarser grained bioclastic and oolitic limestone beds that form distinctive resistant ledges. Definition of these cycles has varied, from the 28 described by Laudon and Severson (1953) in the Bridger Range to the 5 to 8 described by Wilson (1969) and Smith (1972) from central Montana. A typical Woodhurst cycle begins with a mixed oolite and bioclastic lime grainstone interval that commonly overlies an undulating surface on top of the capping bed of the previous cycle. Overlying this initial unit are thin beds of pellet and bioclastic grainstone and packstone beds. Bioclastic components in these beds include crinoid fragments, dasyclad algae, fragments of bryozoans, and endothyrid foraminifers. These beds grade upward into more bioclast-rich, cross-stratified lime grainstone beds which are capped by resistant thick-bedded oolitic and bioclastic grainstone beds that complete the cycle. This capping bed is composed of crinoid-cored oolites and abundant crinoid debris and is characterized by trough cross-stratification and ripple-drift cross-lamination (Jenks, 1972).

The contact of the Woodhurst Member is conformable with the overlying Mission Canyon Formation and is generally placed at the base of the first massive cliff-forming limestone bed above the base of the Madison Group. According to Sando and Dutro (1974, p. 4), ". . . the top of the Woodhurst is placed at the top of the highest shaly, thin-bedded, predominantly fine grained limestone beneath the thicker crinoidal beds of the Mission Canyon."

According to Sando, Mamet, and Dutro (1969), the Lodgepole Limestone includes two foraminiferal zones and part of a third, four coral-brachiopod megafaunal zones (fig. 7), and, at its base, three conodont zones (fig. 5). The Cottonwood Canyon Member contains conodonts of the early Kinderhookian *Siphonodella sandbergi–S. duplicata* Zone and is included in the Cordilleran megafaunal pre-A Zone and global foraminiferal pre-7 Zone. The Paine Member is also included in the pre-7 Zone but is characterized by Kinderhookian Zone A corals and brachiopods and by conodonts of the *Siphonodella crenulata* Zone (Sandberg and Klapper, 1967). The Woodhurst Member includes the latest Kinderhookian-early Osagean megafaunal C_1 Zone and foraminiferal zone 7 and the lower part of zone 8 (fig. 7).

The depositional history of all or part of the Lodgepole Formation has been reviewed by Sando (1976); Gutschick, McLane, and Rodriguez (1976); Rose (1976); Rodriguez and Gutschick (1970); Sandberg and Klapper (1967); and Smith (1972, 1977).

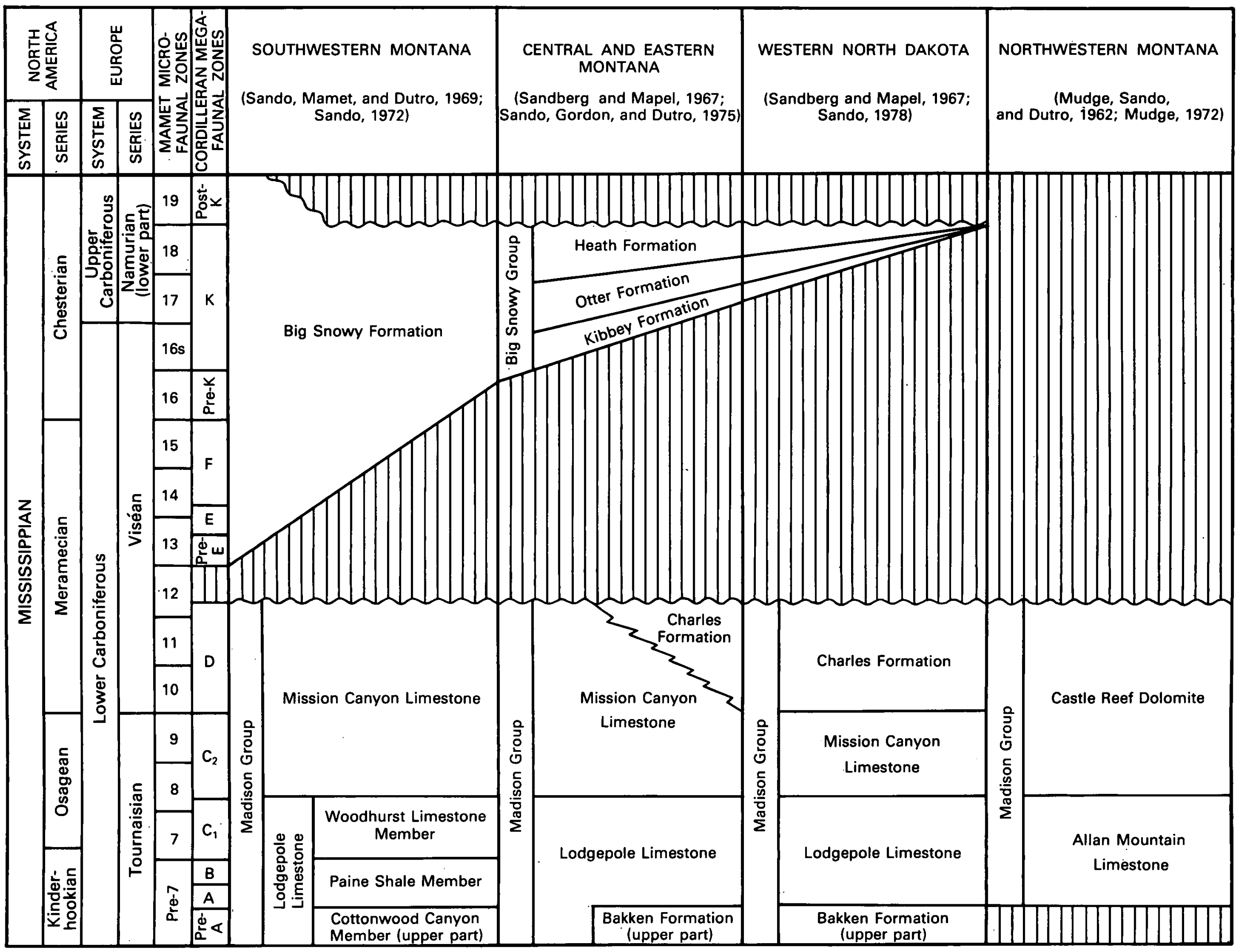


FIGURE 7.—Nomenclature, faunal zones, and temporal relationships of Mississippian rock units in Montana and adjacent areas. Vertical lines indicate hiatus. (Modified from Sando, 1976).

Lodgepole strata record the initial incursion of the Madison sea from the Cordilleran miogeosyncline onto the craton in Montana. This transgression began in latest Devonian time and resulted in the accumulation of the Cottonwood Canyon Member in southwestern Montana and the upper black shale of the Bakken Formation in the northeastern corner of the State. These rocks were deposited over an irregular erosion surface in shallow basins between latest Devonian-earliest Mississippian tectonic elements. Rocks of the Cottonwood Canyon Member occur in two facies tracts—an eastern dolomitic shale and siltstone facies deposited in shallow-water marine environments at the rapidly advancing margin of the Madison sea, and a western siltstone and shale facies that accumulated in slightly deeper offshore marine environments (Sandberg and Klapper, 1967; Rodriguez and Gutschick, 1970). Clastic sediments for these facies were probably provided by low-lying uplifts adjacent to this shallow basin.

Continued rapid transgression in the late Kinderhookian resulted in the deposition in shallow agitated water of the widespread bioclastic and glauconitic limestone at the base of the Paine Member as low-lying terrigenous sediment sources were progressively inundated and effectively mantled. Sedimentation apparently kept pace with the combined effects of downwarping of the incipient Big Snowy trough and rising sea level.

As rapid transgression of the late Kinderhookian Madison sea continued to its latest Kinderhookian maximum, downwarping of the Big Snowy trough and (or) rise of sea level exceeded rates of carbonate and terrigenous sediment influx, resulting in deeper water environments there. Along the slope of this trough, lime-mud bioherms were raised by the combined activities of crinoids and bryozoans from the shallow sea floor into shallower water.

At this same time, shallow-water carbonate sediments were deposited on the Wyoming shelf (Sando, 1972). However, between the Big Snowy trough and the Wyoming shelf, there was no abrupt break-in-slope and attendant marginal carbonate buildups.

During the early Osagean, regression of the Madison sea began and subsidence of the Big Snowy trough diminished. In the trough, deeper water conditions that prevailed during deposition of the Paine Member gave way to shallower depositional environments, resulting in the cessation of bioherm growth and an increase in the bioclastic content of the limestones that overlie the dark rhythmic lime mudstone beds. On the Wyoming shelf, and probably on the Alberta shelf, carbonate production and accumulation exceeded subsidence, resulting in northward and southward progradation of oolitic and bioclastic shoals of the Woodhurst Member across the Big Snowy trough.

Widespread intertidal and subtidal environments prevailed throughout Montana during the first half of Osage time, and bioclastic and oolitic shoal and finer grained intershoal carbonate sediments accumulated. Alternate high- and low-energy shallow-water lithotopes migrated across the State in response to minor transgressions and regressions (Sando, 1976) or to some combination of eustatic change, varying rates of carbonate production, and differential subsidence (Smith, 1977). The migration of these shoal and intershoal lithotopes produced the cycles that characterize the Woodhurst Member.

MISSION CANYON LIMESTONE

The Mission Canyon Limestone is a prominent cliff- and ridge-forming limestone above the less resistant Lodgepole Limestone and below the recessive Big Snowy and Amsden Groups. The formation is present throughout the State, except where it has been removed by Tertiary erosion. However, the original thickness and lithologic distribution is masked by Triassic and Early Jurassic erosional thinning in the northern part of Montana.

The Mission Canyon Limestone conformably overlies the Lodgepole Limestone and is, in turn, overlain by various upper Paleozoic and Mesozoic strata. In the Big Snowy trough, the Mission Canyon is unconformably overlain by the Big Snowy Group. However, to the north, it is unconformably overlain by the Jurassic Sawtooth and Nesson Formations. South of the Big Snowy trough, the Mission Canyon is unconformably overlain by the Pennsylvanian Amsden Formation, the contact marked by a regional karst surface (Roberts, 1966; Sando, 1974).

The Mission Canyon Limestone is thickest in the Big Snowy trough and Williston basin (300 m), thinning northward because of Triassic and early Jurassic erosion and southward because of deposition. This lenticular geometry reflects the relative negative and positive aspects of these paleotectonic elements both during and after deposition of the Mission Canyon Limestone.

The Mission Canyon Limestone is characterized by less detrital sediment than is the underlying Lodgepole; it is composed of massive beds of limestone and dolomite, the percentage of dolomite in the formation generally increasing southward onto the Wyoming shelf (Andrichuk, 1955). Interbedded with these carbonate rocks in the upper part of the formation are beds of gypsum or anhydrite in the subsurface that are manifest in surface sections as solution breccia zones (Roberts, 1966). Evaporite beds are more abundant in the Big Snowy trough and Williston basin than they are on the northern and southern shelves (Craig, 1972).

On the shelf in southern Montana, Sando (1972) delineated three members of the Mission Canyon: the lower limestone member, the cliffy limestone member, and the Bull Ridge Member, in ascending order. The lower limestone member ranges from 85 to 100 m in thickness and is the time-stratigraphic equivalent of the cherty dolomite member of Wyoming (Sando, 1972). The base of this member consists to 15 to 30 m of cross-stratified, oolitic, and crinoidal limestone. This basal unit is overlain by interbedded limestone and dolomite, and the top of the member is placed at the base of a widespread evaporite solution breccia. The cliffy limestone member includes this 7.5 to 15-m-thick breccia at its base. This breccia bed appears in most sections of southwestern Montana and is the "lower solution zone" of Sando (1972). Above this breccia, the cliffy limestone member consists of 55 to 70 m of cliff-forming oolitic and bioclastic limestone and dolomitized limestone. The Bull Ridge Member is 3 to 36 m thick, and its base is characterized by an evaporite solution breccia or brecciated dolomitic siltstone and shale interval 3 to 7.5 m thick, the "upper solution zone" of Sando (1972). Above this breccia, the member consists of cherty bioclastic limestone. Brecciation in the Bull Ridge Member is common; red sand, clay, and silt from Pennsylvanian erosion and deposition fill cavities and sinkholes (Sando, 1972).

Although these three members have not been extended out of the Beartooth Mountains area, the lower breccia zone has been extended to the type section at Logan, as well as to the Monarch section in the Big Snowy Mountains (Sando and Dutro, 1974). The upper breccia zone has been correlated only through the Beartooth Mountains. Because it is roughly coincident with the Osagean-Meramecian boundary, it is a good correlation horizon (Sando, 1972). However, in both the Logan and Monarch sections, this breccia may be absent because of nondeposition and is absent in the type section of the Mission Canyon because of pre-Jurassic erosion.

In its type area in the Little Rocky Mountains, the Mission Canyon section was abbreviated by pre-Jurassic erosion, and the 90 m of section there is considered to be the time-equivalent of the lower two-thirds of the lower limestone member in the Beartooth Mountains (Sando and Dutro, 1974, pl. 1). In the Little Rockies, the Mission Canyon consists of medium- to coarse-grained crinoidal limestone in beds reaching a maximum thickness of 1.5 m. These coarse-grained beds are interbedded with finer grained limestone. Both limestone types contain lentils and nodules of chert that may constitute as much as 20 percent of the section.

In northwestern Montana, Mudge, Sando, and Dutro (1962) and Mudge (1972) used the term Castle Reef Dolomite for time-stratigraphic equivalents of the Mission Canyon Limestone. This unit was divided into a lower member and the Sun River Member. The lower member is 116 to 156 m of thick-bedded, fine to coarsely crystalline dolomite and limestone. Many of the coarsely crystalline limestone beds are crinoidal and are cross stratified. The Sun River Member ranges from 76 to 100 m in thickness and correlates with the upper part of the Mission Canyon in the Three Forks area (Mudge, 1972). It contains thin to thick beds of very fine to medium crystalline dolomite interbedded with thick lenses of dolomitized crinoidal limestone. The upper part of the Sun River Dolomite contains sandstone lenses that are interpreted as Jurassic cave fillings, but the karst surface that is so extensive over the southern part of the State is absent here (Mudge, 1972).

The Charles Formation has been the subject of controversy since Seager (1942, p. 863) applied the name to a "series of interbedded limestones, dolomite, anhydrite, and some shales" in a well in central Montana (fig. 1). Sloss (1952, p. 67) described the Charles of the Williston basin as "three major evaporite cycles which include normal fossiliferous limestones, sugary dolomites, dense dolomites, and anhydrite in upward succession." The base of the Charles is usually picked at the base of the lowest evaporite; the top is the unconformable contact with the Kibbey Formation of the Big Snowy Group. Attempts have been made to identify the Charles Formation in outcrops of central Montana but "it is defined on criteria that are difficult to use with precision in outcrop areas" (Sando and Dutro, 1974, p. 2). Until definitive stratigraphic work has refined knowledge of the relationship between the Charles and the Mission Canyon, the Charles is probably best considered as the subsurface lithic and temporal equivalent of the upper part of the Mission Canyon Formation (Sloss, 1952), and the use of the term should be restricted to the subsurface of the Williston basin (Sando and Dutro, 1974).

The Mission Canyon Limestone and its equivalents include three and part of two more foraminiferal zones and two coral-brachiopod megafaunal zones (fig. 7). Both biostratigraphic schemes give the formation a middle Osagean to early Meramecian age (Sando, Mamet, and Dutro, 1969).

The depositional history of the Mission Canyon Limestone and its equivalents was summarized by Andrichuk (1955) and recently by Rose (1976) and Sando (1976). These summaries suggested that Mission Canyon carbonate and evaporite strata record the regression of the Madison sea from the craton in Montana. As shallowing and exposure progressed, the shelves became the sites of sabkhas and salt pans in which evaporites were deposited; dolomitizaton of intertidal and subtidal limestone took place at the same time. Progressive shallowing of the Madison sea also restricted circulation, which led to the accumulation of thick evaporite beds in the Big Snowy trough and Williston basin. Periodic freshening of the sea by normal marine water, related to transgressive pulses of the major regression or to differential sedimentation and epeirogenic warping, probably accounted for the carbonate-evaporite cycles that characterize both basins and shelves.

After its early Meramecian restricted phase, the Madison sea withdrew from most of the craton in Montana, exposing the upper part of the Mission Canyon and its equivalents to subaerial weathering and subsequent karst formation prior to the later Meramecian and earliest Chesterian transgression of the Big Snowy sea.

BIG SNOWY GROUP

The lithology, stratigraphy, and paleontology of the Big Snowy Group have been described by Scott (1935), Walton (1946), Mundt (1956a, b), Willis (1959), Easton (1962), Maughan and Roberts (1967), Harris (1972), Craig (1972), Jensen and Carlson (1972), and Sando, Gordon, and Dutro (1975). The least known but probably the most detailed of these studies is that of Harris (1972), who measured 58 Big Snowy sections in central Montana and who also used existing subsurface data in his synthesis. Much of the following is from Harris' unpublished work.

The Big Snowy Group is restricted to the Big Snowy trough in central Montana, the Williston basin in eastern Montana, and an extension of the Big Snowy trough in southwestern Montana (fig. 6). In central Montana, the Big Snowy trough was bordered on the north by the "Milk River uplift" of Maughan and Roberts (1967) and on the south by the southern Montana arch (Sando, 1976). The group is thickest (360 m) along the axis of the Big Snowy trough; it thins abruptly north and south away from the trough axis. To the north, this thinning is due, in part, to pre-Jurassic erosion (Maughan and Roberts, 1967). South of the trough, the thinning is due to a combination of depositional thinning and latest Chesterian erosion (Sando, Gordon, and Dutro, 1975; Sando, 1976).

In the Big Snowy trough of central Montana and the Williston basin, the Big Snowy Group consists of three formations—the Kibbey Sandstone and the Otter and Heath Formations, in ascending order (fig. 7). However, in southwestern Montana, where these members have not been formally delineated, the Big Snowy has formational rank. Rocks of the Big Snowy Group are interpreted as products of the second major Mississippian transgression in Montana. The Kibbey is the basal transgressive unit, deposited at the eastward-advancing margin of the Big Snowy sea; the Otter consists of shale and limestone, deposited just offshore from the Kibbey; and the Heath is an accumulation of dark limestone and shale along the axis of the Big Snowy trough (Sando, Gordon, and Dutro, 1975). These formations are a classic diachronous transgressive sequence (fig. 7), older in southwestern Montana (latest Meramecian), progressively younger eastward in the Big Snowy trough of central Montana (earliest Chesterian), and youngest in the Williston basin of eastern Montana (middle Chesterian) (Sando, 1976).

KIBBEY SANDSTONE

The contact of the Kibbey Sandstone with the underlying Madison Group is reported to be conformable in eastern Montana and in the center of the Big Snowy trough in central Montana (Maughan and Roberts, 1967; Harris, 1972), where it is said to intertongue with or conformably overlie the Charles Formation or its Madison Group equivalent. However, these same workers stated that the Big Snowy Group rests unconformably on Madison strata in southwestern Montana, as well as in the southern part of the Big Snowy trough in central Montana. These interpretations suggest that during the regressive phase of the latest early Meramecian, the Madison sea regressed from the central and western parts of the Big Snowy trough; it remained as an isolated marine body in the more negative eastern parts of the trough and Williston basin, producing the Madison-Big Snowy unconformity in western and central Montana and the conformable and intertonguing relationships between these groups in the eastern trough and Williston basin.

Counter to this view is the concept that the Madison sea retreated from the Big Snowy trough and the Williston basin, as well as from the surrounding shelf areas during the latest early Meramecian (Sando, Gordon, and Dutro, 1975; Sando, 1976), producing a major disconformity between the Madison and Big Snowy Groups. In support of this concept, Sando (1978) identified megafaunal zone D corals 15 to 20 m below the top of the Charles Formation in the subsurface of the Williston basin. Additional faunal evidence bearing on this problem was provided by Scott (1945) and Easton (1962), who indicated that faunas from the Otter Formation are Chesterian. Inasmuch as the age of the intervening Kibbey Formation has not been definitely established by fossil evidence, at least three and possibly four megafaunal zones are absent at the Madison-Big Snowy contact. The absence of these fossil zones, the classic transgressive stratigraphic and sedimentologic sequence of the Big Snowy Group, the paleogeographic problem presented by an isolated marine body in a cratonic basin, and similar stratigraphic relationships of the Darwin Sandstone in Wyoming, all strongly suggest the presence of an interregional disconformity between the Madison and Big Snowy Groups.

The Kibbey Formation is thickest along the northern edge of the Big Snowy trough (76 m), thinning abruptly to the north and south. In southwestern Montana, the Kibbey equivalent is 45 m thick. Where the formation is thickest and most extensive in central and eastern Montana, Harris (1972) delineated three informal members. The lower member is 0 to 30 m thick and consists of red shale containing beds and lenses of green shale, sandstone, and gypsum. The middle member is 1.5 to 12 m thick and occurs only in the middle and eastern parts of the Big Snowy trough. In the central Montana part of the trough, this member is composed of dolomite containing thick gypsum beds. However, in the eastern part of the trough, this dolomite grades into oolitic and fragmental limestone (Rawson, 1968). The upper member of the Kibbey consists of fine-grained quartz sandstone and interbedded red and gray shale and lenses of dolomite. This member is 46 m thick in the axis of the Big Snowy trough, where shale is the predominant rock type. At the southern margin of the trough, sandstone is dominant.

The Kibbey Formation is interpreted as intertidal to subtidal deposits at the leading edge of the advancing Big Snowy sea. During deposition of the lower member, sand from the craton accumulated along high-energy shorelines at the same time as finer-grained detritus was deposited in intertidal and shallow subtidal environments. Thick beds of gypsum interbedded with these clastic rocks indicate that circulation of shallow marine water in the Big Snowy trough must have been spatially and temporally restricted, probably by a combination of the irregular topography of the eroded Madison surface, minor transgressive-regressive fluctuations of the Big Snowy sea, and gentle epeirogenic activity in the trough. Similar environmental controls and configurations prevailed during deposition of the middle member of the Kibbey, but the supply of coarse clastic debris from the craton had diminished and parts of the Big Snowy trough were less restricted, resulting in the accumulation of evaporite and dolomite in central and western Montana and oolitic and bioclastic limestone in the eastern Big Snowy trough and Williston basin. This facies pattern of rocks of hypersaline origin in the western part of the trough and normal marine limestone in the eastern part of the trough and the Williston basin presents a paleogeographic problem concerning the source of the water for this marine embayment. Rawson (1968) suggested that normal marine water fed into the Williston basin from a northwestern link with the Cordilleran miogeosyncline. An equally strong argument may be made for nonrestricted flow of normal marine water through the Big Snowy trough from the Cordilleran miogeosyncline on the west to the Williston basin on the east, peripheral shallow restricted embayments providing sites for evaporite accumulation at the margins of the trough. The upper member of the Kibbey records an influx of coarser clastic material from the Canadian shield (Ballard, 1964; Harris, 1972), sand that was deposited in high-energy intertidal shoreline environments near the margins of the Big Snowy trough. However, shale dominates this member in the central part of the trough, suggesting deeper quieter depositional conditions there.

Throughout most of Montana, the Kibbey grades into the Otter through 3 to 6 m of intertongued typical Kibbey sandstone beds and light-gray shale typical of the Otter. However, in parts of southwestern Montana, the Kibbey and Otter are unconformably overlain by the Tyler Formation equivalent (Sando, Gordon, and Dutro, 1975).

OTTER FORMATION

The Otter Formation or its equivalents occur throughout the Big Snowy trough, the Williston basin, and southwestern Montana (Sando, Gordon, and Dutro, 1975). Along the trough axis, the formation is 150 m thick, but it thins abruptly to a featheredge to the north and south. In central Montana, Harris (1972) divided the Otter into three informal and unnamed members. The lower member is 75 m thick along the trough axis and consists predominantly of gray and dark-gray shale and interbedded green and maroon shale, thin sandstone lenses and beds, dolomite, and thin beds of stromatolitic and oolitic limestone. Green shale and dolomite predominate along the southern margin of the trough, and dark-gray shale and limestone are dominant in the trough axis. The middle member of the Otter is 35 m thick in the Big Snowy trough and is predominantly oolitic, pelletal, bioclastic, stromatolitic, and micritic limestone containing minor beds of green to dark-gray shale, cherty dolomite, and gypsum. The darker rocks generally occupy the center of the trough; the green shale and dolomite are found at the southern trough margin. The upper member is 90 m of green shale containing lenses of variegated calcareous and siliceous shale and beds of chert-rich, stromatolitic, oolitic, and bioclastic limestone. These light-colored rocks dominate the upper member along the trough margins. The axial part of the trough is characterized by dark shale and interbedded limestone.

The carbonate-rich Otter Formation records a change from the coarse clastic deposits of the Kibbey Formation to more normal marine, more agitated, but sand-free depositional conditions. The lower member was deposited in intertidal and shallow subtidal environments, where fine-grained clastic sediment, ooliths, calcareous skeletal debris, and lime mud were synchronously deposited under variable environmental conditions. Concentration of green shale and dolomite along the southern trough margin suggests oxidizing and, possibly, supratidal environments there, close to the source of terrigenous clastic sediment. The predominant gray shale and limestone of the lower member suggest deposition in agitated and calm subtidal environments toward the trough axis. In the center of the Big Snowy trough, deeper water environments favored accumulation of dark shale.

The Otter-Heath contact is gradational through a 15- to 30-m sequence of interbedded black Heath-like shale and limestone and typical green shale of the Otter Formation. However, in parts of southwestern Montana, the Otter is unconformably overlain by the Tyler equivalent, and in the northern part of the Big Snowy trough, by the Tyler Formation, the Quadrant Formation, or the Ellis Group.

HEATH FORMATION

Along the axis of the Big Snowy trough, the Heath is 120 m thick, thinning abruptly toward the northern and southern trough margins. In southwestern Montana, the Heath is 45 m thick; dark-gray to black shale and limestone are dominant. The dark shale is fissile, petroliferous, and is the predominant Heath lithology in the middle of the Big Snowy trough. Heath limestone beds are massive, micritic, sparsely fossiliferous, and may have sparse chert nodules or beds and scattered quartz sand grains. Limestone is the dominant Heath lithology at the southern margin of the trough.

Heath shale accumulated in calm, oxygen-poor environments in the center of the Big Snowy trough. Depositional environments at the margins of the trough, however, were shallower, more agitated, and were conducive to the accumulation of lime mud and well-washed bioclastic debris, both containing admixtures of quartz sand from a cratonic source (Harris, 1972).

Relief of as much as 45 m on the unconformity between the Tyler and the Heath indicates that extensive erosion took place in the Big Snowy trough after deposition of the Heath. At the southern trough margin, as well as in southwestern Montana, the Tyler or its equivalent unconformably overlies Kibbey, Otter, and Heath equivalents (Sando, Gordon, and Dutro, 1975). At the northern margin of the Big Snowy trough, rocks of the Jurassic Ellis Group unconformably overlie and truncate all three formations of the Big Snowy Group (fig. 9).

AMSDEN GROUP

The Amsden Group as defined by Maughan and Roberts (1967) consists of three formations—the Tyler Formation, Alaska Bench Limestone, and Devils Pocket Formation (fig. 2). This terminology is used for central and eastern Montana. Maughan and Roberts (1967) arbitrarily used Amsden Group terminology eastward to the Montana-North Dakota border, at which point they used Minnelusa Formation. In southwestern Montana, strata approximately equivalent to the Amsden Group were referred to as the Amsden Formation (Maughan and Roberts, 1967).

The Amsden Group ranges from a featheredge to more than 270 m in thickness in central Montana (Mallory, 1972). More commonly, the group is about 150 m thick through the center of the Big Snowy

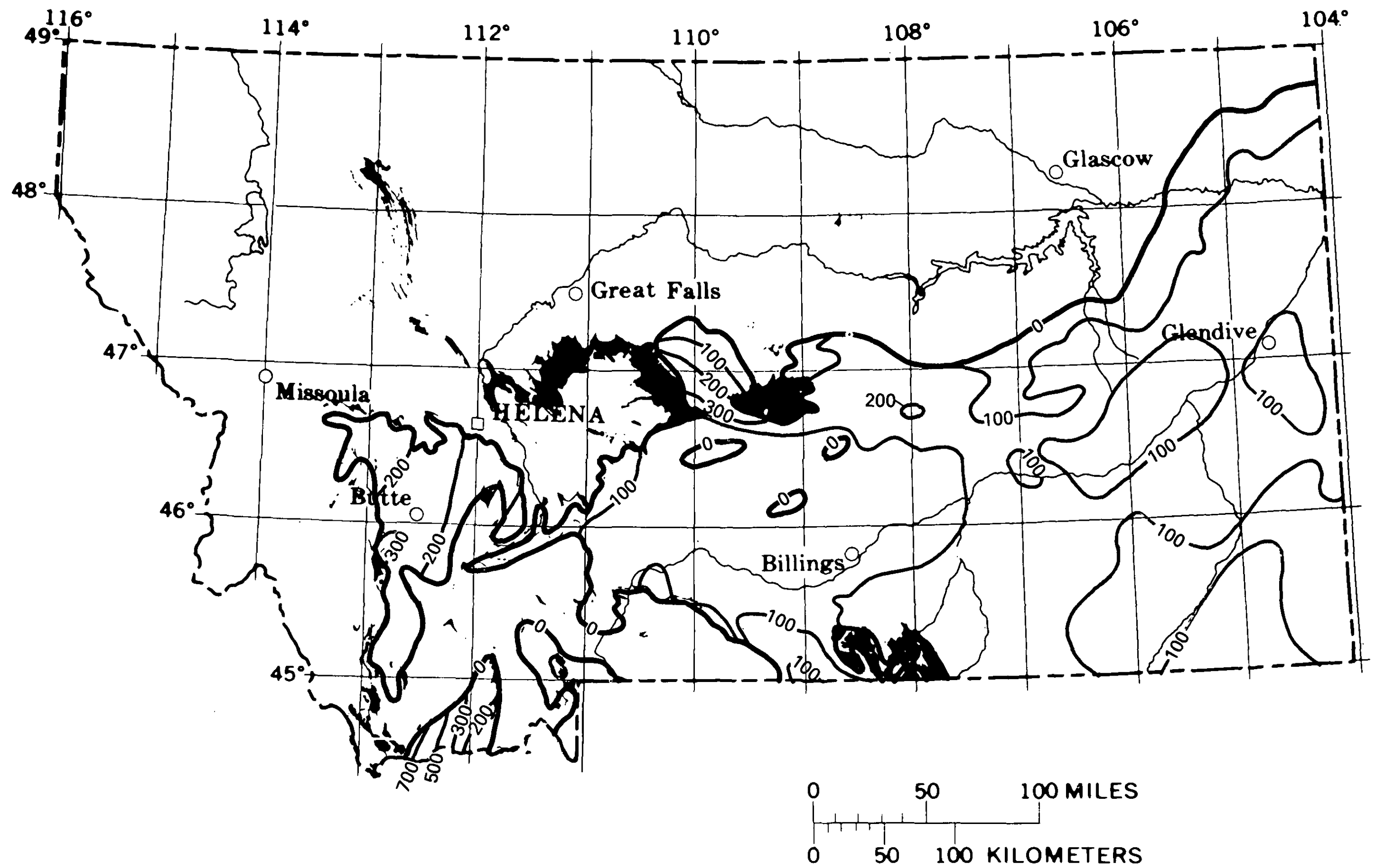


FIGURE 8.—Map of total thickness of the Pennsylvanian System (adapted from McKee and Crosby, 1975, pl. 11). Thicknesses include Stonehouse Canyon Member of the Tyler Formation, part of which may belong in the Mississippian System. Isopach thicknesses in meters.

trough. The group extends from southwestern Montana northeastward through the Big Snowy trough to the Montana-North Dakota border (fig. 8). Northward, pre-Jurassic erosion has truncated the group (Maughan and Roberts, 1967); southward, the group thins over the southern Montana arch, where it grades into the Ranchester Limestone Member of the Amsden Formation (Sando, Gordon, and Dutro, 1975). Lithologies present are sandstone, siltstone, shale, limestone, dolomite, sandy dolomite, and a few dolomite beds with chert nodules.

The Amsden Group grades upward from dark-gray shale and sandstone to red beds to interbedded limestone, shale, and dolomite to sandy dolomite and sandstone. This vertical relationship is similar to that described by Sando, Gordon, and Dutro (1975, p. A3–A7) for the Amsden Formation in Wyoming. Lateral gradations of rocks between and within formations have been reported by several workers (Gardner, 1959; Maughan and Roberts, 1967; Gilmour, 1969; Sando, Gordon, and Dutro 1975). Sando, Gordon, and Dutro (1975, p. A64) postulated an east-west facies relationship across Montana of a nearshore sand belt, an intermediate offshore lagoonal facies, and a central dolomitic carbonate-shale facies.

TYLER FORMATION

The lowest unit of the Amsden Group is the Tyler Formation, named by Freeman (1922) for exposures in the Big Snowy Mountains (fig. 1). The Tyler Formation was redefined by Maughan and Roberts (1967) as consisting of two members: a lower member called the Stonehouse Canyon Member and an upper member called the Cameron Creek Member. The two members were established "largely on color and partly on lithology" (Maughan and Roberts, 1967, p. B12).

In Easton's (1962) section at Stonehouse Canyon, designated as a reference section and reinterpreted by Maughan and Roberts (1967, p. B12), the Stonehouse Canyon Member consists of 31 m of covered black fissile shale and poorly exposed dark-greenish-gray to very dark brownish-gray shale. At the type section given by the authors, the member is 76 m thick and consists of buff to brown sandstone (21 m), black and dark-gray fissile shale (53 m), and light-gray to buff limestone (2 m). Both these sections were included in the Heath Formation by Easton (1962).

The Cameron Creek Formation was originally named by Gardner (1959) and was reduced in rank to member status by Willis (1959) and again by Maughan and Roberts (1967) (fig. 2). This unit comprises the varicolored shale, gray and brownish sandstone, and thin gray limestone above the sandstone and dark shale of the Stonehouse Canyon Member and below the Alaska Bench Limestone. The two members are lithologically similar, as pointed out by Maughan and Roberts (1967, p. B12): "The boundary between Stonehouse Canyon and Cameron Creek is difficult or impossible to pick consistently at the same stratigraphic position from place to place owing to the gradation and intertonguing of one into the other." At the west end of Alaska Bench (locally called "Beacon Hill"), the Cameron Creek Member is 25 m thick (Easton, 1962, p. 117) and at the Stonehouse Canyon section, it is 67 m thick (Easton, 1962, p. 123).

The extent of the Tyler Formation approximates the same area defined as the Big Snowy trough. It generally thins eastward towards the Cedar Creek anticline in eastern Montana but thickens again eastward into the Williston basin area after crossing the anticline. The formation also thins southward across the southern Montana arch toward Wyoming. Westward, the Tyler Formation or its equivalent extends into the Cordilleran miogeosyncline. Total thickness for the Tyler Formation ranges from a featheredge to more than 242 m near the Little Belt Mountains in central Montana. Generally, the formation is 30 to 90 m thick.

Vertically, the formation changes from dominantly sandstone and black shale at the base to sandstone and red beds in the middle to interbedded limestone and red shale at the top in central Montana. The percentage of sandstone in the lower part (Stonehouse Canyon Member) increases westward toward southwestern Montana; the percentage of limestone in the upper part (Cameron Creek Member) increases eastward toward the Williston basin (Mallory, 1972, p. 112).

Most of the fossils found in the Tyler Formation indicate an Early Pennsylvanian (Morrowan) age. Several detailed reports on the fauna and flora of the Tyler Formation have been published by Easton (1962), Maughan and Roberts (1967), Gordon (1975), and Sando, Gordon, and Dutro (1975). Easton (1962) interpreted the fauna in the Tyler Formation (Cameron Creek Formation of Easton) as Late Mississippian (Chesterian). Sando, Gordon, and Dutro (1975) listed three brachiopods from the Tyler Formation that are exclusively Mississippian in Wyoming and two species that are Mississippian and Pennsylvanian in Wyoming. Exclusively Mississippian are *Pugnoides quinqueplecis* Easton, *Anthracospirifer curvilateralis curvilateralis* (Easton), and *A.* cf. *A. occiduus* (Sadlick) form A. *Schizophoria depressa* Easton and *Eolissochonetes pseudoliratus* (Easton) are found in both Mississippian and Pennsylvanian rocks. Sando, Gordon, and Dutro (1975) also listed six species of brachiopods in the Tyler Formation that are exclusively Pennsylvanian. These include *Orthotetes* sp. A. Gordon, *Echinoconchus* sp. A. Gordon, *Antiquatonia* cf. *A. coloradoenis* (Girty), *Linoproductus eastoni* Gordon, *Composita ovata* Mather, and *Anthracospirifer occiduus* (Sadlick). *Petrocrania chesterenis* (Miller and Gurley) was collected near the top of the Stonehouse Canyon Member and is regarded as Chesterian, according to Gordon (Maughan and Roberts, 1967, p. B20). This brachiopod occurs with *Eolissochonetes pseudoliratus* (Easton), considered by Sando, Gordon, and Dutro (1975) as Mississippian and Pennsylvanian. This type of information tends to support the contention of Easton (1962, p. 25) ". . . that the fauna of the Big Snowy group [Easton included the Cameron Creek Formation in the Big Snowy Group] may prove to be particularly significant in subsequent attempts to recognize the Mississippian-Pennsylvanian boundary, because the fauna occurs in the critical interval and is replete with species."

Maughan and Roberts (1967, p. B12–B14) gave particular attention to the age of the Tyler Formation and concluded that the Tyler Formation above the regional unconformity with the Heath Formation is Morrowan. They reported Early Pennsylvanian plant spores from the upper part of the Stonehouse Canyon Member and extended this age assignment to the base of the Tyler. However, they (Maughan and Roberts, 1967, p. B21) also quoted R. H. Tschudy as saying that spores from the lower part of the Stonehouse Canyon Member "may be from a Pennsylvanian horizon not yet examined, or may represent a transitional flora between the Late Mississippian and the Early Pennsylvanian." Mallory (1972) also described the Tyler Formation as Morrowan in age.

The Cameron Creek Member contains several brachiopods that are restricted to Morrowan or younger rocks (Maughan and Roberts, 1967). These include *Linoproductus eastoni* Gordon, *Rugoclostus nivalis* Easton, and *"Marginifera" planocosta* Easton. Easton (1962, pl. 3) illustrated *Millerella* collected from the Cameron Creek Member. According to B. A. Skipp, this form "may be considered Pennsylvanian as much as they may be Mississippian forms" (Maughan and Roberts, 1967, p. B21).

From the foregoing discussion, it appears that the upper part of the Stonehouse Canyon Member and the Cameron Creek Member are Morrowan on the basis of the faunal evidence. However, the exact age of the lower part of the Stonehouse Canyon Member (upper part of Heath Formation of Easton, 1962), as redefined by Maughan and Roberts (1967), is still open to question, and additional work is needed.

Both marine and nonmarine environments of deposition have been proposed for the Tyler Formation in central and eastern Montana (Mundt, 1956b; Gardner, 1959; Willis, 1959; Foster, 1961; Ballard, 1964; Maughan and Roberts, 1967; Jensen and Carlson, 1972). The Stonehouse Canyon Member is believed to have been deposited on an erosional surface of the Mississippian Heath Formation (Beekly, 1955). The interbedded sandstone and dark-gray to black shale of the Stonehouse Canyon Member are lagoonal, deltaic, and estuarine deposits varying from marine to nonmarine (Beekley, 1955). Mundt (1956b) interpreted the dark gray and black shale in the lower Tyler as nonmarine and the erratic sands in the lower part of the formation as channel deposits. He believed that "a gradation upward from nonmarine to normal marine shale apparently coincides with the color change" from dark gray to reddish shades.

According to Gardner (1959, p. 337, 344), the Tyler Formation represents deposition in a shallow marine basin where currents of water, flowing seaward from tributary rivers, cut local channels across tidal flats and shallow parts of the sea. These channels were filled with sand, as well as debris torn from the sides and bottom of the channel. When changes in current velocity or positions of currents took place, typical marine sediments could be interbedded with channel deposits. Gardner (1959) presented evidence opposing a nonmarine interpretation of the paleontological data. Thus, Gardner supported an interpretation that both the Heath and the Tyler Formation are part of a continuous marine sequence.

THE BEAR GULCH LIMESTONE PROBLEM

Mundt (1956a, b) included a marine limestone tongue in the Tyler Formation that is locally present near the top of the formation. This limestone was shown in a diagrammatic correlation section (Mundt, 1956a, p. 1925). Mundt (1956c) also measured and described a section designated as "Bear Gulch composite section" that contained a detailed description of the Tyler Formation and the included limestone tongue. Willis (1959, p. 1953) referred to this limestone as the Bear Gulch limestone tongue for exposures along Bear Gulch Creek, south of Forest Grove. Other workers (Foster, 1956, p. 122; Norton, 1956, p. 58, 62; Todd, 1959) also made reference to this limestone tongue.

All the above workers placed the Bear Gulch limestone tongue in some part of the Tyler Formation and above the regional unconformity between the Heath and Tyler described by Maughan and Roberts (1967). Maughan and Roberts (1967, p. B11, fig. 5) also showed an extensive limestone tongue in the middle of the Tyler Formation that thins from north to south (fig. 9). However, they did not discuss the lithology, age, or extent of the limestone unit. Mundt (1956a, p. 1924) depicted a depositional model for the Tyler Formation showing sandstone deposited simultaneously with the Bear Gulch limestone tongue, both of which are above the regional unconformity between the Heath and Tyler.

On the basis of studies of conodonts and fish (W. G. Melton and J. R. Horner, written commun., 1978), the age of the Bear Gulch limestone tongue is believed to be Springeran (latest Mississippian). This age may be equivalent to that of foraminiferal zone 19, which was included in the Chesterian by Sando, Gordon, and Dutro (1975). Mackenzie Gordon, Jr. (written commun., 1978), in 1970 studied the cephalopod fauna of the Bear Gulch limestone tongue at the Allen Surprise quarry in Fergus County and concluded that because it contained *Epistroboceras, Tylonautilus,* and *Anthracoceras,* it was of late Chesterian (Late Mississippian) age. This information throws doubt on the interpretation that the unconformity between the Heath and the Stonehouse Canyon Member of the Tyler Formation represents the Mississippian-Pennsylvanian boundary. The position of the systemic boundary appears to be somewhere in the Stonehouse Canyon-Cameron Creek sequence (fig. 9).

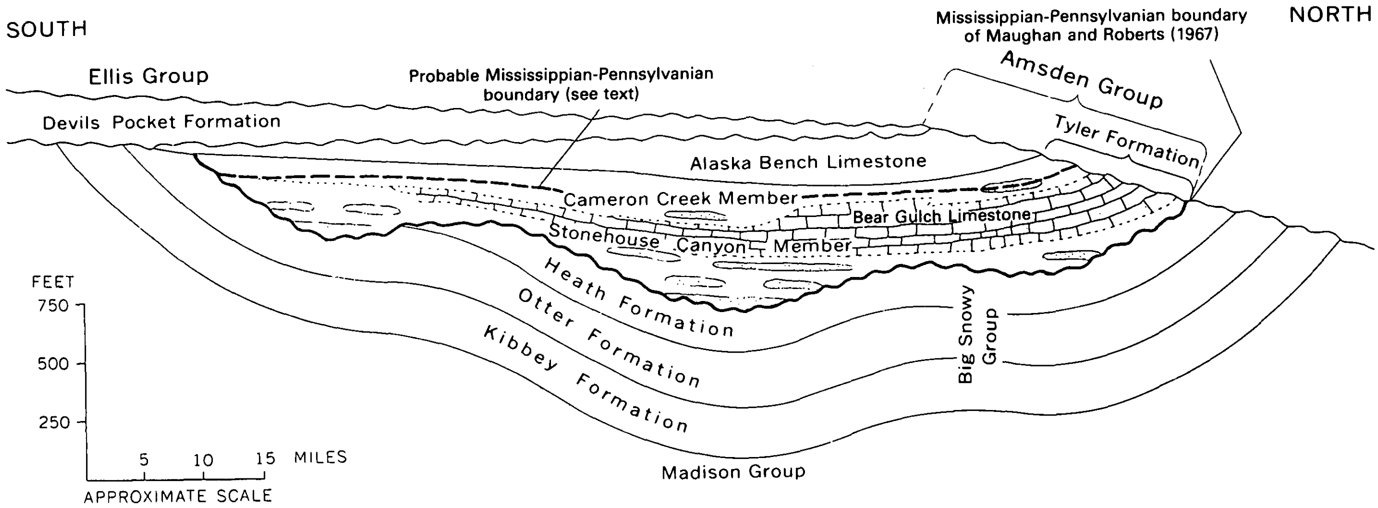


FIGURE 9.—Diagrammatic cross section of Big Snowy and Amsden Groups in central Montana. (Modified from Mundt, 1956a, and Maughan and Roberts, 1967.)

ALASKA BENCH LIMESTONE

Freeman (1922) was the first to apply the name Alaska Bench Limestone to the gray fossiliferous limestone that forms the sloping benches and hogbacks around the Big Snowy Mountains (fig. 1). Mundt (1956a), Gardner (1959), Easton (1962), Maughan and Roberts (1967), and Gilmour (1969) used the name as a formational designation. Only the limestone in the Big Snowy trough is called the Alaska Bench, although Willis (1959) stated that the lower limestone of the Minnelusa Formation in the Williston basin is probably equivalent to the Alaska Bench. Carbonate units at the same interval in southwestern Montana constitute the upper part of the Amsden Formation (Maughan and Roberts, 1967). In Wyoming and southern Montana, limestone of similar age belongs to the Ranchester Limestone Member of the Amsden Formation (Sando, Gordon, and Dutro, 1975).

The Alaska Bench Limestone thins northward toward the depositional edge of the Big Snowy trough (Gilmour, 1969). The formation is also truncated northward by pre-Jurassic erosion. This combination of depositional thinning and postdepositional erosion accounts for the northward thinning and truncation of the Alaska Bench. The formation is truncated southward, where it was probably eroded from the Montana uplift of Sando, Gordon, and Dutro (1975) during Atokan time.

Thickness of the formation varies considerably over short distances owing to nondeposition and periods of postdepositional erosion. Maximum thicknesses reported are 43 m at Durfee Creek Dome (Easton, 1962), 40 m at Beacon Hill (Gilmour, 1967), and 88 m at Judith Gap (Maughan and Roberts, 1967). The formation thins eastward toward the Williston basin, where it does not exceed 38 m.

Lithologically, the Alaska Bench consists of interbedded gray limestone, red mudstone, and dolomite. Limestone beds generally range from 0.3 m to 0.6 m in thickness, but beds as thick as 1.5 m do occur. Beds of red mudstone 0.3 to 1.5 m thick occur throughout the formation (Gilmour, 1967). Dolomite beds are also found throughout the Alaska Bench. Maughan and Roberts (1967) reported a greater proportion of carbonate in the lower 30 m of the Alaska Bench and equal amounts of carbonate and mudstone in the thicker sections (Judith Gap). In the North Fork of Flat Willow Creek, the Alaska Bench Limestone is 25 m thick and contains 33 percent mudstone (Gilmour, 1967, pl. 20). The lower 10 m of the Stonehouse Canyon section is 50 percent mudstone (Gilmour, 1967, pl. 3).

Contacts between the Alaska Bench Limestone and the underlying Cameron Creek Member of the Tyler Formation are gradational with interbedded gray limestone and red mudstone. Lateral intertonguing between these two lithologies was mentioned by Maughan and Roberts (1967) and discussed in detail by Gilmour (1967). Evidence for an unconformity between the Alaska Bench Limestone and the overlying Devils Pocket Formation was presented by Mundt (1956a) and further supported by Maughan and Roberts (1967, p. B15). However, the use of a limestone-dolomite contact between the two formations as the unconforimity is open to question because of secondary dolomitization downward into the Alaska Bench Limestone.

Fossils compose a large percentage of the limestone in the Alaska Bench Limestone. Occurrence of specific fossils or fossil debris depends on which microfacies is represented by a particular limestone bed (Gilmour, 1969). Ostracodes are particularly abundant, as they occur in most of the microfacies described, especially in the algal biolithites and ostracodal muds. Brachiopods, bryozoans, pelmatozoans, echinoids, opthalmid Foraminifera, gastropods, and millerellids are major components in the normal marine microfacies.

Six species of brachiopods were listed by Sando, Gordon, and Dutro (1975) as present in the Alaska Bench Limestone and exclusively Pennsylvanian in Wyoming. These species are: *Antiquatonia* cf. *A. coloradoensis* (Girty), *Linoproductus eastoni* Gordon, *Composita ovata* Mather, *Anthracospirifer occiduus* (Sadlick), *Orthotetes* sp. A. Gordon, and *Echinoconchus* sp. A. Gordon.

Scott (1945) reported fusulinids from the lower part of the Alaska Bench Limestone near Beacon Hill, identified as *Millerella marblensis* Thompson and *Millerella advena* Thompson. Easton (1962) identified *Dicromyocrinus granularis* Easton from the Alaska Bench.

Most paleontological evidence supports a Morrowan age for the Alaska Bench Limestone (Scott, 1945; Gilmour, 1967; Sando, Gordon, and Dutro, 1975), although some writers place the uppermost part of the Alaska Bench in the Atokan (Willis, 1959; Maughan and Roberts, 1967; Mallory, 1972; Maughan, 1975). The Alaska Bench is equivalent to the upper part of the Namurian Series and possibly extends upwards into the base of the Westphalian Series (Sando, Gordon, and Dutro, 1975).

The Alaska Bench Limestone was deposited in a shallow-water marine environment, resulting in a series of microfacies, described by Gilmour (1969). These microfacies represent rocks deposited in supratidal-intertidal, marginal subtidal marine, and normal subtidal marine environments. These microfacies were deposited as cyclic units and record the transgressive-regressive movements of the strandline of the Morrowan sea across central Montana. The strandline was oriented east-west, the sea floor gently sloping to the south. Maughan (1975) stated that an increase of terrigenous sediments in western Montana and the Dakotas suggests land areas west and east of central Montana. Because of the fine size of the material, he believed that the sedimentary source areas probably were moderately distant or low lying or both. Gilmour (1969) believed that these interbedded terrigenous sediments in central Montana were deposited in both marine and nonmarine environments concurrently with the various carbonate microfacies.

DEVILS POCKET FORMATION

Gardner (1959, p. 347–348) proposed the name Devils Pocket Formation for 43 m of cherty dolomite, limestone, red sandstone and shale, and chert breccia that overlies the Alaska Bench Limestone in Road Canyon, southeastern Big Snowy Mountains (fig. 1). According to Maughan and Roberts (1967, p. B15) similar rocks in the equivalent position extend throughout eastern and southern Montana and have been included previously within the Minnelusa Formation. Dolomite and sandstone of the Devils Pocket Formation intertongue with and grade into sandstone of the Quadrant Formation toward western Montana (Maughan, 1975, p. 286). The Devils Pocket overlaps older strata toward the south.

Thickness of the Devils Pocket Formation varies considerably because of pre-Middle Jurassic erosion, which removed much of the formation. A thickness of 67 m in southwestern Big Snowy Mountains is the most complete exposure of the formation. Gardner (1959, p. 348) listed thicknesses of 5.5, 11.5, and 21.5 m in central Montana. The formation thins eastward toward the Williston basin (Willis, 1959, fig. 12).

The Devils Pocket Formation consists of sandstone and siliceous dolomite and some dolomitic limestone and siltstone (Gardner, 1959, p. 338, 342–343). The formation changes from predominantly dolomite in the lower part to sandstone in the upper part. Mundt (1956a, p. 1931) presented evidence for an unconformity between the Devils Pocket and the underlying Alaska Bench Limestone. Maughan and Roberts (1967) and Mallory (1972) also supported this idea. Alternatively, Gardner (1959, p. 335) stated that the Devils Pocket rests conformably on the Alaska Bench Limetone and that lithologies are transitional from one formation to the other. In central Montana, the Devils Pocket Formation is overlain unconformably by Jurassic rocks of the Ellis Group. In eastern Montana, southern Montana, and western Montana, the carbonate rocks of the Devils Pocket grade upward into the sandstone or quartzite of the Quadrant Formation. Maughan and Roberts (1967, p. B16) reported an increase in sand westward until the Devils Pocket Formation cannot be separated from the overlying Quadrant Formation. They also stated that the dolomitic middle member of the Minnelusa Formation in eastern Montana grades laterally into the Quadrant of central Montana and the Tensleep of Wyoming and should be included in the Devils Pocket Formation.

Few fossils have been reported from the Devils Pocket Formation. Henbest (1954, p. 50, 51) and Easton (1962, p. 16–17) reported fusulinids from clasts and matrix of the breccia near the top of the formation. Included in their faunal lists are: *Climacammina* sp., *Endothyra* sp., *Bradyina* sp., *Tetrataxis* sp., *Millerella* sp., *Pseudostaffella* sp., *Profusulinella* sp., *Cornuspira* sp., *Spiroplectammina* sp., *Climacammina magna*? Roth and Skinner, 1930, *Derbyia* sp., and *Straparollus* (Euomphalus) sp. Both authors believed that the formation is Atokan because of the presence of *Profusulinella* sp. George Verville (oral commun. reported by Maughan, 1975) assigned a late Atokan age to the Devils Pocket Formation on the basis of fusulinid studies.

Dolomite and sandstone of the Devils Pocket Formation are believed to have been deposited in ". . . a marine environment of above normal salinity" (Maughan, 1975, p. 288). This assumption is based on the type of fossils described and a belief that the dolomite is either primary or penecontemporaneous. Easton (1962, p. 26) believed that the formation "represents warm, emergent conditions, which favored the deposition of interbedded red clastics and sandy calcareous deposits which altered penecontemporaneously or later to cherty dolomite." Some of the dolomite beds in the Devils Pocket Formation are very similar to those in the underlying Alaska Bench Formation described by Gilmour (1967) as dolomitized normal marine beds. Fossils listed in the Devils Pocket Formation indicate normal marine conditions (fusulinids and brachiopods). Present available evidence suggests that the Devils Pocket Formation was deposited in a marine environment of normal salinity.

A rising land area to the west or northwest is suggested by the great thickness of sandstone in western Montana, the rapid eastward thinning, and the overlap of sandstone over carbonate rocks eastward (Maughan, 1975, p. 287). Maughan believed that Ordovician sandstone was eroded and served as the principal source for the Pennsylvanian sand. He also suggested that the mudstone in the eastern part of the area was derived from land areas to the east or southeast.

QUADRANT FORMATION

The Quadrant Formation as used in this report refers to the quartzite or sandstone sequence that overlies the Devils Pocket Formation in central Montana and the Amsden Formation in southwestern Montana. The Quadrant grades eastward into dolomite and sandy dolomite that are included in the middle member of the Minnelusa Formation in eastern Montana.

The Quadrant ranges in thickness from zero in central Montana to more than 80 m at the northwest corner of Yellowstone Park (Williams, 1962). South and west it thickens markedly to more than 800 m near the Idaho-Montana State line (Sloss and Moritz, 1951, p. 2163). Sandstone and dolomitic sandstone referred to as Quadrant or Tensleep extends from the Williston basin southwestward to the Idaho-Mantana border and southward to the Wyoming-Montana border. Scott (1935) described the Quadrant "quartzite" as well-bedded, white to pink, fine- to medium-grained quartzite, containing thin beds of siliceous limestone. Gardner and others (1945) described the Quadrant as light-gray quartzitic sandstone with pink or yellowish-brown tints, composed of fine- to medium-grained, angular to subangular quartz grains. In southwestern Montana, the Quadrant is composed of sandstone and quartzite; sandy dolomite beds are found in the lower part of the formation. The percentage of dolomite increases towards central Montana, and dolomite is the dominant lithology in eastern Montana.

The contact between the Quadrant Formation and the underlying Devils Pocket Formation is gradational; the boundary is generally arbitrarily placed between the dominantly carbonate sequence and the overlying dominantly quartzite or sandstone sequence (Maughan and Roberts, 1967, p. B16). Near the Idaho-Montana border, Sloss and Moritz (1951) placed the base of the Quadrant at the base of the lowest massive sandstone bed, leaving several thin sandstone beds in the Amsden and some dolomite beds in the basal Quadrant.

The top of the Quadrant is marked by a major unconformity throughout most of Montana. Rocks overlying the unconformity range in age from Early Permian to Middle Jurassic and, locally, Cretaceous (Mallory, 1972, p. 122).

The age of the Quadrant Formation is Desmoinesian, on the basis of fusulinids (Thompson and Scott, 1941; Henbest, 1954, 1956). Henbest (1954, p. 52) listed the following fossils for the Tensleep (Quadrant Formation of this paper) in Montana: *Endothyra* sp., *Wedekindellina euthy-*

septa (Herbert), *Fusulina tregoensis* (?) Roth and Skinner, *Bradyina* sp., and *Fusulina* sp. Henbest believed that this fauna represents the lower half or two-thirds of the Desmoinesian.

Scott (1935) and Williams (1962) supported a marine origin for the Quadrant Formation in southern Montana and Wyoming. Mallory (1972, p. 122) and Maughan (1975, p. 288, 289) stated that during deposition of the Quadrant, the entire area was inundated by a shallow sea in which the circulation of marine water was restricted. The uplift to the west was the source of clastic materials in western Montana; an eastward source continued to supply fine clastic sediments to the east and southeast. Thin beds of chert and dolomite in the Quadrant contain the fusulinids described by Henbest (1954). As fusulinids are known only from normal marine environments, the conclusion is that at least some of the Quadrant Formation was deposited under such conditions.

PETROLEUM AND NATURAL GAS

MISSISSIPPIAN STRATA

Mississippian strata produce oil and gas from approximately 1,100 wells in 37 fields in Montana. Of these fields, 26 produce only oil, 2 produce only gas, and 9 produce both oil and gas (Montana Board of Oil and Gas Conservation, 1976).

Production figures for the 21 largest of these fields indicate average production depths of 1925 m, cumulative production through 1976 of 252,793,000 bbls of oil, and reserves of 58,457,000 bbls of oil. Total 1976 oil production from these wells was

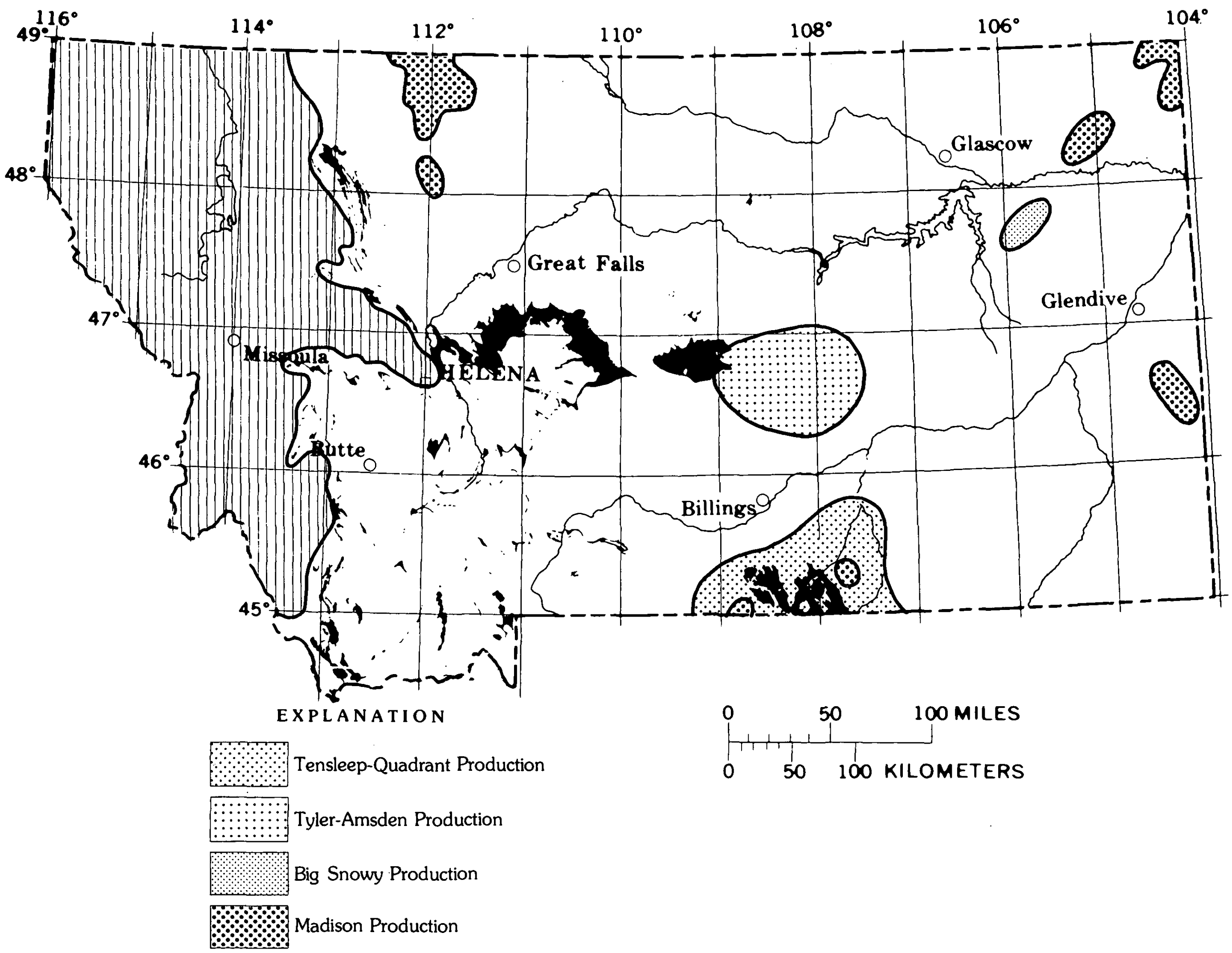


FIGURE 10.—Carboniferous oil and gas fields in Montana.

6,697,853 bbls. Total 1976 gas production from Mississippian wells was 524,282 MCF.

Of these 21 largest fields, 18 produce from Madison Group strata, primarily the upper Mission Canyon and Charles Formations, and 3 produce from Big Snowy Group rocks (Kibbey and Heath Formations).

Madison fields occur in front of the overthrust belt in northwestern Montana, along the northwestern margin of the Williston basin, on the Cedar Creek anticline on the southwestern margin of the Williston basin, and on the flanks of the Bighorn and Powder River (fig. 10). Big Snowy Group fields are centered on the Big Snowy trough, along Tertiary structures east of the Big Snowy Mountains.

PENNSYLVANIAN STRATA

Thirty-four fields produced oil from 305 wells in Pennsylvanian strata during 1976 (Montana Board of Oil and Gas Conservation, 1976). No fields produce only gas, but three fields produce both oil and gas (Elk Basin, Keg Coulee, and Sumatra). Approximate cumulative production from Pennsylvanian fields in Montana is 127,984,000 bbls. Known oil reserves in the 26 largest fields are 33,346,000 bbls. Sixty-two percent of the Pennsylvanian fields produce from the Tyler Formation. Tyler oil ranges from 28 to 34 gravity °API, having a mean value of 32 °API. The Amsden Formation accounts for 22 percent of the producing fields; the oil ranges from 19 to 30 gravity °API. Sixteen percent of the Pennsylvanian fields produce from the Tensleep Formation (Quadrant Formation); the oil ranges from 27 to 37 gravity °API.

Nearly all fields producing from the Tyler and Amsden Formations are in central Montana (fig. 10). Fields producing from the Tensleep (Quadrant) are in south-central Montana. Largest Pennsylvanian fields on the basis of past production are Elk Basin (Carbon County), Sumatra (Rosebud County), and Stensvad (Musselshell County). In 1976, Sumatra field produced 2,019,813 bbls of oil and 160,915 MCF of gas. Jim Coulee (Musselshell County) produced 489,808 bbls of oil, and Elk Basin produced 482,390 bbls of oil and 369,660 MCF of gas. In 1976, two new fields were found in the Tyler Formation and two extensions were made in Tyler producing fields.

The dark organic shale and limestone of the Heath Formation (Mississippian) and the black limestone of the Bear Gulch Limestone tongue (Mississippian?) are believed to be the source of oil for the Tyler Formation (Varland, 1956; Willis, 1969). Principal production is from sandstone in the Stonehouse Canyon Member of the Tyler. Traps are found in the crests or axes of domes or anticlines and on the flanks of anticlines (Jensen and Carlson, 1972).

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Utah

By JOHN E. WELSH *and* HAROLD J. BISSELL

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Utah

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—UTAH

By JOHN E. WELSH[1] and HAROLD J. BISSELL[2]

ABSTRACT

A late Kinderhookian to Osagean Redwall carbonate bank prograded northwestward from southeastern Utah over a starved phosphatic basin and formed the clinoform Monte Cristo Limestone in southwestern Utah and the Deseret Limestone in central Utah. The interior of the Redwall carbonate bank was extensively dolomitized in southeastern Utah; the same lithofacies in northern Utah is Brazer Dolomite. Later, in late Meramecian and Chesterian time, the Great Blue carbonate bank prograded westward nearly to the present Nevada border and covered Antler flysch deposits. Clastic materials in the Mississippian were derived primarily from two sources. Erosion of the Antler Highlands in central Nevada provided fine clay and silt to the Chainman Formation and fine quartz sand and coarse chert pebbles to the Diamond Peak Formation. Erosion on the craton northeast of Utah provided clay, silt, and sand which were transported westward down the Doughnut trough and then deposited as prograding prodeltaic and deltaic deposits of the Deseret Limestone, Humbug Sandstone, and Manning Canyon Formation.

The Pennsylvanian Oquirrh basin in north-central Utah and the Paradox basin in southeastern Utah were elongated downwarps that received feldspathic sands beginning in the Morrowan and received an increasing volume of clastic material in late Des Moinesian through Virgilian time from the Uncompahgre uplift. Evaporites were deposited during a short period of the early Des Moinesian (Cherokee) time when the southeast inlet to the Paradox basin was barred by algal stratigraphic reefs. The Antler Highlands in central Nevada contributed fine clastic chert and quartz to the Hogan Member of the Ely Limestone on the Ely shelf in western Utah during the Des Moinesian. The Antler Highlands never were a source for the feldspathic sandstones of the Oquirrh Formation. Pre-Wolfcampian erosion stripped all the Pennsylvanian rocks off the Emery high in central Utah and beveled the Pennsylvanian rocks across the entire southwest quarter of the State.

Late Mesozoic (Sevier) structures are strongly influenced by the Carboniferous stratigraphy and paleogeography. The Leamington Canyon tear fault and the Charleston-Nebo allochthon are spatially controlled by the original northwestern edge of the Redwall carbonate bank and the south side of the Doughnut trough. The Chainman decollement of northeastern Nevada and northwestern Utah has caused the overlying Pennsylvanian, Permian, and Triassic sequence to shear into multiple nappes. Now these nappes, consisting of distinctly different parts of the upper Paleozoic and Triassic sequence, rest structurally upon plastically deformed shale of the Chainman Formation or directly upon the footwall of the Devonian carbonate rocks along the flanks of the Mesozoic anticlinoria and gneiss domes. Lower Mississippian limestone, particularly the Joana Limestone, was widely boudinaged by the Chainman decollement.

The Carboniferous rocks of Utah do not contain any economic coal deposits; however, they have yielded more than 361 million barrels of oil, mostly from the Paradox basin. Carboniferous limestones contain ores of copper, lead, silver, gold, zinc, and arsenic. Potash, clay, and limestone are produced from the Carboniferous deposits.

INTRODUCTION

Carboniferous rocks crop out in the fault-block mountains of western Utah, in the asymmetrical overthrust anticlines of the Wasatch Hinge Line, in the deep canyons of the Canyon Lands (fig. 1, locs. 42, 45), in the Paradox salt anticlines (fig. 1, loc. 43), and along the flanks of the Uinta Mountains (fig. 1, loc. 29). The Carboniferous sequence has been densely drilled for oil and gas only in the Paradox basin in southeastern Utah. Widely spaced drilling in the High Plateaus and Uinta Basin has provided useful Carboniferous stratigraphic control for this compilation. A few holes have penetrated the sequence in the Basin and Range province, but samples and logs are incomplete. Excellently exposed sequences of the Carboniferous are found in the tilted ranges in western Utah.

In northeastern Utah, the Lower Mississippian limestone sections crop out either as hogbacks or as near-vertical canyon walls. In southeastern Utah, where no Mississippian outcrops exist, the Pennsylvanian cyclical limestones form ledge and slope topography at Cataract Canyon (fig. 1, loc. 42) of the Colorado River and at the Goosenecks (fig. 1, loc. 45) of the San Juan River. Tilted fault blocks of Pennsylvanian limestone, anhydrite, and black shale form linear ridges in the Paradox salt anticli-

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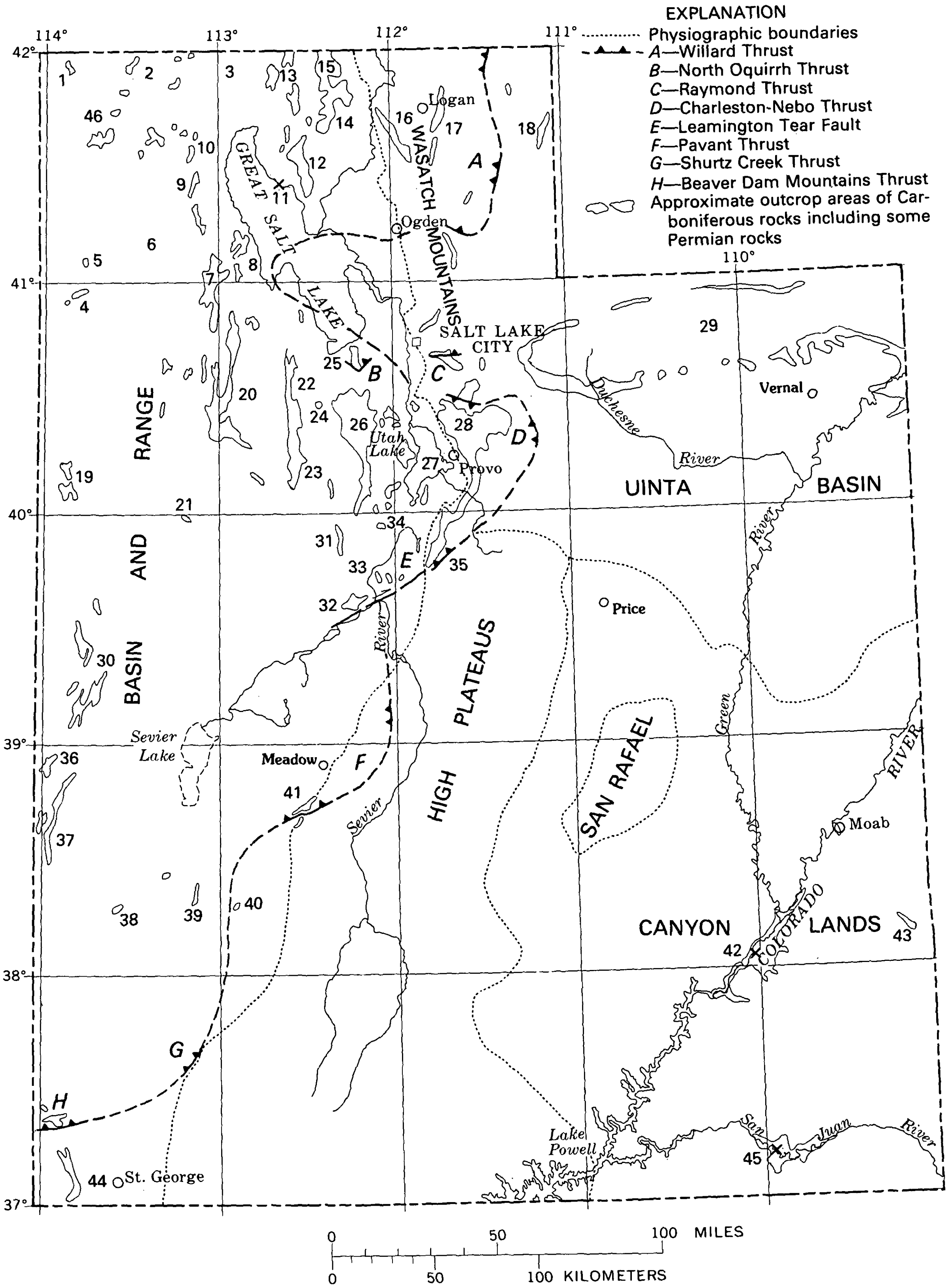

Figure 1.—Locations of outcrops of Carboniferous rocks in Utah. Outcrops west of the thrust belt are allochthonous. Numbers refer to locality list. (See locality list on facing page.) Uppercase letters refer to thrust and tear faults recognized along eastern boundary of the Sevier Orogenic Belt. Physiographic boundaries are indicated by a dotted line.

nal valleys near Moab and Lisbon (fig. 1, loc. 43). Salt solution in these anticlines has caused collapse structures to form.

In the Basin and Range province, Upper Mississippian rocks generally crop out as long strike valleys. Where resistant limestones or sandstones are present in the Mississippian sequences, these strike valleys may have a series of parallel minor hogbacks. Cuestas are less common than hogbacks because the dip is generally greater than 10°. Pennsylvanian cyclical limestones and sandstones in the same province give rise to steplike topography which usually extends to the crest of the ranges. Most of the magnificent skyline of Mt. Timpanogos (fig. 1, loc. 28) and Mt. Nebo (fig. 1, loc. 35) in the southern Wasatch Mountains is formed by the Pennsylvanian Oquirrh Formation.

A generalized biostratigraphic zonation of Carboniferous deposits in Utah is shown on figure 2; figure 3 is a lithostratigraphic correlation chart of the Utah Carboniferous. Displacement of allochthonous sequences is recognized in the belt of overthrusting along the Wasatch Hinge Line from southwesternmost Utah to near Logan and in areas of denudation adjacent to the Raft River gneiss dome in northwestern Utah and the Snake Range-Deep Creek gneiss dome in eastern Nevada and western Utah.

The stratigraphic data from the Uinta and Wasatch Mountains and northern Utah were compiled by Bissell; the remaining surface and subsurface information was compiled by Welsh. This paper includes a generalization of much unpublished stratigraphic information originated by Welsh. The writers have attempted to simplify the stratigraphic terminology and to relate it to specific lithostratigraphic facies.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Utah Geological and Mineral Survey.

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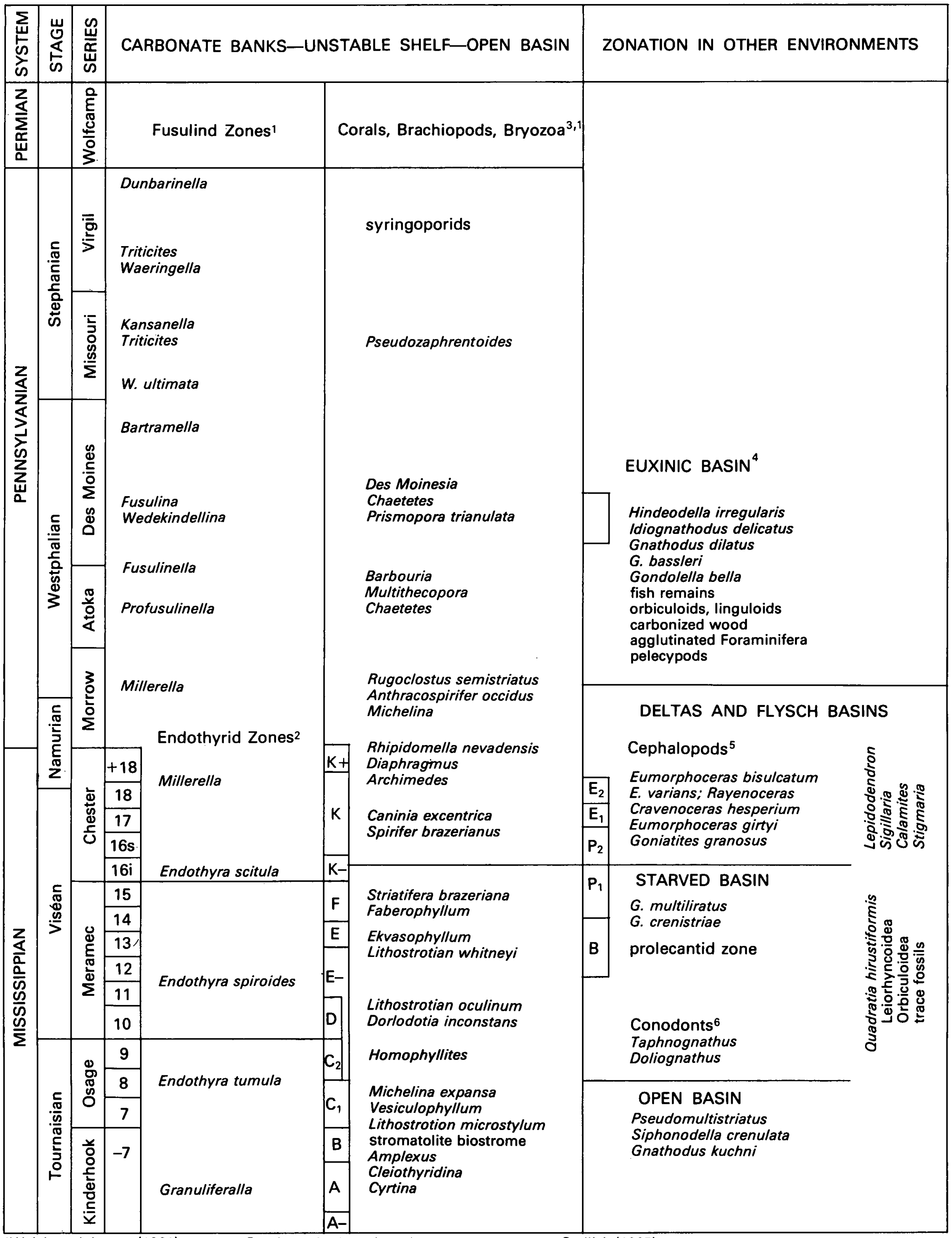
The writers acknowledge the exchange of information with C. A. Sandberg, W. J. Sando, and R. Gutschick concerning the starved-basin concept and paleontological zonation. The writers are, however, wholly responsible for the data presented. Graduate students of many universities deserve credit for undertaking mapping and stratigraphic projects in remote areas which have complex geology.

Welsh is indebted to A. Shaw for a rigorous introduction to paleontology, to R. Spivey for an introduction to fusulinids, and to A. James and W. L. Stokes for an appreciation of the value of stratigraphy in solving geologic problems. Bissell has benefited greatly from source data collected by some of his graduate students and also is grateful to various

FIGURE 1.—Continued

Locality list

1. Goose Creek Mountains
2. Raft River Range
3. Curlew Valley
4. Silver Island Range
5. Crater Island
6. Newfoundland Mountains
7. Grassy Mountains
8. Lakeside Mountains
9. Terrace Mountains
10. Hogup Mountains
11. Rozel Point
12. Promontory Mountains
13. Hansell Mountains
14. Blue Hill Mountains
15. West Mountains
16. Wellsville Mountains
17. Bear River Range
18. Crawford Mountains
19. Gold Hill
20. Cedar Mountains
21. Dugway Range
22. Stansbury Mountains
23. Onaqui Mountains
24. South Mountain
25. North Oquirrh Mountains
26. South Oquirrh Mountains
27. Lake Mountains
28. Timpanogos Mountain
29. Uinta Mountains
30. Confusion Range
31. West Tintic Mountains
32. Gilson Mountains
33. East Tintic Mountains
34. Tintic Mountains
35. Mt. Nebo
36. Burbank Hills
37. Needles Range
38. Southern Wah Wah Mountains
39. Star Range
40. Bradshaw Mountain
41. Pavant Range
42. Cataract Canyon
43. Lisbon Valley
44. Beaver Dam Mountains
45. Goosenecks of San Juan River Canyon
46. Grouse Creek Mountains



[1]Welsh and James (1961)
[2]B. L. Mamet and Betty Skipp, *in* Sando and others (1969)
[3]Sando and others (1969); J. T. Dutro, Jr., *in* Tooker and Roberts (1970)
[4]Stone (1968)
[5]Sadlick (1965)
[6]Sandberg and Gutschick (1977)

FIGURE 2.—Generalized biostratigraphic zonation of Carboniferous deposits in Utah.

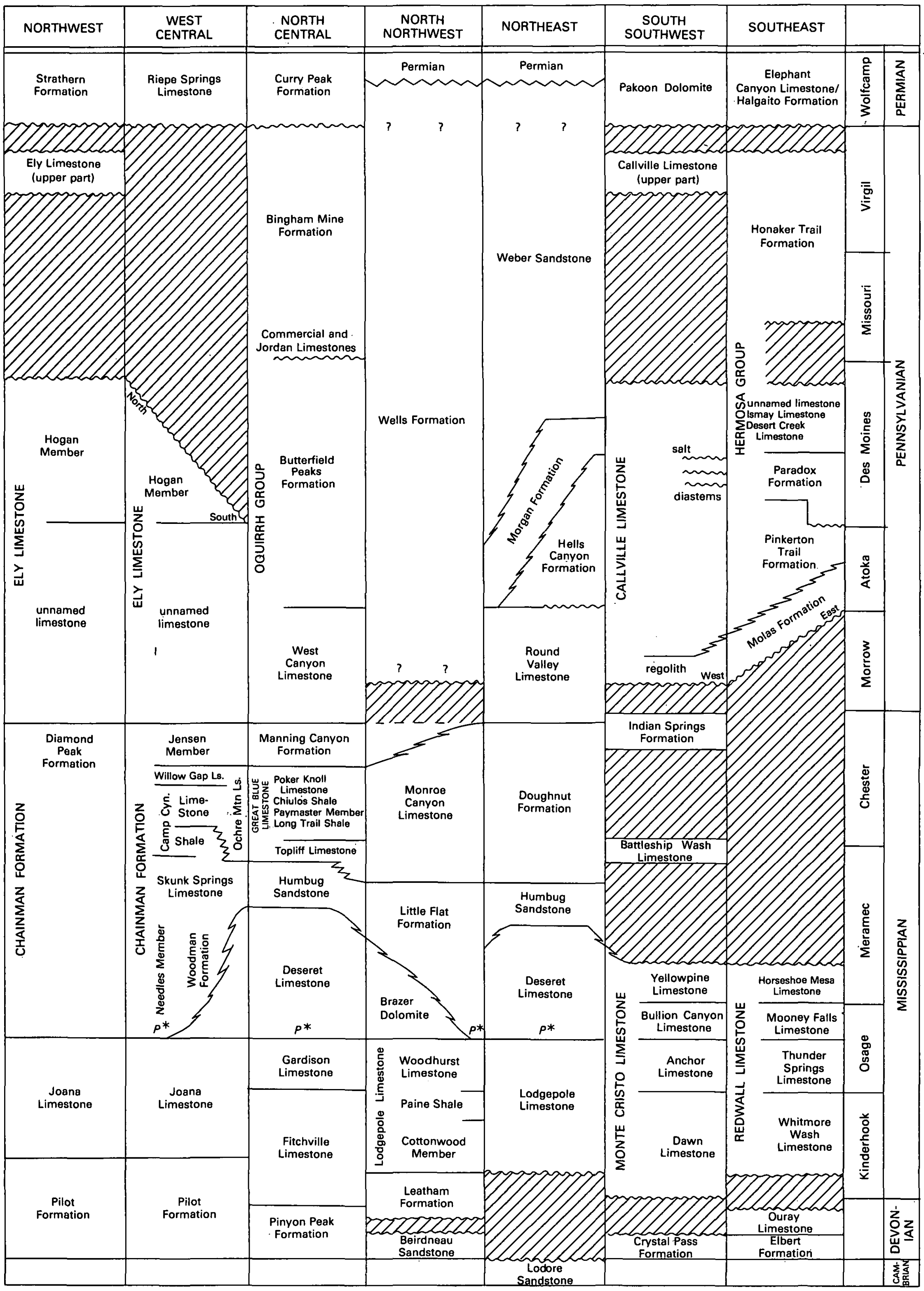


FIGURE 3.—Lithostratigraphic correlation of Carboniferous deposits in Utah. The presence of phosphatic shale is indicated by p*.

oil companies for releasing some data accrued while he was in their employ.

The writers appreciate the cooperation of the Utah Geological and Mineral Survey personnel under the directorship of Don McMillan. Brent Jones supervised the drafting of the illustrations. Martha Smith has immeasurably improved the manuscript by helpful suggestions and editing. Carlton Stowe provided the oil-production data.

HISTORY OF STRATIGRAPHIC NOMENCLATURE

HISTORICAL BACKGROUND

More than a century ago, Clarence King (1876) named the Weber Quartzite in Weber Canyon east of Ogden. Blackwelder, in 1910, separated the red sandstones and shales at the base of the quartzite cliffs as the Pennsylvanian Morgan Formation. Shortly thereafter, in 1912, Richards and Mansfield gave the name Wells Formation to rocks below the Phosphoria Formation in Wells Canyon, Bannock County, Idaho. The Wells Formation is in the same stratigraphic position as the Weber Formation, but their correlation is not well documented, even today.

Discovery of replacement ore bodies of precious and base metals in Carboniferous rocks at Park City, Bingham, Tintic, Ophir, Gold Hill, and other, lesser districts resulted in generalized stratigraphic studies and a proliferation of formation names. Formations were poorly defined because of structural and alteration complications, and names were extended to other areas on the basis of paleontology rather than lithology. This practice has resulted in an imprecise use of lithostratigraphic terms. The nomenclature of the Mississippian sequences has been sufficiently revised, as discussed below, that these sequences represent coherent lithofacies. The Pennsylvanian sequences are still in a state of confusion, particularly in the Oquirrh basin where 9,000 m of Pennsylvanian and Permian section are referred to as the "Oquirrh Formation." Unfortunately, paleontologic extension of formation names continues.

MISSISSIPPIAN NOMENCLATURE

The Mississippian sequences were lithologically divided by Gilluly (1932) in the Oquirrh Mountains (fig. 1, loc. 26) and by Nolan (1935) in the Gold Hill district (fig. 1, loc. 19). The Oquirrh Mountain terminology of the Deseret Limestone, Humbug Sandstone, Great Blue Limestone, and Manning Canyon Formation has been extended as the accepted terminology of the allochthonous sequences west of the Wasatch Hinge Line. The Deseret Limestone and Humbug Sandstone have also been accepted to the east for the autochthonous sequences. The Lower Mississippian sequences, formerly called "Madison," are now called the Gardison and Fitchville Limestones because of correlation with those limestones in the Tintic district (Morris and Lovering, 1961). Equivalent rocks in northern Utah are now called the Lodgepole (Holland, 1952). The Chesterian Manning Canyon Formation of the Oquirrh Mountains has been found in all the allochthonous sections of the proto-Oquirrh basin, whereas the thinner Doughnut Formation includes the late Meramecian to Chesterian rocks on the shelf to the east in the Wasatch and Uinta Mountains. Nolan's Upper Mississippian units of the Woodman Formation and Ochre Mountain Limestone at Gold Hill were extended by Staatz (1972) into the Dugway Range (fig. 1, loc. 21), but these units need to be further defined by larger scale mapping at Gold Hill before they can be correlated regionally.

Williams (1948) and his students measured reconnaissance sections in northern Utah, but the descriptions of the rocks were generalized. Parks (1951) first described the coral zones of the Upper Mississippian section in the Wellsville Mountains (fig. 1, loc. 16). Sando and his colleagues (1959 and 1976) have further revised the lithostratigraphy and biostratigraphy in northern Utah by restricting the name Brazer Dolomite to the Crawford Mountains (fig. 1, loc. 18) and by using the name Little Flat Formation for basinal siltstones equivalent to the Deseret Limestone. Limited stratigraphic studies of the Mississippian in northern Utah were done in conjunction with the U.S. Geological Survey's phosphate program in the late 1940's and early 1950's.

Oil companies initiated regional stratigraphic studies in the early 1950's, but very little critical work was accomplished until oil was discovered in the mid-1950's in the Paradox basin, Utah, and at Eagle Springs, Nevada. These discoveries spurred investigations throughout both States. The U.S. Geological Survey geologists restudied the stratigraphy of the Tintic district in the 1950's, then later initiated an investigation of the northern Oquirrh Mountains (fig. 1, loc. 25) following the completion of the stratigraphic field work in the southern Oquirrhs by Bissell (1959) and in the overall Oquirrh Mountains by Welsh and James (1961). Sadlick, in graduate studies at the University of

Utah (1956, 1957, 1965), contributed substantially to the biostratigraphy in northeastern and western Utah and was the first to describe the flysch facies of the Upper Mississippian sections in western Utah. Sadlick and Mackenzie Gordon, Jr. (oral commun., 1960), did most of the goniatite zonation of the Upper Mississippian sequences. By the late 1950's a geologic mapping program, supervised by Lehi Hintze and Lee Stokes, was initiated and was funded by the Utah State Land Board. Graduate students at Brigham Young University and the University of Utah received partial field expenses for much of the original geologic mapping in remote areas of the State. These mapping and stratigraphic theses resulted in new regional information on the Carboniferous rocks and made possible the compilation of the 1:250,000-scale "Geologic Map of Utah" (Hintze and others, 1962–1964).

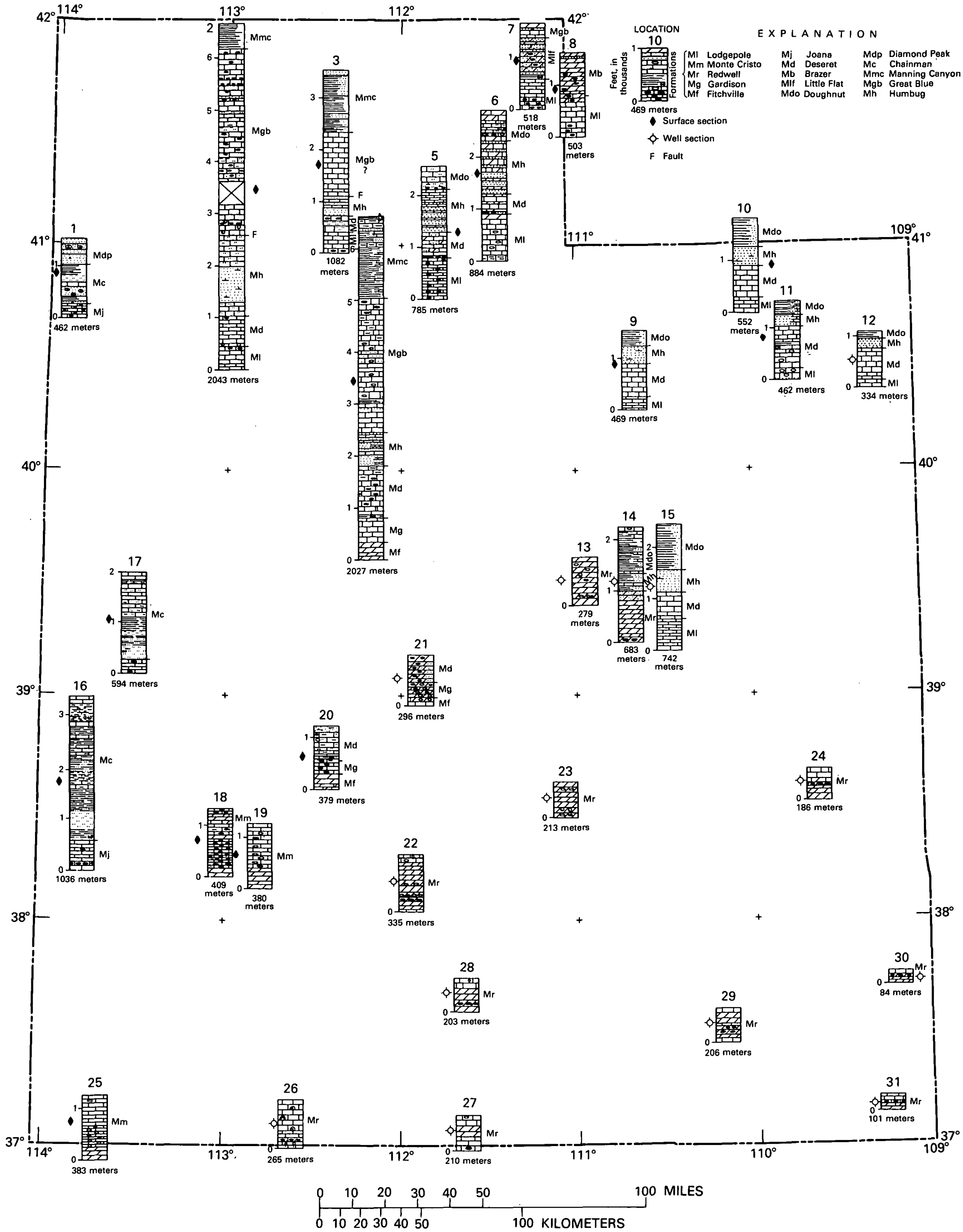
Hose and Repenning (1959) and Langenheim (1963) formally extended the Joana Limestone, Chainman Formation, and Monte Cristo Limestone terminology into western Utah, where oil company geologists had been using the terminology informally since the early 1950's. Parker and Roberts (1966) formally used McKee and others' (1969) members of the Redwall Limestone to designate units in the subsurface of southeastern Utah. The recognition that the Redwall Limestone was the interior carbonate bank deposit and that the Monte Cristo Limestone was the clinoform slope deposit was made in the early 1960's by oil company geologists in southern Nevada. Rose (1976b) found this relationship to extend along the Wasatch Hinge Line across Utah into Idaho. The basinal siltstone facies of the Deseret in the Pavant Range (fig. 1, loc. 41), now known to be equivalent to the Little Flat Formation, was first recognized by Welsh (1972). Several recent oil tests (fig. 4, secs. 13, 21) between Meadow in the Pavant Range and Hiawatha southwest of Price (fig. 1) have provided the control necessary for defining the initial northeast trend of the Osagean Redwall carbonate bank (fig. 5).

Gutschick (1976) and Sandberg and Gutschick (1977) interpreted conodont zones in the phosphatic shales of the lower part of the Deseret Limestone as having been deposited in a starved basin of the early Osagean. The recognition of this starved basin (fig. 5) helped explain the northwest progradation of the Osagean and lower Meramecian limestones. Dolomitization of the interior carbonate bank (fig. 5) was recognized by Sando and others (1959) as the Brazer Dolomite. This same dolomitization had been recognized in outcrops earlier by oil geologists at Frenchman Mountain near Las Vegas. Drilling since the mid-1950's in south-central Utah has shown the universality of dolomitization in the interior Redwall carbonate bank.

Oil geologists in 1954 recognized that the western margin of the Chesterian carbonate bank of the Great Blue and Ochre Mountain Limestones, also called the Great Blue Carbonate Bank (fig. 6), was in the Confusion Range (fig. 1, loc. 30)-Gold Hill (fig. 1, loc. 19) area. This bank margin crops out in the Skunk Springs section (fig. 4, sec. 17) of the Confusion Range. Rose (1976b) first illustrated the regional progradation of this Chesterian carbonate bank almost to the Nevada border. Research on source beds by Sandberg and Gutschick (1977) has renewed interest in Mississippian stratigraphy and we hope that more detailed investigations will result from this economic interest.

PENNSYLVANIAN NOMENCLATURE

The quality of Pennsylvanian stratigraphic data from Utah is directly proportional to the past economic incentive to study the geology of the State. Oil exploration in southeastern Utah, and the resultant discovery of the Greater Aneth field in 1956, stimulated the early synthesis of subsurface data by Wengerd and Strickland (1954), Herman and Sharps (1956), and Herman and Barkell (1957). These investigators assumed lateral facies changes from evaporites to carbonate rocks around the margins of the basin; thus, their correlations cross time–stratigraphic units. In 1958, Welsh showed that in the subsurface of southeastern Utah, an unconformity separates the Des Moinesian and Missourian series and a disconformity separates the lowermost Des Moinesian and Atokan series. Wengerd and Matheny (1958) revised the lithostratigraphy of the Paradox basin and used the top of the Desert Creek Limestone or the equivalent Horn Point Limestone at Honaker Trail (fig. 1, loc. 45) as the top of the Paradox Formation. They included all the superjacent limestone in the Honaker Trail Formation. In 1963, Welsh used fusulinid data to show that the evaporites of the Paradox Formation were equivalent to disconformities on the western margin of the basin and were not equivalent to the fossiliferous marine limestones that overlie and underlie the evaporites. The fact that the contact between the Missourian and Wolfcampian series is unconformable was further documented by fusulinid data. Baars, Parker, and Chronic (1967) reverted


EXPLANATION
LOCATION
10
Feet, in thousands
Formations
469 meters
Ml Lodgepole
Mm Monte Cristo
Mr Redwell
Mg Gardison
Mf Fitchville
Mj Joana
Md Deseret
Mb Brazer
Mlf Little Flat
Mdo Doughnut
Mdp Diamond Peak
Mc Chainman
Mmc Manning Canyon
Mgb Great Blue
Mh Humbug
Surface section
Well section
F Fault
462 meters
2043 meters
1082 meters
2027 meters
785 meters
884 meters
518 meters
503 meters
469 meters
552 meters
462 meters
334 meters
279 meters
683 meters
742 meters
594 meters
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0 10 20 30 40 50 100 KILOMETERS

to the lateral facies concept, ignored time-stratigraphic units, and correlated the evaporites of the Paradox Formation with open-circulation limestones of both Des Moinesian and Missourian ages. Since 1967, many papers have been written on the Paradox basin, but the time-stratigraphic correlation of the facies has not been adequately documented in the literature.

At present, the name Hermosa Group can logically be restricted to the limestone-clastic sequence in the immediate area of the Paradox basin and to the type area in southwestern Colorado. The Pinkerton Trail Formation is the preevaporite carbonate rock of the Atokan and early Des Moinesian; the Paradox Formation is the evaporite sequence; the Desert Creek, Ismay, and unnamed Des Moinesian limestones are the overlying, open-marine formations. The Honaker Trail Formation is restricted to the Missourian and Virgilian limestones. The Callville Limestone as used in this paper is the western platform facies of the Hermosa in south-central Utah. The Wolfcampian Pakoon Dolomite is the western platform facies of the Elephant Canyon Limestone and the red beds of the Halgaito Formation, all of which unconformably overlie Pennsylvanian rocks of either the Callville Limestone or the Hermosa Group.

The Pennsylvanian rocks of the type Oquirrh Formation in the southern Oquirrh Mountains (fig. 1, loc. 26) were not divided by Gilluly (1932). He included approximately 9,000 m of Pennsylvanian and Permian rocks in one map unit and implied complex facies changes within the Stockton and Fairfield quadrangles. Nolan (1930) had earlier applied the Oquirrh name to Pennsylvanian and Permian sequences in the Gold Hill district of western Utah; today the Pennsylvanian part of these sequences is considered to be the Ely Limestone. Nolan's correlation was based upon similar faunal elements, not lithology. Detailed descriptions of the "Oquirrh" sequences at Gold Hill have never been published. Bissell (1959) was the first to publish descriptions of mappable formations in the southern Oquirrh Mountains (fig. 1, loc. 26), and he divided the Pennsylvanian sequence into the Morrowan Hall Canyon Member, the Atokan Meadow Canyon Member, the Des Moinesian Cedar Fort Member, the Missourian Lewiston Peak Member, and the Virgilian Pole Canyon Member. Unquestionably these lithologic units are mappable; however, the youngest rocks in the southern Oquirrh Mountains are Des Moinesian, not Virgilian. Welsh and James (1961) divided the entire Pennsylvanian and Permian sequence of the Oquirrh Mountains into mappable time-strati-

FIGURE 4.—Numbered stratigraphic sections of the Mississippian in Utah. Formational subdivisions and generalized lithologies are shown. Where no section number is supplied, the stratigraphy is a composite of information from more than one section. Stratigraphic sections 13–23 and 25–28 are based on original unpublished data of Welsh; the other sections are modified from published sources as indicated.

1. Silver Island Mountains, T. 1 N., R. 19 W. (modified from Schaeffer, 1960)
2. Lakeside Mountains, T. 6 N., R. 9 W. (modified from Doelling, 1964)
3. Promontory Mountains, T. 7 N., R. 6 E. (modified from Olson, 1960)
4. South Oquirrh Mountains, T. 5 S., R. 4 W. (modified from Gilluly, 1932)
5. Morgan, T. 4 N., R. 3 E. (modified from Nohara, 1966)
6. Causey Dam, T. 7 N., R. 3 E. (modified from Mullens and Izett, 1964)
7. Old Laketown Canyon, T. 13 N., R. 6 E. (modified from Sando and others, 1976)
8. Crawford Mountains, T. 11 N., R. 8 E. (modified from Sando and others, 1976)
9. Duchesne River, sec. 14, T. 1 N., R. 8 W. (modified from Sadlick, 1957)
10. Sols Canyon, sec. 11, T. 2 N., R. 18 E. (modified from Sadlick, 1957)
11. Whiterocks Canyon, T. 2 N., R. 1 E. (modified from Kinney, 1955)
12. Ute Federal, sec. 12, T. 4 S., R. 22 E. (modified from Sadlick, 1957)
13. Hiawatha, sec. 13, T. 15 S., R. 7 E.
14. Miller Creek, sec. 26, T. 15 S., R. 10 E.
15. Mounds, sec. 33, T. 15 S., R. 12 E.
16. Needles Range, T. 25 S., R. 19 W.
17. Skunk Springs, T. 17 S., R. 16 W.
18. Elephant Canyon, T. 28 S., R. 12 W.
19. Bradshaw Mountain, T. 29 S., R. 10 W.
20. Cove Fort, T. 24 S., R. 6 W.
21. Scipio Lake, sec. 14, T. 20 S., R. 2 W.
22. Antimony Canyon, sec. 30, T. 30 S., R. 2 W.
23. South Last Chance, sec. 18, T. 26 S., R. 7 E.
24. Little Valley, sec. 29, T. 26 S., R. 20 E. (modified from Parker and Roberts, 1966)
25. Beaver Dam Mountains, T. 42 S., R. 18 W.
26. Kanab, sec. 2, T. 43 S., R. 8 W.
27. Judd Hollow, sec. 19, T. 43 S., R. 2 E.
28. Upper Valley, sec. 12, T. 36 S., R. 1 E.
29. Moqui, sec. 33, T. 37 S., R. 15 E. (modified from Parker and Roberts, 1966)
30. Coalbed Canyon, sec. 20, T. 35 S., R. 26 E. (modified from Parker and Roberts, 1966)
31. Desert Creek, sec. 2, T. 42 S., R. 23 E. (modified from Parker and Roberts, 1966)

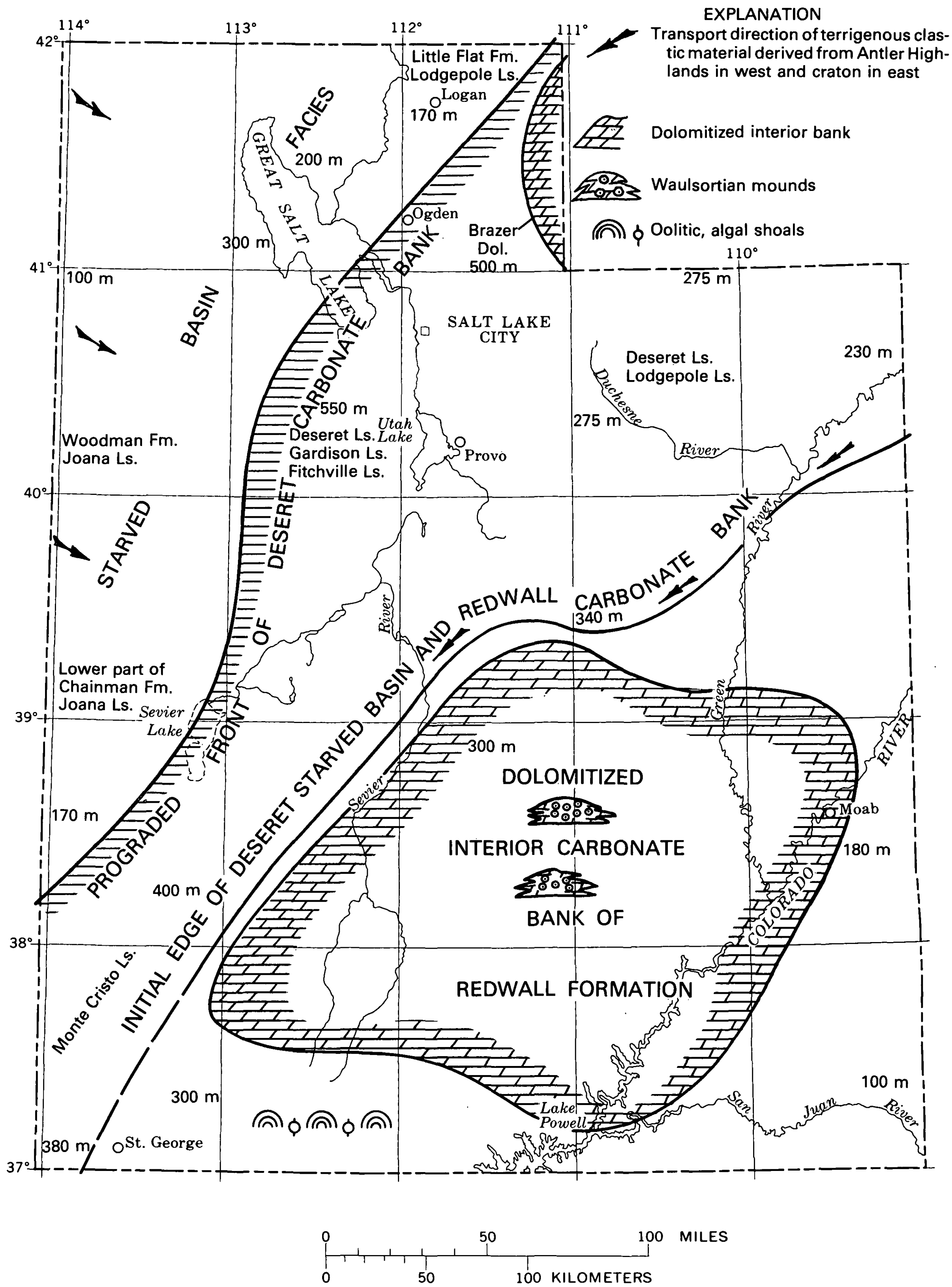


FIGURE 5.—Paleogeographic map of Utah showing approximate present thicknesses in meters of deposits of the late Kinderhookian through Osagean into early Meramecian time.

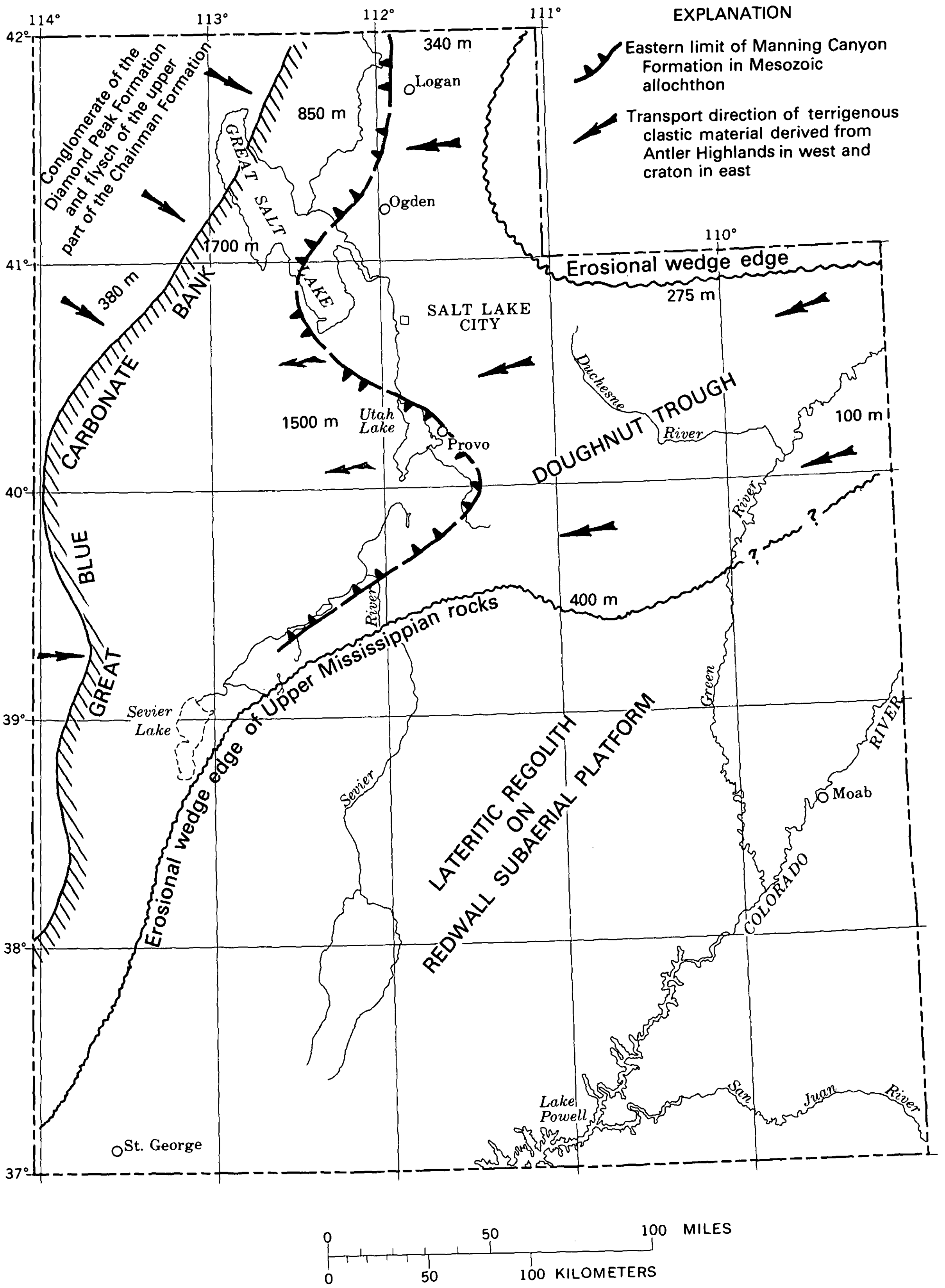


FIGURE 6.—Paleogeographic map of Utah showing approximate present thicknesses in meters of deposits of the late Meramecian to late Chesterian time.

graphic units in conjunction with structural mapping in the Bingham district. Welsh (*in* Welsh and James, 1961) recognized the importance of the South Mountain (fig. 1, loc. 24) section of Des Moinesian through Wolfcampian rocks for understanding the correlation of other structural blocks. He also first recognized the Permian Kirkman Limestone, Diamond Creek Sandstone, and Park City Formation at South Mountain (fig. 1, loc. 24) and in the northern Oquirrh Mountains (fig. 1, loc. 25); these formations completed the sequence. Lateral continuity of Pennsylvanian formations and marker limestone beds throughout the Oquirrh Mountains was demonstrated by Welsh and James (1961). The Oquirrh Group was logically restricted to the Pennsylvanian sequence, and these mappable Permian units in ascending order were established for the first time in the Oquirrh Mountains: Curry Peak Formation, Freeman Mountain or "Clinker" Sandstone, Kirkman Limestone, Diamond Creek Sandstone, and Park City Formation. Tooker and Roberts in 1960 obtained complete access to Welsh's stratigraphic and structural data for the 1961 guidebook (Welsh and James, 1961), and in 1970 published their interpretation of the Oquirrh stratigraphy.

Bissell (1937) and Baker (1947) recognized the usefulness of fusulinids in dividing the Oquirrh Formation in the central Wasatch Mountains (fig. 1, loc. 28). M. L. Thompson and George Verville identified fusulinid collections for Bissell, and Lloyd Henbest identified collections for Baker. Baker's (1972, 1973, 1976) geologic maps of the Charleston-Nebo allochthon (fig. 1, thrust fault D) make it possible to correlate time-stratigraphic units with units in the type section of the Oquirrh Formation in the Oquirrh Mountains. As accurate time-stratigraphic data become available in a few mountains besides the type area, we will be able to synthesize the Oquirrh depositional history.

Chamberlain and Clark (1973) began an environmental interpretation by describing trace fossils in the deeper water environment of the Pennsylvanian and Permian of the Oquirrh basin, but unfortunately they used Bissell's (1959) inaccurate time-rock units of the southern Oquirrh Mountains (fig. 1, loc. 26) for the South Mountain section.

Geologic mapping in isolated mountain ranges by Croft (1956) in the Onaqui Mountains (fig. 1, loc. 23), Rigby (1958) in the Stansbury Mountains (fig. 1, loc. 22), Costain (1960) in the East Tintic Mountains (fig. 1, loc. 33), and Maurer (1970) in the Cedar Mountains (fig. 1, loc. 20) has contributed data about the Oquirrh. Few lithostratigraphic units have been mapped; reliance on time units based on fusulinids has not stimulated structural interpretations or regional lithologic correlations. J. K. Rigby (oral commun., 1977) now suspects that the described onlap of Pennsylvanian Oquirrh Formation on the Mississippian Manning Canyon Formation in the Stansbury Mountains is structural rather than stratigraphic. Many similar structural problems in western Utah await resolution before the Pennsylvanian stratigraphy can be further resolved.

The Callville platform and Ely shelf (figs. 7–9) have been extensively studied by oil company geologists. Hose and Repenning (1959) described in detail the Ely Limestone in the Confusion Range (fig. 1, loc. 30) of western Utah. Bissell (1962) and Brill (1963) presented reconnaissance overviews of the Pennsylvanian in the Cordilleran region. Roberts and others (1965) compiled the Pennsylvanian and Permian data for northwestern Utah, but the critical stratigraphic information was and still is lacking for this region.

Much of the better stratigraphic information gathered by oil company geologists and students is in guidebooks of the Intermountain Association of Geologists, Utah Geological Association, Four Corners Geological Society, and Rocky Mountain Association of Geologists. The Brigham Young University Geology Studies, the Utah Geological Society, and the Utah Geological and Mineral Survey have published many of the graduate theses.

GEOLOGIC SETTING

The lowermost Carboniferous limestones disconformably overlie calcareous siltstones of the Pilot, Pinyon Peak, and Leatham Formations in the miogeosyncline of central and western Utah. The time-stratigraphic boundary between the Famennian and lower Kinderhookian is placed by conodont studies near the top of the Pilot and Leatham Formations and within the Fitchville Limestone that overlies the siltstone of the Devonian Pinyon Peak Formation. In the Colorado Plateau region in southeastern Utah, the Whitmore Wash Member of the Redwall Limestone rests disconformably either upon the siltstone of the middle Famennian Pinyon Peak Formation or upon the Ouray Limestone. In the central Wasatch Mountains (fig. 1, thrust fault C), the Fitchville Limestone rests unconformably upon the Middle Cambrian Maxfield Limestone near the

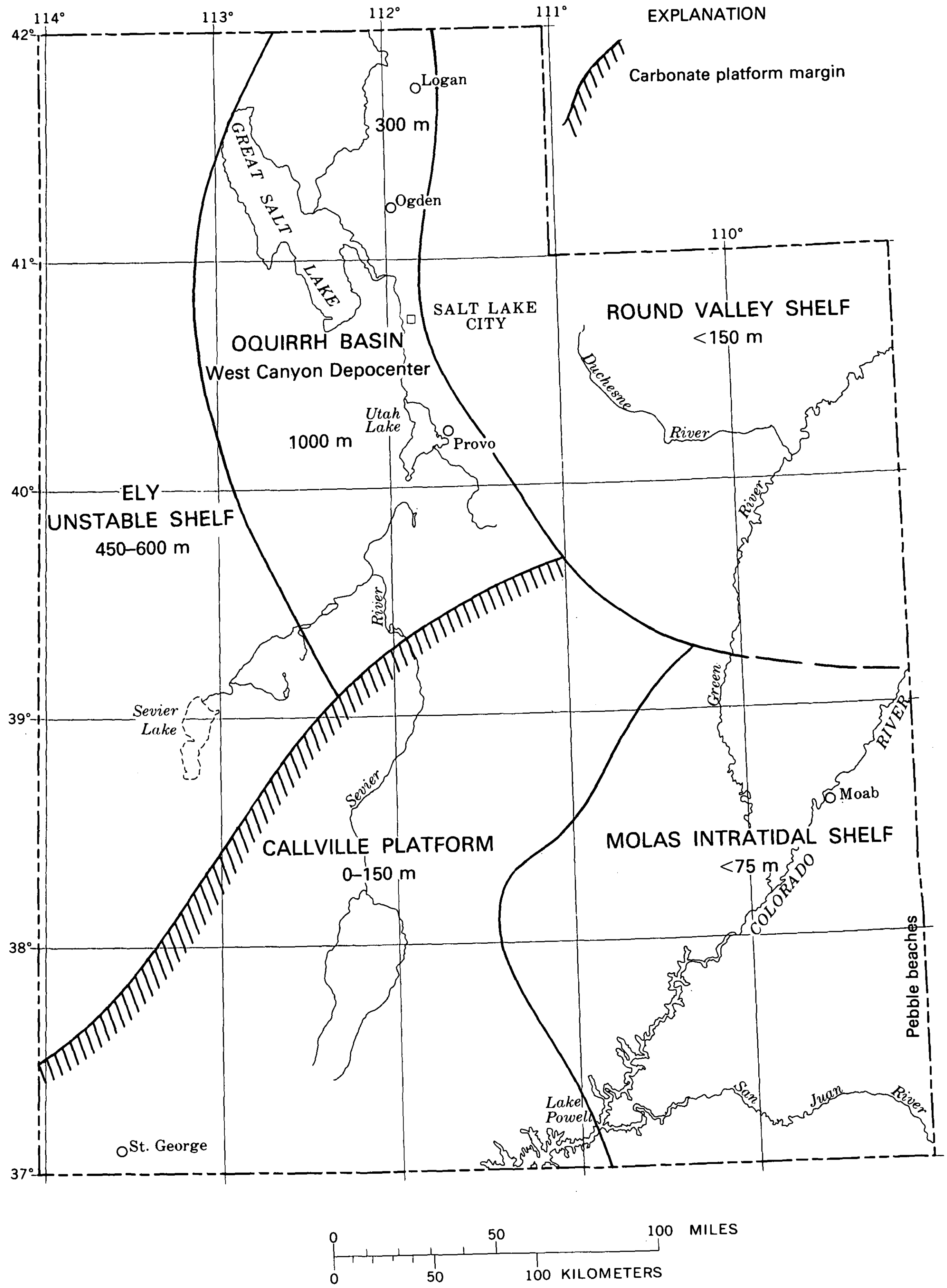


FIGURE 7.—Paleogeographic map of Utah showing approximate present thicknesses in meters of deposits of Morrowan and Atokan time.

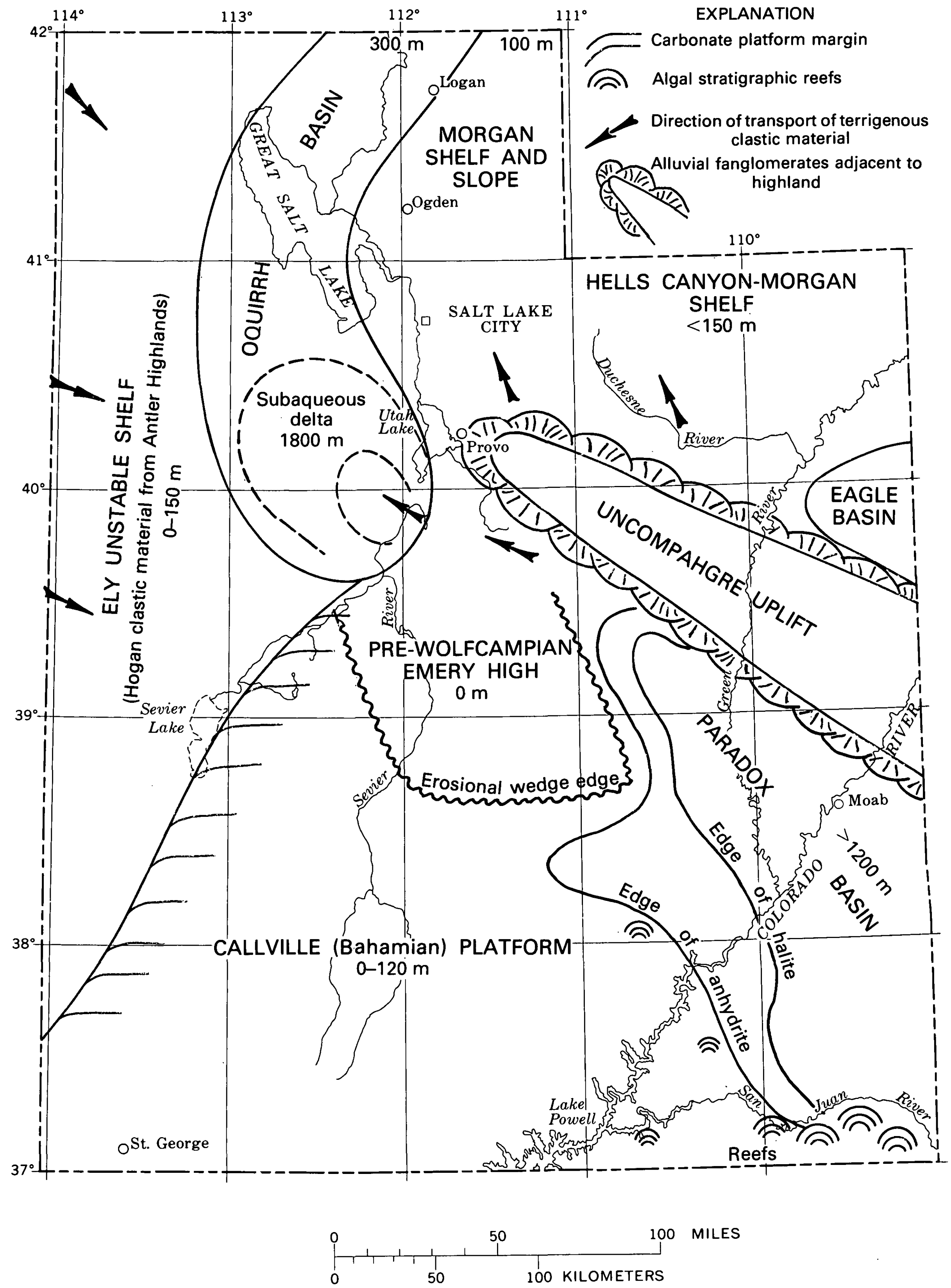


FIGURE 8.—Paleogeographic map of Utah showing approximate present thicknesses in meters of deposits of Des Moinesian time.

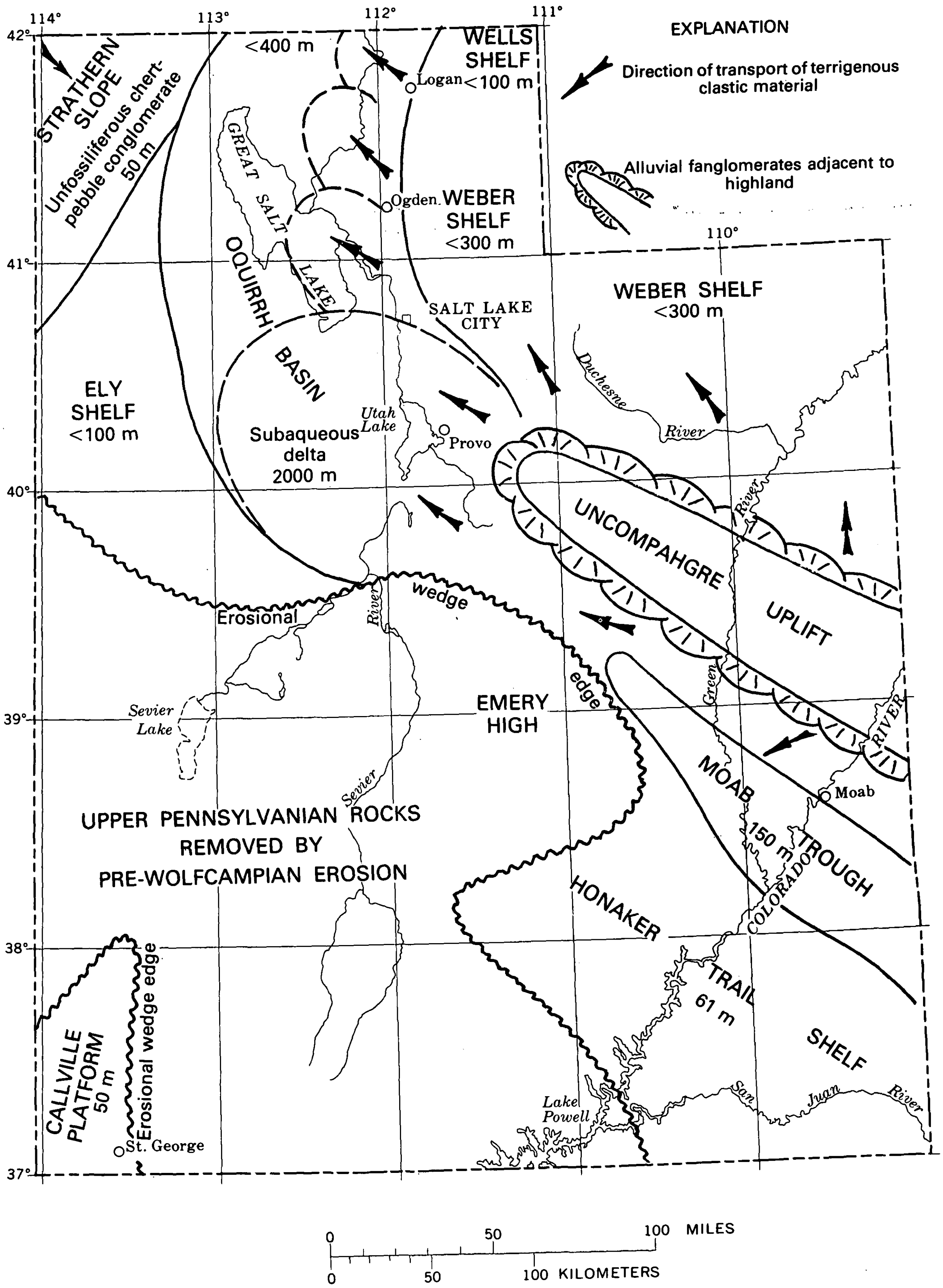


FIGURE 9.—Paleogeographic map of Utah showing approximate present thicknesses in meters of deposits of Missourian and Virgilian time.

site of the Upper Devonian Stansbury uplift; in the Uinta Mountains (fig. 1, loc. 29), the Lodgepole Limestone rests unconformably upon the Cambrian Lodore Sandstone (fig. 3).

Regional depositional breaks within the Carboniferous are (1) a diastem of slow deposition or nondeposition at the top of the Joana, Gardison, Thunder Springs, and Lodgepole Limestones; (2) a lateritic regolith on top of the Horseshoe Mesa Member of the Redwall Limestone in southeastern and southern Utah; (3) an unconformity at the top of the Brazer Dolomite in the Crawford Mountains of northern Utah; (4) a regional disconformity between the Des Moinesian and Missourian series; and (5) an angular unconformity below the Wolfcampian which beveled on a regional scale the Pennsylvanian rocks down to the Mississippian rocks on the Emery high (figs. 8 and 9).

Early Mississippian depositional patterns were affected by the Upper Devonian subaerial highs and restricted basins. This is reflected in the Redwall, Joana, Fitchville, and Lodgepole lithostratigraphic units. Renewed downwarping of local Devonian restricted basins in western and northern Utah produced a regional starved basin in late Osagean and Meramecian time. This downwarping, which began in Late Devonian and continued intermittently through the Chesterian, was east of the central Nevada Antler orogenic belt.

By Late Mississippian, the region of thick carbonate deposition in northwestern Utah of the Deseret and Great Blue Limestones (figs. 5 and 6) became the proto-Oquirrh basin, which extended eastward as a downwarp (known as the Doughnut trough) into the craton. Clastic material was eroded from the Roberts Mountain overthrust sheets of oceanic sediments and lava flows in central Nevada and was redeposited in eastern Nevada and western Utah as flysch. Renewed thrusting or uplift and erosion in the Antler belt during the Des Moinesian provided fine quartzose and chert clastic deposits to the Ely shelf in western Utah (fig. 8).

The Uncompahgre uplift in eastern Utah and southwestern Colorado (figs. 8 and 9) raised the Precambrian basement approximately 6,000 m during Des Moinesian through Early Permian time. The Paradox basin and Oquirrh basin were negative areas at the same time, and both received the clastic material eroded from the Uncompahgre uplift. Other positive elements of the ancestral Rocky Mountains had less influence on Utah depositional patterns but did provide some sand that crossed the carbonate banks. The Weber shelf and the Wells slope contained clastic sediments that were moving toward the Oquirrh basin.

Lower Permian depositional patterns were controlled by renewed uplift of the Uncompahgre in east-central Utah and by influx of chert-pebble conglomerates from the Antler belt in extreme northwestern Utah. Wolfcampian limestones and dolomites unconformably overlie Pennsylvanian carbonate deposits in southern and western Utah. Wolfcampian siltstones unconformably overlie Pennsylvanian sandstones in the Oquirrh basin, but the Permian contact is problematical within the Wells and Weber sandstones of the northeastern shelf. Red beds of the Halgaito Formation disconformably overlie the Pennsylvanian limestones in southeastern Utah.

The Upper Jurassic and Laramide overthrusts of the Sevier belt had a profound effect upon the distribution pattern of the Carboniferous rocks in the eastern Great Basin (fig. 1). The eastward piling of overthrust asymmetrical anticlines onto the Wasatch Hinge Line belt has telescoped lithofacies. Contrastingly, the decollement-type thrusting in the area of denudation associated with gneiss domes in eastern Nevada and western Utah has structurally reduced the thickness of Mississippian sections and scattered Pennsylvanian outcrops in a grandiose chaos of nappes. The one major palinspastic problem is the Oquirrh basin, because the amount of transport on the Charleston-Nebo thrust (fig. 1, D) is unknown. The allochthon is restricted to rocks of the Oquirrh basin and is separated from the thin eastern and southern Carboniferous sections by the Leamington tear fault (fig. 1, E) and the Charleston-Nebo thrust (fig. 1, D). The Leamington fault appears to follow the northeasterly ancestral break between the Redwall platform and the Doughnut trough (fig. 6).

LITHOSTRATIGRAPHY

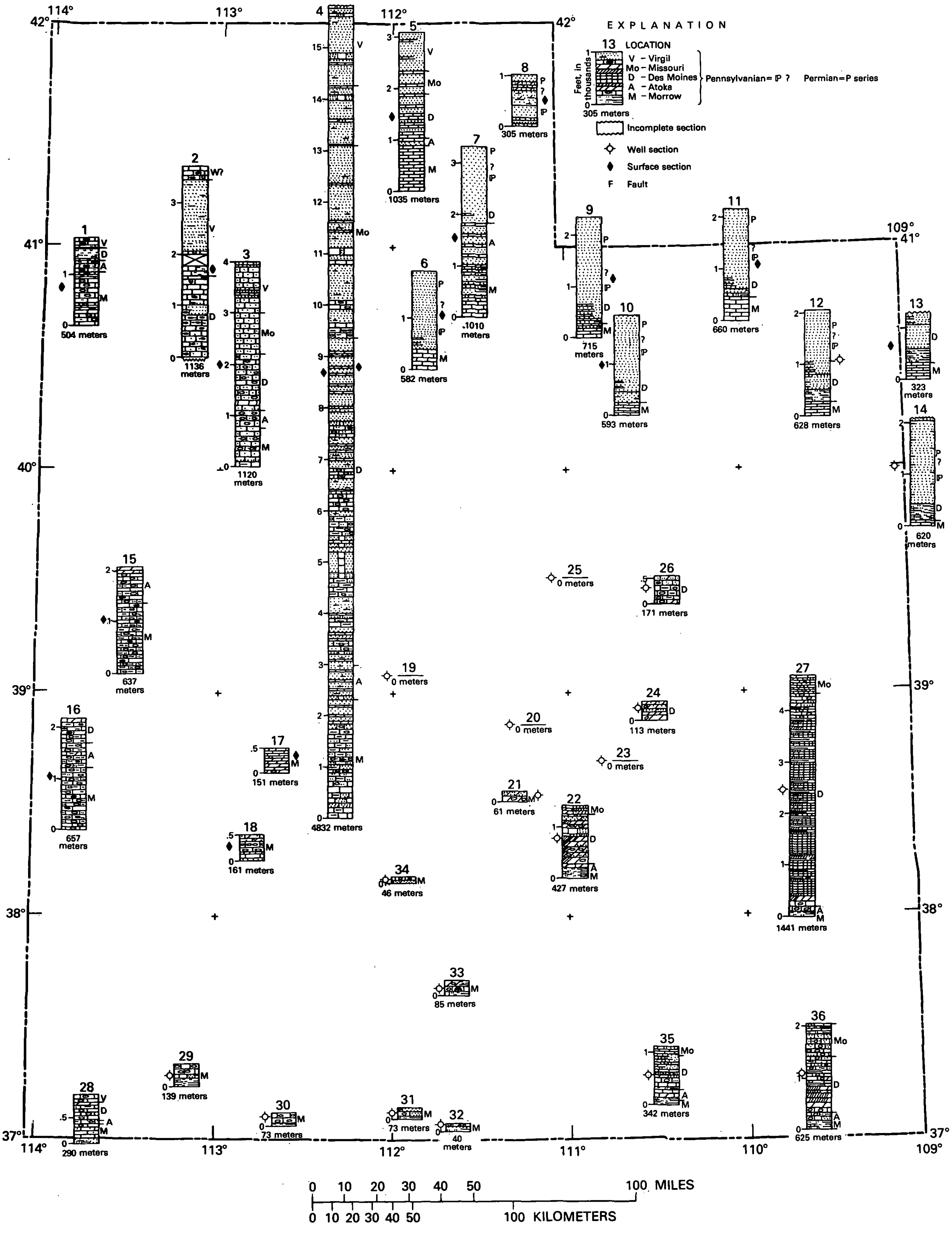
The Utah Carboniferous lithostratigraphic terminology is shown on the correlation chart in figure 3. Rock units and biostratigraphic zones (fig. 2) are shown in relationship to different biofacies and lithofacies. Regional stratigraphic sections of the Mississippian are illustrated in figure 4 and those of the Pennsylvanian, in figure 10. The Mississippian sections are divided into formational units, whereas the Pennsylvanian sections are divided into approximate series units.

MISSISSIPPIAN LITHOSTRATIGRAPHY

The Mississippian depositional history may be summarized as a succession of prograding carbonate banks and adjacent restricted reducing basins (figs. 5 and 6). The earliest Mississippian basins were a continuation of the restricted siltstone deposition in the Devonian Pilot, Pinyon Peak, and Leatham Formations (fig. 3). Carbonate deposition expanded across Utah by late Kinderhookian time, resulting in the time-equivalent limestone units of the Joana in the west, the Fitchville in the central, the Whitmore Wash in the southern, and the Lodgepole in the northern regions (fig. 3). The Whitmore Wash oolitic and algal limestones were deposited on a Bahamian-type platform, whereas the other time-equivalent units were deposited on open-marine shelves. The darker carbonate, gray chert, and widespread stromatolitic "Curley" bed of the Fitchville Limestone indicate deposition in a deeper, lower energy environment than the oolites and oncoliths of the Whitmore Wash Limestone. These differences in the limestones of the late Kinderhookian indicate that the Redwall carbonate bank was already differentiated from the basin to the northwest (fig. 5).

The Thunder Springs Member of the Redwall, Gardison, and upper parts of the Joana and Lodgepole Limestones are lithostratigraphic equivalents (fig. 3). All units are characterized by dark-gray thin-bedded limestones and abundantly bedded and nodular dark chert. These clinoform units were deposited in deeper water than the earlier Mississippian limestones. Thicknesses for this interval range from 15 m at Desert Creek (fig. 4, sec. 31) in the Four Corners area to 134 m at Elephant Canyon (fig. 4, sec. 18) in the Star Range. On parts of this slope, Waulsortian mounds accumulated within the Thunder Springs and the Joana. Where these encrinite mounds were dolomitized in the interior Redwall bank, they are porous petroleum reservoirs in the subsurface at South Last Chance, Ferron, and Big Flat fields (fig. 12, locs. 19, 17, 21).

At the end of the deposition of the Thunder Springs cherty limestones and equivalent units, subsidence in northwestern Utah accelerated, resulting in the sharp differentiation of the starved basin from the Redwall carbonate bank in southeastern Utah. From middle Osagean to early Meramecian, the carbonate bank prograded northwestward over this starved basin (fig. 5). The Mooney Falls and Horseshoe Mesa Limestones of the interior carbonate bank of the Redwall Limestone are equivalent to the prograded slope deposits of the Bullion Canyon and Yellowpine Limestone Members of the Monte Cristo Limestone in extreme southwestern Utah and the Deseret Limestone of central and northern Utah (fig. 3). The original edge of the carbonate bank is defined by fine clastic quartz in the Mooney Falls Member and lower part of the Deseret Limestone. These quartzose clastic deposits directly overlie the phosphatic shales of the starved basin. The phosphatic shales crop out in the basal Deseret Limestone at the Gilson, Tintic, Oquirrh, Wasatch, Pavant, and Wah Wah Mountains (fig. 1). The quartzose clastic deposits are present in the surface sections at Bradshaw Mountain (fig. 4, sec. 19) in the Mineral Range, at Cove Fort (fig. 4, sec. 20) in the Pavant Range, and in the subsurface wells near Meadow, Scipio Lake (fig. 4, sec. 21), North Springs, and Miller Creek (fig. 4, sec. 14). Clastic deposits and gamma-ray logs are the basis for defining the original edge of the starved basin (fig. 5). Biostratigraphic studies by Gutschick (1976) have confirmed the synchronous start of the starved basin in central Utah. Conodont studies by Sandberg and Gutschick (1977) show that the phosphatic shales of the basal parts of the Deseret, Chainman, and Little Flat are the time-stratigraphic equivalents of the late Redwall carbonate bank. The phosphorite beds of the starved-basin facies of the Deseret are overwhelmed by increased deposition of quartzose silt. The surface section at Cove Fort (fig. 4, sec. 20), the Scipio Lake (fig. 4, sec. 21) subsurface section, and the Shell Sunset Canyon subsurface section near Meadow in the Pavant Range (fig. 1, loc. 41) have a predominance of silt over carbonate material in the Deseret Limestone. The similarity of these three sections contrasts with the all-carbonate facies at the Elephant Canyon section (fig. 4, sec. 18) in the Star Range. The subsurface sections at North Springs, Mounds (fig. 4, sec. 15), and Miller Creek (fig. 4, sec. 14) south of Price are also in the basinal siltstone facies, whereas the subsurface section at Hiawatha (fig. 4, sec. 13) is in the dolomitized interior carbonate bank. The surface sections in the allochthon of the Charleston-Nebo thrust (fig. 1, thrust D) contain predominantly limestone above the phosphatic shales, which suggests that the siltstone of the Deseret was deposited parallel to the carbonate bank near the original basin margin by bypassing quartz from the craton to the northeast (fig. 5). No evidence exists for any quartzose source on the Redwall platform to the southeast. The Horseshoe Mesa Limestone, the uppermost member of the Redwall Limestone, progrades north-


EXPLANATION
13 LOCATION
Feet, in thousands
V - Virgil
Mo - Missouri
D - Des Moines
A - Atoka
M - Morrow
Pennsylvanian= IP ? Permian=P series
305 meters
Incomplete section
Well section
Surface section
F Fault
504 meters
1136 meters
1120 meters
4832 meters
1035 meters
582 meters
1010 meters
305 meters
715 meters
593 meters
660 meters
628 meters
323 meters
620 meters
637 meters
657 meters
151 meters
161 meters
0 meters
61 meters
427 meters
46 meters
113 meters
171 meters
1441 meters
85 meters
290 meters
139 meters
73 meters
40 meters
342 meters
625 meters
114°
113°
112°
111°
110°
109°
42°
41°
40°
39°
38°
37°
0 10 20 30 40 50 100 MILES
0 10 20 30 40 50 100 KILOMETERS

westward over the Deseret siltstone facies in the Price area and Pavant Range (fig. 1, loc. 41) sections.

In the interior of the Redwall carbonate bank, reflux dolomitization is common in the subsurface sections (figs. 4 and 5). This dolomitization is also present in outcrop in the Brazer Dolomite of the Crawford Mountains (fig. 4, sec. 8) in northern Utah and in the Redwall Limestone at Frenchman Mountain in southern Nevada. Superposition of imbricate thrusts in southern Nevada has juxtaposed the clinoform limestones of the Monte Cristo against the interior platform dolomite of the Redwall. In southwestern Utah, the Redwall or Monte Cristo Limestones reach their maximum thicknesses of 383 m in the Beaver Dam Mountains (fig. 4, sec. 25) and 409 m at Elephant Canyon (fig. 4, sec. 18) in the Star Range. The Yellowpine, Bullion Canyon, Anchor, and Dawn members of the Monte Cristo Limestone are recognized in these prograded carbonate deposits. From the High Plateau province to the Four Corners region, the Redwall Limestone thins approximately from 300 to 90 m. Parker and Roberts (1966), using the top of the Thunder Springs cherty limestone as a datum, demonstrated that there is both erosional wedging out and depositional thinning of the individual limestone members of the Redwall in a southeasterly direction. At Rockwood quarry, the nearest surface section in LaPlata County, Colorado, the Redwall (Leadville) Limestone is only 30 m thick (Baars and Knight, 1957). All the Mississippian rocks of the Redwall Limestone in southeastern Utah are on the carbonate bank and are oolitic, pelletal, birdseye micritic, stromatolitic, and fossiliferous. Dolomitization is variable and crosscuts lithologies, but it is spatially restricted to the interior of the carbonate bank (fig. 5).

In the Confusion Range (fig. 1, loc. 30) synclinorium of western Utah, there are well-exposed, unfaulted sections of the Carboniferous in the Needles Range (fig. 4, sec. 16), Burbank Hills (fig. 1, loc. 36), and Confusion Range (fig. 4, sec. 17). The Mississippian-Devonian boundary here is determined on the basis of conodont zones to be within the siltstones of the upper part of the Pilot Formation. For mapping, the base of the Joana Limestone is used as the base of the Mississippian. The Joana Limestone is approximately equivalent to the Lodgepole, Fitchville, and Gardison Limestones and the lower half of the Redwall and Monte Cristo Limestones (fig. 3). Most stratigraphers have traditionally placed a

FIGURE 10.—Numbered stratigraphic sections of the Pennsylvanian in Utah. Series subdivisions and generalized lithologies are shown. Where no section number is supplied, the stratigraphy is a composite of information from more than one section. Stratigraphic sections 1 and 15–36 are based on original unpublished data of Welsh; the other sections are modified from published sources as indicated.

1. Rishel Peak, T. 1 N., R. 18 W.
2. Lakeside Mountains, T. 2 N., R. 11 W. (modified from Doelling, 1964)
3. Cedar Mountains, T. 4 S., R. 10 W. (modified from Maurer, 1970)
4. Oquirrh Mountains, T. 4 S., Rs. 3–5 W. (modified from Welsh and James, 1961)
5. Wellsville Mountain, T. 10 N., R. 1 W. (modified from Williams, 1948)
6. Mt. Aire, T. 1 S., R. 2 E. (modified from Crittenden, 1959)
7. Weber Canyon, T. 4 N., R. 3 E. (modified from Bissell and Childs, 1958)
8. Crawford Mountains, T. 11 N., R. 8 E. (modified from Sando and others, 1959)
9. Deadman Mountain, T. 1 N., R. 11 E. (modified from Bissell and Childs, 1958)
10. Duchesne River, sec. 14, T. 1 N., R. 8 W. (modified from Sadlick, 1957)
11. Sols Canyon, sec. 11, T. 2 N., R. 18 E. (modified from Sadlick, 1957)
12. Ute Federal, sec. 12, T. 4 S., R. 22 E. (modified from Sadlick, 1957)
13. Whirlpool Canyon, sec. 27, T. 3 S., R. 25 E. (modified from Sadlick, 1957)
14. Watson, sec. 34, T. 9 S., R. 25 E. (modified from Sadlick, 1957)
15. Skunk Springs, T. 17 S., R. 16 W.
16. Needles Range, T. 25 S., R. 19 W.
17. Cove Fort, T. 24 S., R. 6 W.
18. Bradshaw Mountain, T. 29 S., R. 10 W.
19. Scipio Lake, sec. 14, T. 20 S., R. 2 W.
20. Emery, sec. 34, T. 22 S., R. 5 W.
21. South Last Chance, sec. 18, T. 26 S., R. 7 E.
22. Cainville, sec. 29, T. 28 S., R. 8 E.
23. San Rafael, sec. 28, T. 24 S., R. 10 E.
24. Sinbad, sec. 5, T. 22 S., R. 12 E.
25. Hiawatha, sec. 13, T. 15 S., R. 7 E.
26. Grassy Trail, sec. 1, T. 16 S., R. 12 E.
27. The Knoll, sec. 11, T. 26 S., R. 19 E.
28. Beaver Dam Mountains, T. 42 S., R. 18 W.
29. LaVerkin, sec. 30, T. 40 S., R. 12 W.
30. Kanab, sec. 2, T. 43 S., R. 8 W.
31. Kaibab Gulch, sec. 34, T. 42 S., R. 2 W.
32. Judd Hollow, sec. 19, T. 43 S., R. 2 E.
33. Upper Valley, sec. 12, T. 36 S., R. 1 E.
34. Antimony Canyon, sec. 30, T. 30 S., R. 2 W.
35. Nokai, sec. 27, T. 40 S., R. 12 E.
36. Lime Ridge, sec. 28, T. 40 S., R. 20 E.

regional unconformity at the top of the Joana Limestone in western Utah and eastern Nevada. Sadlick (1965) followed this interpretation but was first to recognize a fondothem facies in the overlying Chainman Formation. Current interpretation (Gutschick, 1976) is that the lowermost part of the Chainman Formation represents the starved-basin facies and that its phosphatic shales are equivalent to the lower part of the Deseret Limestone (fig. 3). Most of the terrigenous clastic deposits of the Chainman Formation are turbidite shales and silts derived from the Antler Highland about 161 km farther west. Sadlick (1965) divided the Chainman Formation into six lithostratigraphic members. In ascending order, they are: the Needles Siltstone, Skunk Springs Limestone, Camp Canyon, Donner, Willow Gap Limestone, and Jensen. The Donner has not yet been recognized in Utah. The Needles Siltstone and Skunk Springs Limestone Members and the shale and siltstone in the lower part of the Camp Canyon Member represent the basinal infillings. The phosphatic shale in the Needles Siltstone Member above the Joana Limestone represents the time equivalent of the Osagean-lower Meramecian carbonate bank farther east (Rose, 1976b). The shale in the lower part of the Camp Canyon Member in the Confusion Range is a restricted basinal facies, which is equivalent to, and is covered by, the upper Meramecian-Chesterian carbonate bank of the Great Blue and Ochre Mountain Limestones (fig. 6). The Woodman Formation (fig. 3) of the Gold Hill district (fig. 1, loc. 19) is approximately equivalent to the Needles Siltstone Member and the lower part of the Camp Canyon Member of the Chainman Formation.

The upper part of the Camp Canyon Member and the Willow Gap Limestone Member of the Chainman Formation are approximately equivalent to the Ochre Mountain or Great Blue Limestones (fig. 3). In the Confusion Range synclinorium, the facies on the east limb are markedly different from those on the west limb. At Skunk Springs (fig. 4, sec. 17) and in the Burbank Hills (fig. 1, loc. 36), coarse clastic limestones of the Camp Canyon and Willow Gap Limestone Members represent the westernmost exposures of the Chesterian carbonate bank. In the Needles Range (fig. 4, sec. 16) and on the west side of the Confusion Range, the Chainman Formation is entirely within the basinal facies. The limestones of the Chainman Formation are black calcilutites, whereas the Great Blue carbonate bank consists of light-gray calcarenites.

In northwestern Utah, a decollement in the Chainman Formation has complicated the stratigraphic sequence of the Mississippian rocks. Thicknesses of lithostratigraphic units are variable because younger units have been thrust over older units. Sadlick (1965) reported that the lower part of the Camp Canyon Member of the Chainman rests unconformably upon either the Joana Limestone or the Pilot Formation in the Silver Island Mountains (fig. 4, sec. 1). Earlier, Sadlick and Schaeffer (1959) had interpreted this observation as evidence for their Wendover phase of the Antler orogeny. The present writers, recognizing the structural complications in northwestern Utah, suggest that the Mississippian stratigraphy needs to be reexamined. At this time, no reliable sections have been published for Mississippian deposits in northwestern Utah or northeastern Nevada.

In northern Utah, the Tintic nomenclature of Morris and Lovering (1961) is applicable to all the outcrops west of Logan (fig. 1). The Lodgepole Limestone of the Logan area is equivalent to the Fitchville and Gardison Limestones of the East Tintic Mountains (fig. 1, loc. 33). The Deseret Limestone and Humbug Sandstone are present in both the allochthonous and autochthonous sequences of the Wasatch, Uinta, and Basin and Range Mountains (fig. 1). The Great Blue Limestone and Manning Canyon Formation terminology has been extended into the Wellsville Mountains (fig. 1, loc. 16) west of Logan and into the Deep Creek Mountains of southeastern Idaho. To the east in the Uinta and Wasatch Mountains, the Upper Mississippian rocks of late Meramecian and Chesterian age are better termed the Doughnut Formation rather than the Great Blue or Manning Canyon. The Doughnut Formation includes the rocks of late Meramecian through Chesterian age that are reduced in stratigraphic thickness and occupy the stratigraphic position between the Humbug Sandstone and the Pennsylvanian Round Valley Limestone (fig. 3). This abbreviated Upper Mississippian section extends eastward in the outcrop to Whiterocks Canyon (fig. 4, sec. 11) along the south flank of the Uinta Mountains and into the subsurface at Mounds (fig. 4, sec. 15) in the northern San Rafael area near Price. A similar section of Upper Mississippian rocks is present in southwestern Utah in the southern Wah Wah Mountains (fig. 1, loc. 38) and northern Beaver Dam Mountains (fig. 1, loc. 44). Both these sections are allochthonous and are equivalent to the Battleship Wash Limestone and Indian Springs Formation in southern Nevada (fig. 3).

In Old Laketown Canyon (fig. 4, sec. 7) at the southeast corner of Bear Lake in northern Utah, Sando and others (1976) have described an Upper Mississippian section that is similar to the Skunk Springs (fig. 4, sec. 17) section in the Confusion Range in western Utah. The section above the Lodgepole Limestone includes the Little Flat Formation and the Monroe Canyon (Great Blue) Limestone. The lower phosphatic shales of the Little Flat Formation are equivalent to the starved basin facies of the lower part of the Deseret Limestone, and the upper siltstones of the Little Flat are equivalent to the Humbug Sandstone. The Monroe Canyon Limestone has prograded over the siltstone of the Little Flat Formation as the Great Blue-Ochre Mountain Limestones have prograded over the Chainman or Woodman Formations in western Utah.

Shale of the Manning Canyon Formation was deposited in deltaic, estuarine, and near-shore marine environments. The formation extends from the East Tintic Mountains (fig. 1, loc. 33) northward into southern Idaho. In contrast to the Diamond Peak and Chainman Formations that are flysch that filled the Antler Foreland basin west of the Great Blue carbonate bank, the Manning Canyon Formation consists of clastic deposits that prograded westward through the Doughnut trough across the interior of the Chesterian carbonate bank (fig. 6). The deposition of these clastic sediments on the Upper Mississippian carbonate bank in Utah caused a swamp to form near sea level. Shale deposited in temporary swamp environments had earlier encroached upon the Great Blue carbonate bank during deposition of the Long Trail, Chiulos, and Herat Shales (fig. 3). The Manning Canyon Formation and the upper part of the Chainman Formation eventually buried the carbonate bank in shale and sandstone. Contemporaneously during the Late Mississippian, a lateritic regolith formed on the exposed Redwall subaerial platform to the southeast. The present edge of the erosional wedge of the Upper Mississippian rocks is shown in figure 6.

PENNSYLVANIAN LITHOSTRATIGRAPHY

Key stratigraphic columns of surface and subsurface sections of the Pennsylvanian are illustrated on figure 10 and the approximate series time units are designated. Formational names of rock units are shown in the correlation chart (fig. 3). Figures 7–9 are paleogeographic maps of Utah in Early, Middle, and Late Pennsylvanian time.

Morrowan and Atokan (fig. 7) deposits are well represented throughout Utah; the thickest deposit is 1,000 m of calcilutite in the Oquirrh basin. On the unstable Ely shelf, 350 to 425 m of cyclical calcarenite, calcisiltite, and calcilutite was deposited. Less than 150 m of pelletal and birdseye calcilutite, oolitic calcarenites, and biostromal calcirudite was deposited on the Bahamian-type Callville platform. Coral biostromes are common rocks deposited in all three environments.

The West Canyon Limestone in the Oquirrh basin was deposited in deeper water than deposits on the Ely shelf or Callville platform, but the environment was still in the photic zone. This lower limestone of the Oquirrh Group has been recognized north to the Utah-Idaho border in Cache and Box Elder Counties, where the Oquirrh and Sublette basins merge. No evidence is preserved to indicate that the Oquirrh basin of Utah and the Bird Spring basin of southern Nevada were connected along the Wasatch Hinge Line; instead, the connection between these basins was farther west. In both basins, marine carbonate deposition was continuous from Morrowan through Atokan to Des Moinesian time.

Southeastern Utah was invaded by Early Pennsylvanian seas which reworked the lateritic regolith on the Redwall carbonate platform into the Molas Formation. Chert-pebble conglomerates, derived from Paleozoic rocks stripped during the initial uplift of the Uncompahgre, are common in the lower part of the Molas in southwestern Colorado, but red siltstones are more characteristic in southeastern Utah. Reduction in the marine environments produced green shales and siltstones. Overlying the time-transgressive clastic units of the Molas Formation is a predominantly carbonate section of Atokan and earliest Des Moinesian age; this section is generally placed in the Pinkerton Trail Formation, although a disconformity probably exists between the Atokan and Des Moinesian in southeastern Utah.

In northern Utah on the Round Valley shelf (fig. 7), which overlies the Mississippian Doughnut trough (fig. 6), sedimentation was continuous or only slightly interrupted from Chesterian to Morrowan time. The Round Valley Limestone of Morrowan age represents a marine invasion over the deltaic and estuarine environments of the Doughnut Formation. Because the Upper Mississippian rocks in northern Utah contained only a few lateritic beds, the Round Valley does not contain a basal red-bed unit comparable to that in the Molas Formation. The Round Valley Limestone is the shelf equivalent of the West Canyon Limestone of the Oquirrh basin.

By Atokan time, the seas had submerged the entire State of Utah, and carbonate deposition was dominant. The only clastic material being deposited was interbedded with limestone in southeastern

Utah near the Uncompahgre (fig. 8). Atokan rocks are not reported to be in the Round Valley Limestone or Hells Canyon Formation in the Uinta Mountains; however, Bissell and Childs (1958) reported that *Fusulinella* is present 250 m above the base of the type section of the Weber Sandstone northeast of Morgan (fig. 10, sec. 7). The absence of reports of Atokan rocks in the Uinta Mountains may be the result of nondeposition, pre-Des Moinesian erosion, or a lack of fusulinids in the rocks. Probably some rocks in the Hells Canyon and Morgan Formations are Atokan in age (fig. 3).

The Des Moinesian depositional patterns were strongly affected by the eroding of the Uncompahgre Mountains and the sinking of the adjacent Oquirrh, Paradox, and Eagle basins (fig. 8). Erosion of the Precambrian crystalline rocks resulted in thick arkosic alluvial fans that intertongued with sabka evaporites and euxinic shales. Northwest of the Uncompahgre, the Oquirrh basin was a major depocenter for fine arkosic sandstones. Submarine sandstones episodically prograded over the carbonates of the basin. North of the Uncompahgre, the sands were distributed as a thick uniform blanket on the Hells Canyon-Morgan shelf and then passed down the Morgan slope into the Oquirrh basin. The Oquirrh basin received 1,900 m of Des Moinesian strata in a sandstone-to-limestone ratio of 1:1. Actually, detrital quartz is present in the limestone as well as in the sandstone because many of the limestones are calcisiltites. This influx of sand definitely had its source in the Uncompahgre uplift.

The Callville platform remained a broad, stable, Bahamian-type environment. Biostromes flourished along its western and southeastern edges. Algal stratigraphic reefs were a barrier across the only access into the Paradox basin during salt deposition at Aneth in southeastern Utah (fig. 8). Smaller algal patch reefs are in limestones equivalent to the evaporites in the outcrops of the Hermosa Group in the San Juan River Canyon (fig. 1, loc. 45). The western margin of the Paradox basin was intermittently a subaerial tidal flat, and the Paradox salt units are represented by diastems in the limestone sequence. Figure 8 shows the inner depositional edge of halite and the outer depositional edge of anhydrite. Primary dolomite was precipitated contemporaneously with gypsum; however, extensive secondary dolomitization of the Callville Limestone around the Emery high is related to reflux replacement below the Wolfcampian unconformity. Black organic dolomitic shales interbedded with the halite beds of the Paradox Formation are the principal hydrocarbon source rocks. Anhydrite beds of the Eagle basin had a similar depositional history north of the Uncompahgre. Most of the sabka evaporites are in Colorado. The Paradox Formation, containing a maximum of 1,200 m of evaporites, represented a very short period of the early Des Moinesian. Deposition of these evaporites was followed by a sudden extensive marine invasion which produced the widespread Desert Creek and Ismay Limestones which have equivalents in all areas of Utah, except where removed by pre-Wolfcampian erosion.

The Morgan shelf of the Uinta Mountains and northern Utah (fig. 8) received clastic material throughout the Des Moinesian. The Hells Canyon Formation on the south flank of the Uintas is thin-bedded red-gray-purple shale, siltstone, sandstone, and fossiliferous limestone containing early Des Moinesian fusulinids. The overlying Morgan Formation is predominantly reddish-brown siltstones and sandstones and thin limestone. Estuarine environments are common in both formations. The red beds and clastic deposits of the Morgan are approximately equivalent to the evaporites of the Paradox Formation. The reddish Morgan is overlain by and is laterally equivalent to the yellowish-gray rocks in the lower part of the Weber Sandstone. These yellowish-gray rocks are overlain by gray fossiliferous limestone which is also in the lower part of the Weber Sandstone and is equivalent to the post-Paradox Desert Creek and Ismay Limesones and unnamed limestones (fig. 3).

The Ely shelf in western Utah received increasing quantities of very fine quartz and chert silt from the Antler belt. Calcisiltites of the Hogan Member of the Ely Limestone (Robinson, 1961) are as much as 70 percent silica, as fine clastic quartz and chert and spicules. Chert-pebble conglomerates are present locally in the Des Moinesian sections 161 km farther west in central Nevada. The Des Moinesian lithologies are cyclical, like those of the Morrowan and Atokan, but calcarenites make up a much smaller percentage of the rock column. Calcisiltites also characterize the Des Moinesian rocks of the Bird Spring basin in southeastern Nevada.

A disconformity exists between the Middle and Upper Pennsylvanian deposits in all the marine sequences in Utah, but the boundary is undefined in the Weber Sandstone and in the arkoses adjacent to the Uncompahgre uplift.

Missourian and Virgilian time was a period of accelerated uplift of the Uncompahgre (fig. 9). Continental sedimentation on alluvial fans was continuous in the Late Pennsylvanian, and more than 1,500 m of arkose was deposited adjacent to the south flank of the Uncompahgre Mountains. Much of this arkose was reworked by marine currents and bypassed the Weber shelf to be deposited in the Oquirrh basin. The Jordan and Commercial Limestones at the base of the Bingham Mine Formation are the oldest Missourian rocks in the Oquirrh basin. The Jordan Limestone rests unconformably upon the Des Moinesian Butterfield Peaks Formation (fig. 3). Above the Commercial Limestone, more than 1,200 m of arkosic sandstone is in the 2,000-m-thick Bingham Mine Formation (fig. 10, sec. 4). Even though the ratio of clastic material to limestone is 4:1, subsidence exceeded the rapid deposition in the Oquirrh basin. Fine-grained laminated sandstone containing trace fossils on bedding planes suggests low-energy below-wave-base deposition. In contrast, sandstones in the Weber are tabular and crossbedded, indicating a high-energy shelf environment. At the base of the Wolfcampian Curry Peak Formation, a polymictic carbonate, chert-pebble conglomerate containing reworked Pennsylvanian silicified fossils marks the Lower Permian boundary in the Oquirrh Mountains (fig. 3). This boundary is not defined in the Weber Sandstone.

Upper Pennsylvanian limestones and clastic sedimentary rocks in southeastern Utah are named the Honaker Trail Formation (fig. 3). In the type section at Honaker Trail, no Virgilian rocks are present, and red beds of the Permian Halgaito Formation unconformably overlie Missourian limestones. In the Paradox basin, Missourian rocks rest unconformably upon Des Moinesian limestones that have been more deeply eroded toward the west margin of the Paradox basin where the Honaker Trail Formation rests unconformably upon the Desert Creek Limestone. Virgilian rocks are restricted to the Moab trough (fig. 9). Erosion on the pre-Wolfcampian unconformity has removed much of the Upper Pennsylvanian sequence in southwestern and south-central Utah. Now in the Paradox basin, the Wolfcampian Elephant Canyon Limestone to the east and the Pakoon Dolomite to the west unconformably overlie, respectively, the beveled Pennsylvanian rocks of the Hermosa Group and Callville Limestone. Near St. George (fig. 10, sec. 28), a small area contains Virgilian limestones in the upper part of the Callville; no Missourian rocks have been reported. Similar calcarenites and calcilutites of the Virgilian series are present on the Callville platform in southern Nevada and on the Ely shelf in northeastern Nevada and northwestern Utah. At Frenchman Mountain near Las Vegas, clinoform Virgilian limestones are transitional between the Callville platform and the Bird Spring basin. The Pakoon Dolomite overlies unconformably the Pennsylvanian Callville Limestone everywhere in southwestern Utah, except at Scipio Lake where the Kaibab Formation rests unconformably upon the Mississippian in a subsurface section (fig. 10, sec. 19).

The Weber Sandstone, which is 700 m thick at its type locality in Weber Canyon (fig. 10, sec. 7) northeast of Morgan (Bissell and Childs, 1958), includes both Upper Pennsylvanian and Lower Permian sandstone (fig. 3). The highest Des Moinesian fusulinids are present approximately 300 m above the base, so the upper 400 m may be either Late Pennsylvanian or Permian. Bissell and Childs (1958) also indicated that the Weber Sandstone is 310 m thick along the Duchesne River (fig. 10, sec. 10) on the south flank of the Uinta Mountains. Poorly preserved *Triticites* are reported to be 90 m above the base of the Weber Sandstone in the Duchesne River section (Bissell and Childs, 1958). An accurate age assignment was impossible because of the poor preservation. *Schwagerina* was also collected in the same study in the upper 30 m of the Weber Sandstone at the Morris Ranch section northeast of Vernal (fig. 1). Bissell estimates that 100 to 200 m of the lower Weber Sandstone in northeastern Utah is Pennsylvanian and that the remaining part is Permian.

In west-central Utah, Wolfcampian Riepe Springs Limestone containing a thin basal chert-pebble conglomerate overlies unconformably the Atokan or Des Moinesian Ely Limestone. In northwestern Utah, Virgilian limestones rest unconformably upon the Des Moinesian. Chert-pebble conglomerates of the Virgilian (?)-Wolfcampian Strathern Formation overlie fusulinid-bearing Virgilian limestones. Most of these conglomerates are Wolfcampian; however, Schaeffer (1960) included those in the Silver Island Range in the Virgilian. These conglomerates are well exposed at Rishel Peak (fig. 10, sec. 1) north of Wendover and in the Spruce Mountain area of northeastern Nevada. Detailed descriptions of the Pennsylvanian rocks of the Gold Hill district have not been published, but the district does contain Morrowan, Atokan, Des Moinesian, and Virgilian limestones that are not of the Oquirrh facies. The Wolfcampian Ferguson Mountain Formation, not the Strathern Formation, overlies the Virgilian limestones at Gold Hill.

Late Pennsylvanian fusulinids have been reported in isolated ranges in northern Utah. Maurer (1970) reported a thin, 30- to 120-m Missourian section in the Cedar Mountains (fig. 10, sec. 3); Rigby (1958) reported 2,000 m of Missourian in the Stansbury Mountains (fig. 1, loc. 22); and Williams (1948) reported 900 m of Missourian and Virgilian in the Wellsville Mountains (fig. 1, loc. 16). Reported fusulinid-bearing strata of Virgilian age are more widespread than those of Missourian age in the Basin and Range province.

The lithostratigraphy of Upper Pennsylvanian rocks in northern Utah has not been sufficiently investigated to synthesize a meaningful paleogeography for the Oquirrh basin, the Wells slope, or the Weber shelf. Stratigraphy of the Pennsylvanian rocks in the northwest quarter of Utah is described sufficiently for general characterization in only two ranges: the Silver Island Range (fig. 10, sec. 1) and the Cedar Mountains (fig. 10, sec. 3). Stratigraphic knowledge is lacking because (1) all the rocks between the Upper Mississippian strata and the Upper Permian Phosphoria Formation have been lumped into the Oquirrh Formation, and (2) denudation faulting has caused the rocks of the Oquirrh Formation to have a chaotic present distribution. Graduate students have not adequately lithologically divided the Oquirrh Formation in order to resolve the structural complexities in northwestern Utah. Rock sequences have only been grossly assigned to time series on the basis of scattered fusulinid collections. Even series designations are few and scattered in the Grouse Creek and Goose Creek Mountains and Raft River Range (fig. 1) of northwesternmost Utah.

The Silver Island Range north of Wendover (fig. 1, loc. 4) has excellently exposed sections of the Pennsylvanian at A-1 Canyon and Rishel Peak (fig. 10, sec. 1). Schaeffer (1960) did not divide the cyclical limestone, but Morrowan, Atokan, and Des Moinesian calcarenites, calcisiltites, and calcilutites are disconformably overlain by Virgilian-Wolfcampian dolomites and chert-pebble conglomerates of the Strathern Formation. Des Moinesian calcisiltites of the Hogan Member of the Ely Limestone are distinctive. Anderson (1957) reported that limestones at Crater Island (fig. 1, loc. 5) may be Virgilian and mentioned that two lithologies of Permian age there rest unconformably upon the conglomerate of the Mississippian Diamond Peak Formation. He described thinning of the Chainman Formation from 365 to 0 m and of the Joana Limestone from 10 to 0 m along strike. Neither Schaeffer (1960) nor Anderson (1957) recognized the decollement thrusting in the Silver Island Range that has caused structural thinning of the shale in the Chainman Formation and of the Joana Limestone. This thrusting juxtaposed different Pennsylvanian and Permian rocks upon the decollement surface.

Paddock (1956) reported that the Leonardian Pequop Formation unconformably overlies the Devonian Stansbury carbonate conglomerate in the Newfoundland Mountains (fig. 1, loc. 6). In the southern Grouse Creek Mountains (fig. 1, loc. 46), the Wolfcampian Ferguson Mountain and Strathern Formations are unconformable upon the Chainman Formation. These omissions of Pennsylvanian sequences have been interpreted previously by several investigators as evidence for a northwest Utah highland during the late Paleozoic. However, firm paleogeographic conclusions should not be drawn from the incomplete stratigraphic data in northwestern Utah until structural studies are completed because denudation faulting above the Chainman decollement has placed structurally different sequences of Pennsylvanian through Triassic rocks as discrete nappes upon both the Mississippian Chainman Formation and Devonian carbonates. Adjacent to the Raft River gneiss dome in the Goose Creek Mountains and Raft River Range, Compton (1972, 1975) has mapped Pennsylvanian rocks in thrust contact with regionally metamorphosed Precambrian, Cambrian, and Ordovician rocks.

Stifel (1964) mapped much of the Terrace and Hogup Mountains (fig. 1, locs. 9, 10) as Oquirrh, but the sequence is better interpreted as Leonardian Pequop Formation, Wolfcampian Ferguson Mountain Formation, and unnamed Virgilian rocks. R. C. Douglass (unpub. data, 1972) reported *Pseudofusulinella* and *Triticites* from collections of R. R. Compton (1975) in the Raft River Mountains (fig. 1, loc. 2). Compton has used the term "Oquirrh" in the Raft River Mountains on the basis of paleontological rather than lithological correlation. Doelling's (1964) section (fig. 10, sec. 2) in the Lakeside and Grassy Mountains (fig. 1, loc. 7) and Maurer's (1970) section (fig. 10, sec. 3) in the Cedar Mountains document that at least this far west, the lithofacies of the Pennsylvanian type Oquirrh are still present. The proportion of clastic material to limestone has, however, dropped drastically, further supporting the thesis that the fine arkosic sandstone of the type Oquirrh was derived from the Uncompahgre uplift. Outcrops in the Cedar (fig. 10, sec. 3) and Lakeside (fig. 10, sec. 2) Mountains are definitely the farthest west for which the term

"Oquirrh" should be used for Pennsylvanian rocks. Maurer (1970) and Doelling (1964) reported fusulinids representing all the Pennsylvanian series, but they were unsuccessful in attempting to correlate the sections in the Cedar and Lakeside Mountains lithologically with the type Oquirrh Formation in the Oquirrh Mountains (fig. 10, sec. 4) or even between their two adjacent ranges.

Olson's dissertation (1960) on the Promontory Mountains (fig. 1, loc. 12) reported Morrowan, Atokan, and Des Moinesian fusulinids, but his lithostratigraphic data are grossly generalized as 2,150 m of sandstone and limestone bearing Morrowan fossils near the base and Wolfcampian fossils above. About 50 km north in the Tremonton-Portage-Clarkston Mountains area, Bissell reports that the Morrowan West Canyon Limestone is 400 m thick and that the remaining Pennsylvanian and Permian sequence below the Phosphoria Formation is approximately 1,400 m thick. Bissell (*in* Peace, 1956) reported Atokan and Morrowan fusulinids from the subsurface at Rozel Point and Curlew Valley (fig. 1, locs. 11, 3). The documented widespread distribution of Lower and Middle Pennsylvanian rocks in northwestern Utah further indicates that the hypothetical northwest Utah paleohigh interpreted from Paddock's (1956) observation of Leonardian rocks resting upon Devonian rocks in the Newfoundland Mountains (fig. 1, loc. 6) probably did not exist.

The use of the name Wells Formation for Pennsylvanian rocks in the Logan area of northern Utah is another problem of terminology that has yet to be resolved. Richards and Mansfield (1912) named the formation in Wells Canyon in Bannock, Idaho, for 740 m of limestone and sandstone below the Phosphoria Formation. Williams (1948) reported Morrowan, Des Moinesian, Missourian, and Virgilian fusulinids in 1,800 m of "Wells" (Oquirrh) Formation at Wellsville Mountain. Nygreen (1958) was the first to show that the West Canyon Limestone was present at the Dry Lake section at Wellsville Mountain. Bissell states that recent fusulinid studies at Wellsville Mountain (fig. 10, sec. 5) indicate that the Oquirrh Formation has approximately 270 m of Morrowan and Atokan, 340 m of Des Moinesian, 135 m of Missourian, and 290 m of Virgilian. The Wells Formation is now best restricted to the Bear River Range (fig. 1, loc. 17) and Crawford Mountains (fig. 10, sec. 8), where 150–300 m of sandstone, limestone, and dolomite crop out. The correlation with the type area in Wells Canyon, Idaho, is not documented.

CARBONIFEROUS BIOSTRATIGRAPHY

The Carboniferous formations of Utah contain faunas from the earliest Kinderhookian to the youngest Virgilian and essentially all the paleontological zones are represented in carbonate bank, euxinic, deltaic, estuarine, or basin facies. All invertebrate groups and a variety of floras are well represented within the various environments. Taxonomy of the invertebrates is incomplete, but index fossils identified to genera and some species are listed on the biostratigraphic chart (fig. 2).

The earliest Kinderhookian age assignments are based on conodonts in the Fitchville Limestone and the Pilot and Leatham Formations. In the siltstone facies of the Pilot and Leatham, Kinderhookian conodonts overlie the *Syringothyris* zone of Late Famennian age. Late Kinderhookian and early Osagean brachiopod and coral faunas are abundant in limestones of the Whitmore Wash and Thunder Springs Members of the Redwall Limestone and in the Fitchville, Gardison, Joana, and Lodgepole Limestones. Conodonts and endothyrids have also proven useful for zonation in these limestones. The "Curley" bed at the top of the Fitchville Limestone is a widespread stromatolite horizon (Proctor and Clark, 1956). Algal limestones having birdseye textures are also common in the Redwall carbonate bank. Waulsortian crinoidal banks are present in the Thunder Springs Member of the Redwall Limestone in the Paradox basin, and some are also present in the Joana Limestone in eastern Nevada. The bedded cherts of the Thunder Springs and Gardison were derived from abundant siliceous sponges that flourished during times of slightly deeper water than was present during most other times of carbonate deposition.

Beginning in late Osagean time, subsidence produced a starved basin across northwestern Utah. The southern margin of this basin trended northeast from lat 37°45′N. to 40°15′N. (fig. 5), and the Redwall carbonate bank was to the southeast. This separation into two distinct environments produced a marked change in the faunal realms. Sandberg and Gutschick (1977) have shown by conodont zonation that the environmental differentiation began in central Utah at the top of coral zone C_1 in the carbonate bank, and Sadlick (1965) showed that it continued to the top of cephalopod zone P_1, or the top of coral zone F in western Utah. The pre-E zone brachiopods of the starved basin are *Quadratia hirustiformis*, Leiorhyncoidea, and Orbiculoidea. Trace fossils are common. *Goniatites crenistriae* and

G. multiliratus are the indices of the waning stage of the starved basin in western Utah during the late Meramecian.

Dorlodotia inconstans formed widespread coral biostromes during the early Meramecian on the Redwall carbonate bank. These zone D corals are the youngest found in the Redwall bank in southern Utah because the bank was subsequently subaerially exposed until the Morrowan time. The *Ekvasophyllum* corals of zone E are the youngest fauna in the prograded Deseret carbonate bank in central Utah.

In western Utah, a flysch facies of submarine fans gradually filled the starved basin and by Chesterian time, the basin contained a prolific molluscan fauna of goniatites, belemnites, pelecypods, and gastropods. *Goniatites granosus* is the index fossil for the earliest Chesterian; *Cravenoceras* and *Eumorphoceras* are present higher in the sequence.

Corals and brachiopods dominated the megafauna of the Great Blue carbonate bank (fig. 6) which extended almost to the western boundary of Utah. *Cravenoceras* cephalopods are mixed with the coral and brachiopod faunas in calcarenite outcrops at Burbank Hills and the Confusion Range in western Utah (fig. 1, locs. 36, 30). *Faberophyllum* and *Straitifera brazeriana* are common in the lower part of the Great Blue Limestone overlying the nearly barren Humbug Sandstone of the Doughnut trough. The upper Great Blue Limestones have *Caninia excentrica* and *Spirifer brazerianus* as index fossils. Periodically, the interior carbonate bank became emergent, and prodeltaic shales such as the Long Trail Shale of the Oquirrh Mountains and the Herat Shale at Gold Hill were deposited. These shales have *Lepidodendron* plant imprints associated with hematitic regoliths (Chamberlain, 1978). The reestablishment of the carbonate bank in the Late Mississippian continued intermittently until finally the bank was buried by deltaic clastic deposits of the Manning Canyon Formation.

The Manning Canyon Formation and the Jensen Member of the Chainman Formation completely buried the Great Blue carbonate bank with late Chesterian deltaic and estuarine deposits containing a mixed molluscan, bryozoan, and brachiopod fauna in limestone interbedded with carbonaceous shale and siltstone. *Diaphragmus, Archimedes,* and spiriferoids are common. *Eumorphoceras* and *Rayenoceras* are present. *Sigillaria* roots and *Lepidodendron* and *Stigmaria* imprints as well as a diversified Lycopodophyta flora are present in many localities. Tidwell and others (1974) stated that floras in sandstone of the upper part of the Manning Canyon Formation and in sandstone of the upper part of the Diamond Peak Formation are generally Pennsylvanian (Namurian B) in age. Stratigraphically lower floras in the Doughnut and Indian Springs Formations are Chesterian in age.

A return to open-marine circulation at the end of Mississippian time produced a widespread coarse detrital limestone containing the *Rhipidomella nevadenis* zone throughout the miogeosyncline area and into the Doughnut trough. Rocks of this zone overlie the highest rocks from which *Eumorphoceras bisulcatum* and the Pennsylvanian (Namurian B) floras were collected. Because the *R. nevadensis* zone is generally referred to as Chesterian, a discrepancy exists between the age indicated by the flora and that indicated by the invertebrates at the Mississippian-Pennsylvanian boundary.

The Morrowan seas reworked the regolith on the Redwall bank and expanded southeastward. The Doughnut trough was no longer recognizable. Algal limestones accumulated on the very shallow Callville platform (fig. 7). On the Morgan Shelf (fig. 9) in northern Utah and on the Hermosa Shelf in southeasternmost Utah, thin clastic limestones containing a brachiopod-bryozoan-crinoid biofacies were deposited. The Ely shelf was a slightly deeper environment in western Utah where several hundred meters of fossiliferous cyclical limestone accumulated. More rapid subsidence in the Oquirrh basin caused a prolific brachiopod-bryozoan-sponge biocoenose to accumulate below wave base. The Hermosa biofacies, a more diversified invertebrate fauna reflecting shallow nearshore environments, contrasts with the Callville biofacies, which is a sparse population reflecting a Bahamian-type platform.

By Atokan time, Utah was essentially submerged, except perhaps local areas near the Colorado border adjacent to the Uncompahgre uplift. *Chaetetes, Caninia, Multithecopora,* and *Barbouria* corals constructed extensive biostromes across the shelf environments. Fusulinid coquinas of *Profusulinella* and *Fusulinella* are common in beds within the carbonate cycles. Only in the Oquirrh basin where the water was deeper are the *Chaetetes* biostromes small and discontinuous. In the early Des Moinesian (Cherokee), the Paradox sabka formed adjacent to the Uncompahgre uplift (fig. 8). Salt layers in the Paradox alternate with euxinic black dolomitic shales that contain carbonized wood fragments, conodonts, phosphatic brachiopods, agglutinated Foraminifera, and fish remains (Stone, 1968). Algal limestones formed the stratigraphic reef barrier to the salt basin in the Four Corners region. The Callville platform west of the sabka was a supratidal flat. A few small algal patch reefs of Cherokee age are found in the Hermosa Group; the best exposures are in the gorge near the Goosenecks of the San Juan River (fig. 1, loc. 45).

The most widespread of all the Pennsylvanian open-marine environments existed in the middle Des Moinesian. The Desert Creek and Ismay Limestones of the Hermosa Group (fig. 3) were deposited in an open-marine environment that followed the restricted sabka environment. Equivalent limestones containing abundant *Fusulina* and *Wedekindellina* were deposited over the entire State and even covered the arkosic alluvial fans adjacent to the Uncompahgre. *Chaetetes* biostromes containing abundant brachiopod-coral-bryozoan assemblages flourished and extended from shelves into the deeper waters of the Oquirrh basin. Small syringoporid patch reefs tens of meters long have been observed in the Des Moinesian limestones of the Oquirrh Mountains. After this maximum Des Moinesian transgression, the seas gradually became restricted, and a regional hiatus marks the end of the series.

The earliest Missourian fauna containing *Wedekindellina ultimata* was found by Welsh (Welsh and James, 1961) in the Jordan Limestone of the Oquirrh Mountains associated with a biocoenose of productids, bryozoans, gastropods, and sponges. Missourian fusulinid-bryozoan faunas are found in both the Honaker Trail Formation of the Paradox basin and the Bingham Mine Formation of the Oquirrh basin. Virgilian fusulinids are found in the limestones of the northern Ely shelf, the Oquirrh basin, the southwestern Callville platform, and the Moab trough (fig. 9). Upper Pennsylvanian sandstones of the Oquirrh basin have abundant trace fossils on bedding planes indicative of water depths below wave base. Syringoporid biostromes are common in the Virgilian limestones in all basins. Large *Pseudozaphrentoides* corals and siliceous sponges are indigenous in Missourian limestones of the Oquirrh facies. Productids, bryozoans, corals, fusulinids, and sponges are abundant in most limestones of the Bingham Mine Formation. The Virgilian limestones of the Callville platform and Ely shelf are oolitic, pelletal, free of silt-size quartz, and have birdseye texture that suggests an important algal contribution. The Ely shelf was not receiving detritus from the Antler Highlands nor was the Oquirrh basin. The Honaker Trail Formation of the Hermosa Group has a mixed invertebrate fauna of brachiopods, bryozoans, corals, and mollusks reflecting the shallow estuarine facies southwest of the Uncompahgre uplift.

The youngest Virgilian fauna containing *Dunbarinella* has not been collected by the writers in Utah, although this zone is present in southeastern Nevada in the Bird Spring basin. In the Oquirrh basin, where a thick Virgilian sequence is present, the rocks are mostly submarine deltaic sandstone and thin silty limestone. Pre-Wolfcampian erosion has removed most of the Upper Pennsylvanian rocks on the Emery high (fig. 9) and in southwestern Utah so that the Late Pennsylvanian paleogeographic record is incomplete.

Except for the euxinic facies of the Paradox basin, the Pennsylvanian faunas are generally open marine in carbonate banks, shelves, or basins. Preservation of trace fossils and articulate invertebrates in the Oquirrh sedimentary rocks is a reflection of low-energy deeper water. Crinoid columnals are present in most Pennsylvanian limestones, but branches are common in the sedimentary rocks of the Oquirrh that were deposited in quiet water. Mollusks are present as part of the basin and carbonate bank fauna but are nowhere common. Partly restricted environments like those in the Mississippian containing exclusive molluscan assemblages did not form in the Pennsylvanian.

COLLECTING LOCALITIES

Fifteen localities have been selected as representative of the lithostratigraphy of the Carboniferous deposits in Utah (fig. 11). These localities have fossils that are characteristic of the lithofacies and time-stratigraphic units. The following list gives the locality number used in figure 11, the township and range, and the local name. A section designation is not given because several areas along strike are suitable for collecting.

(1) T. 11 N., R. 2 E.	Left Fork, a tributary of Blacksmith Fork south of Logan, has excellent exposures of the Lodgepole Limestone. Kinderhookian and Osagean invertebrates are easily collected in the talus.
(2) T. 10 N., R. 1 W.	The Dry Lake section along the abandoned road in the Great Blue Limestone has brachiopod and coral faunas of Meramecian and Chesterian age.
(3) T. 1 S., R. 3 W.	Rogers Canyon at the northwest corner of the Oquirrh Mountains has one of the best exposed sections for studying the biostratigraphy of the Morrowan, Atokan, and Des Moinesian series. Silica

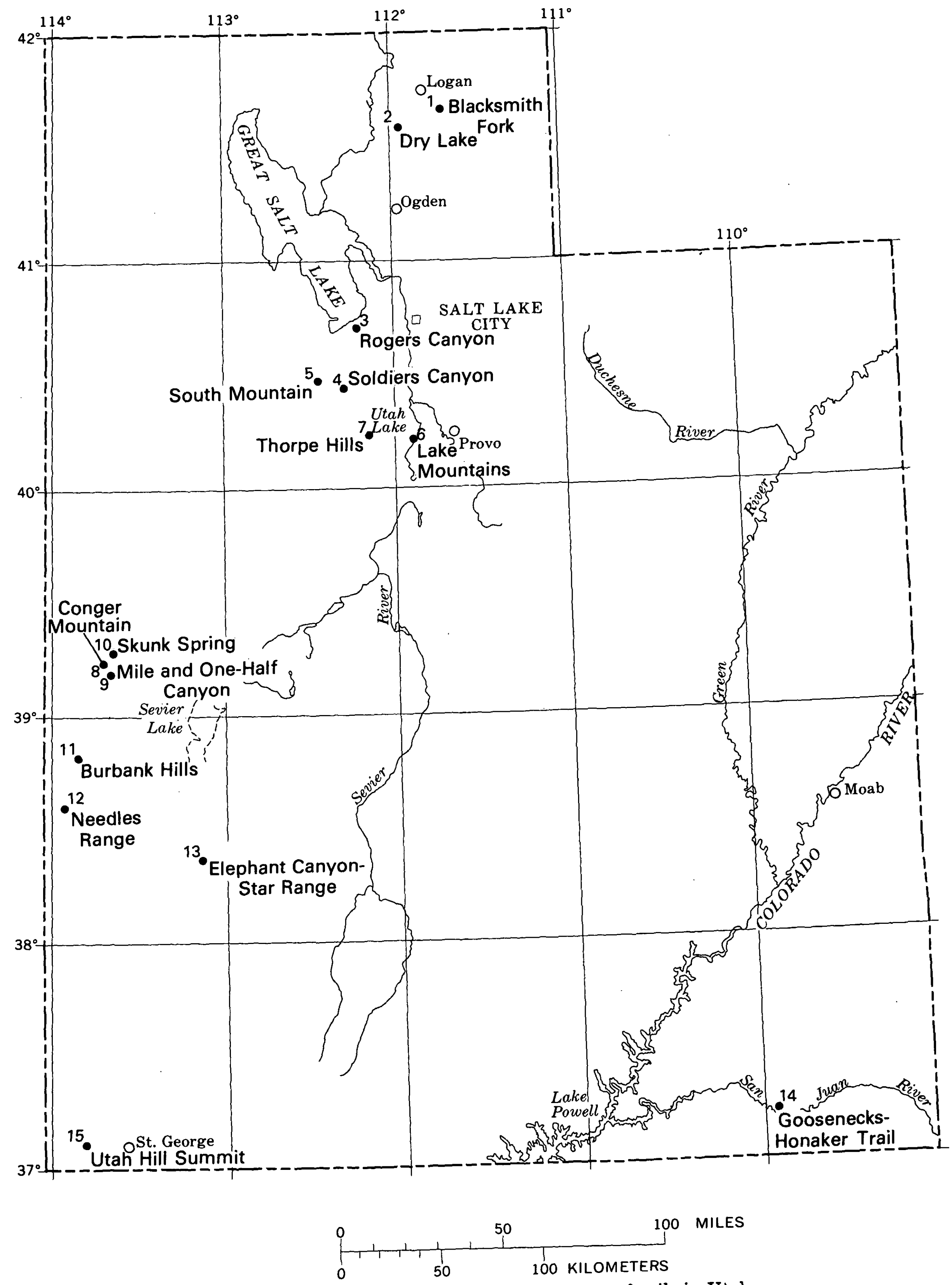


FIGURE 11.—Collecting localities for Carboniferous fossils in Utah.

sponges, crinoids, brachiopods, and corals are part of the biocoenosis.

(4) T. 4 S., R. 4 W. Soldiers Canyon east of Stockton, Utah, has a diversified section of nonmarine and marine deposits containing Chesterian and Morrowan fauna.

(5) T. 4 S., R. 5 W. South Mountain in Tooele valley is the type area of the Missourian and Virgilian Bingham Mine Formation of the Oquirrh Group. This is the most accessible locality for collecting trace fossils, corals, and brachiopods of the Late Pennsylvanian.

(6) T. 7 S., R. 1 W. Clay quarries in the Manning Canyon Formation yield a well-preserved flora of Pennsylvanian (Namurian B) age in the Lake Mountains.

(7) T. 7 S., R. 3 W. The Thorpe Hills, 8 km west of Fairfield, display a diversified brachiopod, coral, and bryozoan fauna of Morrowan age in the West Canyon Limestone of the Oquirrh Group.

(8) T. 18 S., Rs. 16–17 W. Conger Mountain in the Confusion Range has horizontal ledges of Ely Limestone that have an abundant brachiopod, bryozoan, coral, and fusulinid fauna of Morrowan, Atokan, and Des Moinesian age. *Chaetetes* biostromes are thick and continuous.

(9) T. 19 S., R. 16 W. The early Kinderhookian brachiopods and conodonts are most easily collected from the upper part of the Pilot Formation just below the Joana Limestone hogback at Mile and One-Half Canyon in the Confusion Range. Late Kinderhookian and Osagean corals and brachiopods may be collected from the Joana Limestone hogback.

(10) T. 18 S., R. 16 W.
(11) T. 23 S., R. 19 W.
(12) T. 25 S., R. 19 W. The Confusion Range, Burbank Hills, and Needles Range have long strike valleys of the Chainman Formation. These are the best collecting localities for goniatites and other mollusks. The Skunk Springs section (fig. 11, loc. 10) south of Cowboy Pass has abundant Chesterian brachiopods and corals of the Great Blue carbonate bank. These sections represent the transition from the carbonate bank to the flysch basin and provide fossils of different environments.

(13) T. 28 S., R. 12 W. The Redwall (Monte Cristo) Limestone of Kinderhookian and Osagean age is exposed in canyons in the Star Range, 16 km southwest of Milford, Utah. Elephant Canyon has well-exposed ledges for collecting corals, brachiopods, and bryozoans.

(14) T. 41–42 S., R. 18 E. The ledges of the Hermosa Group at the Goosenecks of the San Juan River have excellent collecting for Des Moinesian and Missourian brachiopods and corals. Algal patch reefs in the Paradox Formation are well exposed, and the black shales contain conodonts and fish teeth and bones, as well as phosphatic brachiopods. Honaker Trail is one of the easier accesses to the canyon walls.

(15) T. 42 S., R. 18 W. The Morrowan-Atokan, Des Moinesian, and Virgilian limestones of the Callville Limestone are accessible for collecting just east of the Utah Hill summit on U.S. Highway 91 in the Beaver Dam Mountains. Brachiopods, bryozoans, corals, and fusulinids are common. The Virgin Canyon tributaries off Interstate Highway 15 in Arizona are also favorable localities for collecting these faunas.

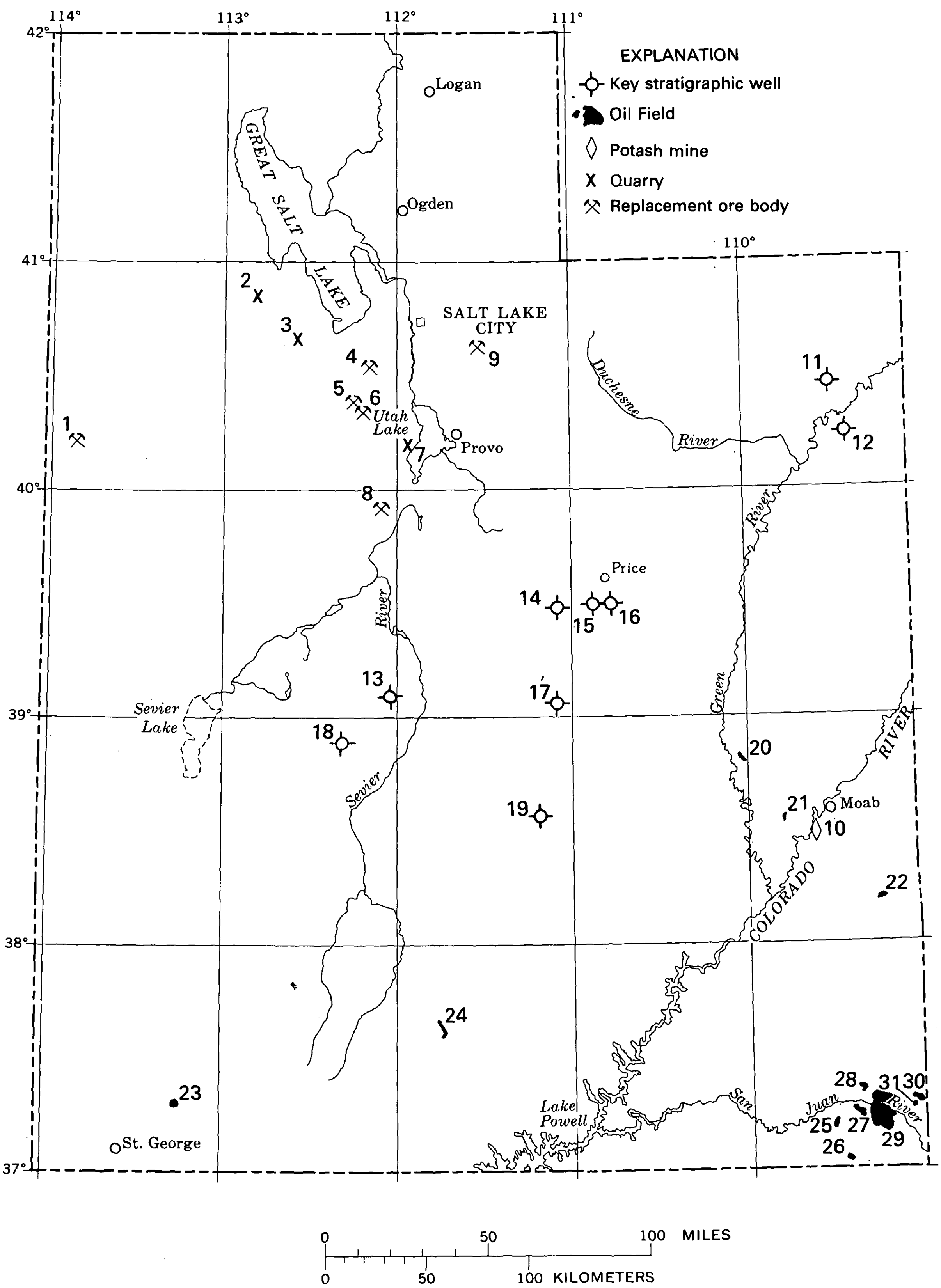
IGNEOUS AND METAMORPHIC ROCKS

Utah does not contain any igneous rocks known to be of Carboniferous age. Carboniferous rocks are regionally metamorphosed near Jurassic gneiss domes in western Utah, and contact metamorphism is common adjacent to Cretaceous and Tertiary stocks.

ECONOMIC PRODUCTS

Coal.—The Carboniferous rocks of Utah do not contain any economic coal deposits. Thin coal beds are present in the deltaic facies of the Manning Canyon and Doughnut Formations in central Utah. Plant fossils and carbonaceous fragments are locally present in the Great Blue Limestone and Chainman, Paradox, and Indian Springs Formations.

Petroleum.—Most of the oil production from the Carboniferous units has been from east of the Wasatch Hinge Line in the Paradox basin. The Anderson Junction field (fig. 12, loc. 23) near St. George produced from the Pennsylvanian Callville Limestone and is the only field near the hinge line.



EXPLANATION
Key stratigraphic well
Oil Field
Potash mine
X Quarry
Replacement ore body
114°
113°
112°
111°
110°
42°
41°
40°
39°
38°
37°
Logan
Ogden
GREAT SALT LAKE
SALT LAKE CITY
Provo
Utah Lake
Duchesne River
Green River
Price
Sevier Lake
Sevier River
COLORADO RIVER
Moab
Lake Powell
San Juan River
St. George
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
0
50
100 MILES
0
50
100 KILOMETERS

Thirty-one Pennsylvanian fields have produced 320 million barrels of oil; Greater Aneth (fig. 12, loc. 29), which has produced 280 million barrels, is the only giant field. Other fields in southeastern Utah that have produced more than 1 million barrels are: Ismay-Flodine Park, 9.5 million; Boundary Butte, 4 million; McElmo Mesa, 2.1 million; Tohonadla, 1.7 million; Bluff, 1.3 million; and Gothic Mesa, 1 million. All these fields are in stratigraphic reefs near the entrance into the Paradox basin. The Ashley Valley field (fig. 12, loc. 11) in the Uinta Basin has minor Pennsylvanian production from the upper part of the Weber Sandstone which is probably Permian; a deep test in the Red Wash field yielded a legitimate show from the middle of the Weber Sandstone.

Mississippian production has been 41 million barrels; Lisbon (fig. 12, loc. 22) has produced 40 million barrels, and Salt Wash (fig. 12, loc. 20) 1.2 million barrels. Upper Valley (fig. 12, loc. 24) and Big Flat (fig. 12, loc. 21) have minor production. Dolomitized crinoidal banks are the main reservoirs, and the oil is probably derived from the Paradox source rocks. The dolomitic shales of the Paradox Formation are the primary source rocks for most of the Carboniferous production.

Potential source rocks for oil in undiscovered Carboniferous reservoirs are shales of the Devonian Pinyon Peak and Pilot Formations, the phosphatic shales of the Mississippian Little Flat Formation, Chainman Group, and Deseret Limestone, and the organic-rich shales of the Great Blue Limestone and Doughnut and Manning Canyon Formations.

Metals.—Mississippian and Pennsylvanian limestones are the host rocks for vein, manto, and skarn deposits in the mining districts of Utah (fig. 12).

Pennsylvanian limestones of the Oquirrh Group are hosts for extensive skarn mineralization. For 50 years, the U.S. and Lark mines in the Bingham district (fig. 12, loc. 4) produced copper, lead, zinc, and silver from replacement deposits in Des Moinesian and lower Missourian limestones. The Anaconda Company is presently developing a large copper skarn ore body in the Missourian Jordan and Commercial Limestones at its Carr Fork underground mine northwest of the Bingham pit.

Some horizons in the Mississippian Humbug Sandstone are hosts for silver, lead, and zinc bedded-ore replacement deposits in the Ontario mine, Park City district (fig. 12, loc. 9). The Chief, Godiva, Iron Blossom, and Plutus veins, the main ore zones in the Tintic district (fig. 12, loc. 8), have bedded replacement deposits in the Fitchville, Gardison, and Deseret Limestones. These manto ores are primarily copper, silver, and gold.

Ochre Mountain Limestone is host of arsenic-gold replacement bodies in the Gold Hill district (fig. 12, loc. 1). Copper-lead-silver replacements are also reported in the Ely Limestone at Gold Hill. Many other districts in Utah have smaller replacement deposits in Carboniferous beds.

Beds of the Mississippian Great Blue Limestone that are rich in organic matter are hosts for disseminated gold deposits of the Carlin type in the Mercur district (fig. 12, loc. 6).

FIGURE 12.—Localities at which economic products have been obtained from Carboniferous rocks in Utah. Also shown are locations of key wildcat wells and oil fields where subsurface stratigraphic data on the Carboniferous are available.

Ore Deposits:
1. Gold Hill (Au, As)
4. Bingham (Cu, Pb, Zn, Ag, Au)
5. Ophir (Pb, Zn Ag)
6. Mercur (Au)
8. Tintic (Ag, Zn)
9. Park City (Pb, Zn, Ag)

Quarries:
2. Lakeside Lime
3. Flux Lime
7. Lake Mountains Clay

Potash mine:
10. Cane Creek

Stratigraphic wells:
11. Ashley Valley
12. Red Wash
13. Scipio Lake
14. Hiawatha
15. North Spring
16. Miller Creek
17. Ferron
18. Meadow
19. Last Chance

Oil fields:
20. Salt Wash
21. Big Flat
22. Lisbon
23. Anderson Junction
24. Upper Valley
25. Tohonadla
26. Boundary Butte
27. Gothic Mesa
28. Bluff
29. Greater Aneth
30. Ismay-Flodine Park
31. McElmo Mesa

Nonmetallic products.—Solution mining at Texas Gulf's Cane Creek mine (fig. 12, loc. 10) yields 200,000 to 300,000 tons of potash yearly from the Paradox Formation near Moab, Utah; the mine has a minimum estimated life of 20 years. The Manning Canyon Formation (fig. 12, loc. 7) is a source of clay for the brick industry. Limestone quarried by Flintkote Corporation from the Great Blue Limestone in the Stansbury Range (fig. 12, loc. 3) is a major source of high-calcium lime which is used for smelter flux and other industrial purposes. Southern Pacific quarries large quantities of the Great Blue Limestone in the Lakeside Mountains (fig. 12, loc. 2) for riprap. Increasing quantities of these limestones are being used as a dust retardant in the coal mines near Price.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States— Arizona

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Prepared in cooperation with the Arizona Bureau of Geology and Mineral Technology

Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Arizona

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—ARIZONA

By H. WESLEY PEIRCE [1]

ABSTRACT

Carboniferous rocks in Arizona are represented by marine and continental deposits ranging in thickness from 0 to 1,100 m. Mississippian marine strata, consisting largely of cliff-forming high-purity limestone, are overlain disconformably by thicker Pennsylvanian strata containing, besides carbonate rocks, varying proportions of siliceous clastic deposits.

The two contrasting outcrop regions are: (1) a northeast or Plateau half, where exposures are limited largely to deep canyons and escarpments; and (2) a southwest or Basin and Range half, where exposures are found only in certain discontinuous range blocks.

Shallow Mississippian seas first transgressed the Arizona region from the northwest and southeast in Kinderhookian time. Strata as young as Chesterian are preserved only locally, an unknown thickness of Upper Mississippian rocks having been removed before the onset of Pennsylvanian sedimentation.

During Pennsylvanian time, northern Arizona was flanked by marine basins to the west, northeast, and southeast, the central part receiving a relatively thin zone of clastic deposits. Precambrian granitic source rocks were exposed on the Defiance positive area in east-central Arizona and were partly onlapped by Pennsylvanian deposits, at least during Missourian and Virgilian time. Faulting during the Pennsylvanian probably gave magnified expression to the southwest margin of this feature.

Carboniferous rocks, chiefly limestone, contributed to products valued at more than $50 million during 1977. Principal commodities were portland cement and quicklime. The welfare of almost every Arizona resident is enhanced by the State's Carboniferous rocks.

INTRODUCTION

DISTRIBUTION AND GENERAL CHARACTERISTICS

Known Carboniferous rocks in Arizona consist only of sedimentary materials. In the Plateau region of northern Arizona, most of these rocks, except for two linear belts of exposures represented by (1) the Grand Canyon and (2) the southern cliffs margin of the Plateau, are buried beneath either Paleozoic or Paleozoic and Mesozoic strata. In the southern half of the State, the Basin and Range part, outcrops are limited to relatively short linear strips within certain mountain or range blocks, especially those in southeastern Arizona. Very little is known of the subsurface nature and distribution of these rocks within the valley, or basin, blocks. Overall, only a very small percentage of the Arizona surface environment is covered with rocks of Carboniferous age. This small percentage is, in turn, unequally distributed (fig. 1).

The most continuous outcrop belt of Carboniferous rocks in Arizona is in the Grand Canyon, a product of late Cenozoic erosion by the Colorado River. Here, because the strata are nearly flat-lying and relatively undeformed, each layer imparts its own set of strengths and weaknesses to form the topographic profile. In particular, the Mississippian strata form one of the more prominent cliffs throughout Grand Canyon, whereas the overlying Pennsylvanian strata support both ledges and slopes. Along the southern edge of the Plateau province, various erosional events have exposed Carboniferous strata. Because of a slight northeast dip, probably imposed initially in Mesozoic time, older strata were truncated southward, and Carboniferous and other rocks were exposed in Mesozoic time. Although several subsequent tectonic-erosional episodes took place, the principal deep canyons that now are being cut into the Plateau edge were initiated during the late Cenozoic (Peirce and others, 1976), probably as a response to drainage integration in contrast to concurrent so-called Plateau uplift, as hypothesized by McKee and McKee (1972).

South of the Plateau, in the Basin and Range region, Carboniferous rocks were subjected to severe tectonic disturbances during Mesozoic and Cenozoic times. Several episodes of plutonism are known, and, locally, Carboniferous strata have been metamorphosed and mineralized. Flat-lying strata are rare, steep dips being the general rule. The present

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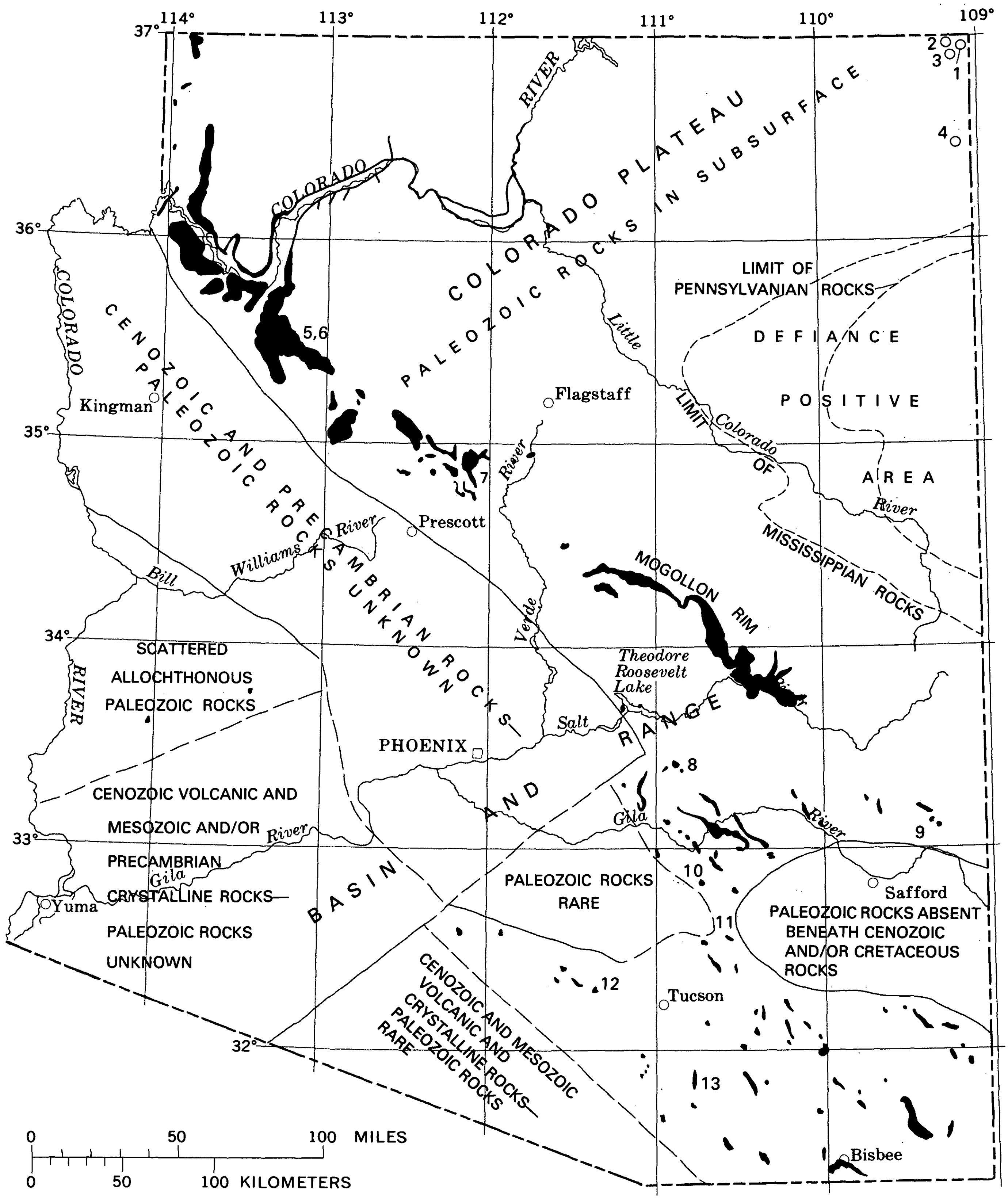

FIGURE 1.—Carboniferous outcrops, distribution of Paleozoic and other rocks, and primary sources of Carboniferous products (numbered and listed in table 1). Areas of broad geologic similarity shown as solid lines; long dashes where inferred. Zero isopach for Mississippian and for Pennsylvanian rocks in northeast part of State shown by short dashed line.

physiographic pattern of disconnected ranges and basins is largely the result of late Cenozoic extensional tectonics. Where present in this province, as in the Plateau province, the Mississippian strata form prominent cliffs, and the Pennsylvanian strata form ledges and slopes.

HISTORICAL REVIEW AND USE OF "CARBONIFEROUS"

The term "Carboniferous" is no longer widely used in Arizona because: (1) the Mississippian and Pennsylvanian parts generally are separable and mappable; (2) their respective histories are notably different; and (3) the top of the Pennsylvanian is, in most places (more so in the subsurface), imprecisely defined.

Arizona Carboniferous rocks initially were included in studies undertaken in two widely separated regions: (1) in northern Arizona at Grand Canyon because of conspicuous exposures, and (2) in southern Arizona mining camps where many of these rocks are associated with ore development.

In 1875, Gilbert gave the name "Red Wall Limestone" to more than 610 m (2,000 ft) of Grand Canyon strata. Apparently, this interval included rocks ranging in age from Devonian to, and perhaps including, Permian. Subsequent refinement by Darton (1910) and Noble (1922) led to the presently prevailing restriction of this name (now Redwall) to strata of Mississippian age. Overlying Pennsylvanian-Permian(?) strata were, in turn, designated "Supai Formation."

In extreme southern Arizona, in the Bisbee quadrangle, Ransome (1904) defined a Carboniferous section as consisting of about 1,128 m (3,700 ft) of combined Mississippian "Escabrosa limestone" (61 m) and Pennsylvanian "Naco limestone" (900 m or more). Subsequently, upon recognition of Permian strata, the "Naco limestone" of Ransome has been altered and restricted.

In 1905, Lindgren, in discussing the mineralized Clifton quadrangle in east-central Arizona, recognized Carboniferous strata. Although outcrops are limited, he defined two outcrop sequences: (1) one to the south exposing about 52 m of Mississippian strata (beneath Mesozoic rocks) to which, apparently, he assigned the name "Modoc limestone" although, in his discussion, he uses "formation" and "limestone" interchangeably, and (2) one to the north exposing about 152 m (500 ft) of strata underlying Tertiary rocks. To this sequence he applied the name "Tule Spring limestone," recognizing that the lower 61 m is Mississippian in age, and the upper 91 m, Pennsylvanian. Whereas this terminology has been applied only locally, the Grand Canyon and southern Arizona terminologies, in contrast, have been applied over much of the Plateau and Basin and Range provinces.

Ransome (1916) attempted to tie together Arizona Paleozoic stratigraphy by correlating sections from Grand Canyon on the north to Bisbee in extreme southern Arizona. He defined a Carboniferous "Tornado limestone" in the Globe-Ray region of central Arizona. He recognized a Mississippian lower massive part and a Pennsylvanian upper part but did not draw a contact in between. He further noted an analogy between the "Tornado limestone" and the Escabrosa-Naco limestones of the Bisbee region. Subsequently, the use of "Tornado limestone" has been replaced by Ransome's original Bisbee area terminology.

Darton (1925) gave one of the more comprehensive coverages of the general geology of Arizona. It includes a summary of the "Carboniferous System" as well as an extensive documentation of the distribution of outcrops of Mississippian and Pennsylvanian strata. Much of the data presented was an outgrowth of fieldwork done cooperatively by the U.S. Geological Survey and the Arizona Bureau of Mines in connection with the first (Darton, 1924) State geologic map of Arizona.

While a professor at the University of Arizona, Stoyanow (1926; 1936), assisted by students seeking advanced degrees (as well as those who worked during summer field seasons with the Arizona Bureau of Mines' geologic program), undertook to establish regional relationships of Paleozoic rocks in Arizona. Although much of the early paleontological work should be credited to G. H. Girty of the U.S. Geological Survey, Stoyanow, also a paleontologist, examined and interpreted the significance of the paleontological highlights of the Carboniferous strata of the State.

McKee (1951) made further paleogeographic interpretations over the entire State, including those relating to Mississippian and Pennsylvanian rocks. Havenor (1958) gave special emphasis to a review of the Pennsylvanian sedimentation framework in Arizona. Most recently, Purves (1976) has initiated a study and review of the Mississippian System of Arizona. In addition to a comprehensive review of past studies, he has presented a list of 33 Basin and Range mountain blocks and other localities in which work on Mississippian strata has been done. Purves is attempting to refine statewide time-facies boundaries by rigorous conodont zonation and petrographic studies.

Many important later contributions have been made by studies of regions of lesser size than the State. Most such studies focus on either the Mississippian or Pennsylvanian systems and not on a "Carboniferous System." Some of these include: Jackson (1951); McNair (1951); Huddle and Dobrovolny (1952); Hughes (1952); Sabins (1957); Thomas (1959); Fetzner (1960); Kottlowski (1960); Armstrong (1962); Hammer and Webster (1962); Lokke (1962); Kottlowski and Havenor (1962); Yochelson (1962); Sabins and Ross (1963); Winters (1963); Brew (1965); Finnell (1966a, 1966b); McKee and Gutschick (1969); Lessentine (1969); Peirce and others (1970); Norby (1971); Blazey (1971); Ross (1973); Racey (1974); Smith (1974); Conyers (1975); McKee (1975a, 1975b); Kent (1975); Peirce and others (1977); and many others.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Arizona Bureau of Geology and Mineral Technology.

GEOLOGIC SETTING

CONTACTS

Mississippian rocks almost everywhere overlie strata of Devonian age. In outcrop, this contact most commonly is described as showing conformity. In very few places, an erosional surface is evident and relatively thin discontinuous conglomerates are present. In turn, Pennsylvanian rocks prevailingly overlie Mississippian strata unconformably except in the subsurface near the Defiance positive area in east-central Arizona, where they overlap both Devonian strata and Precambrian granitic rocks (Peirce, 1976, p. 40).

In general, the cliffs formed by Mississippian rocks contrast markedly with the ledge-slope terrain supported by a relatively thin sequence (usually less than 200 m) of Devonian strata. In detail, the appearance of lithologic gradation often hinders identifying a precise contact. However, faunal evidence suggests that a hiatus involving at least the lowermost Mississippian generally exists. Perhaps the contact between Devonian and Mississippian rocks is most conveniently described as being a paraconformity.

The Carboniferous of Arizona generally may be divided into both Mississippian and Pennsylvanian formational components because an identifiable internal contact is present. That the contact may at times be subtle is suggested by the earlier tendency to lump these components together under one formational entity, such as Tornado limestone and Tule Spring limestone. Nevertheless, on more recent geologic maps, including the State geologic map (Wilson and others, 1969) at a scale of 1:500,000, Mississippian and Pennsylvanian strata are everywhere depicted as separate map units. Again, in outcrop, conformity is the prevailing appearance even though there generally is a hiatus involving late Mississippian and early Pennsylvanian time. In a regional sense, it seems essential that some minor angular discordance exists, at least locally. In central Arizona, a well-recognized thinning of Mississippian strata (fig. 2) is associated, at least in part, with an upper subaerially generated Paleozoic erosion surface. This surface is readily identified in most of the petroleum test holes drilled in northeastern Arizona (Peirce and Scurlock, 1972). Here, Pennsylvanian strata appear to truncate Mississippian strata along the southern part of the Defiance positive area (fig. 1). In outcrop, the Mississippian-aged cliff-forming limestones almost everywhere contrast with the ledge and slope topography of overlying Pennsylvanian strata.

Defining the top of Carboniferous, or Pennsylvanian, rocks in Arizona is a classic problem. The major stumbling block has been a paucity of critical chronological data preserved in either the highest Pennsylvanian or the lowest Permian rocks. Only one generally successful local effort has been made to define an acceptable and mappable Pennsylvanian-Permian systemic boundary, and this was by Winters (1963) in east-central Arizona. Even this boundary, though convenient, is somewhat arbitrary because of the apparent absence of diagnostic Early Permian fossil forms. More recently, McKee (1975a) redefined certain Pennsylvanian-Permian strata of a part of Grand Canyon. In so doing, he suggested that a Pennsylvanian-Permian boundary can be defined by a conglomerate that constitutes the base of Permian rocks in some sections. Elsewhere, thin conglomerate zones exist in many sections along the southern edge of the Plateau province (Peirce and others, 1977). Some are closely associated with a rather prolific record of plant fossils that appear to occupy a position close to the Pennsylvanian-Permian boundary (Blazey, 1971). Additional investigative effort may shed light on this time boundary. In

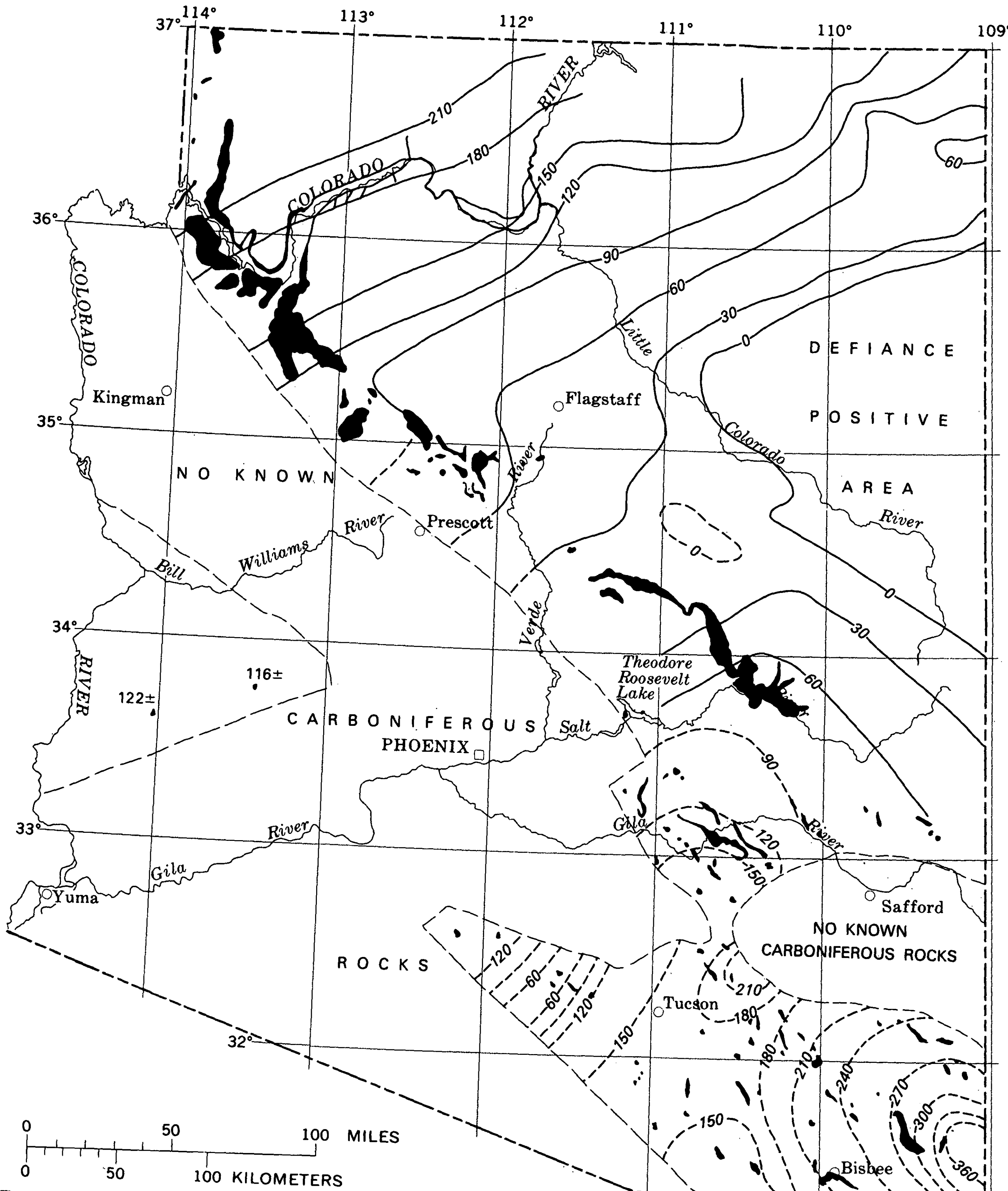

FIGURE 2.—Thickness (in meters) of Mississippian rocks as inferred from present distribution. Dashed where uncertain.

southern Arizona, the Pennsylvanian-Permian boundary is in a sequence of marine limestones that in places can be bracketed but not defined precisely. South of the Plateau is a broad region in which Pennsylvanian strata are overlain by Tertiary rocks. Overall, the upper contact between Carboniferous and Permian rocks remains vague and generally within conformable stratal sequences.

STRUCTURAL FRAMEWORK

During the Carboniferous, Arizona probably occupied a shelf position relative to two geosynclines, the Cordilleran to the west and the Sonoran to the south. Strata of both systems thin to extinction in part of east-central Arizona. A stratal comparison clearly shows that this shelf position was much more stable during the Mississippian Period than during the Pennsylvanian.

Mississippian strata consist of relatively quartz free limestones that accumulated in a shallow open-marine environment. Overall thickness variations appear to be systematic (fig. 2). McKee (1969) noted that exposures in the Grand Canyon region reflect three transgression-regression episodes ranging in age from late Kinderhookian to Chesterian, the latter record being incomplete because of later exposure and erosion. The slight earth movements that affected the deposition and subsequent regional erosion of Mississippian strata might properly be classed as epeirogenic. The thinning and absence of these strata in parts of east-central Arizona seems to be related to both nondeposition and erosion prior to the deposition of Pennsylvanian strata.

In general, Pennsylvanian rocks are thicker, more varied, more siliceous, thinner bedded, and more cyclic than Mississippian rocks. The influence of tectonism, both regional and local, on the Pennsylvanian System in Arizona contrasts markedly with the Mississippian. Thinning over a short distance onto Precambrian granitic rocks in east-central Arizona may signify faulting during the early Pennsylvanian (fig. 3). This region seems to be a southwest edge of the so-called ancestral Rockies (Eardley, 1962) that evolved during Pennsylvanian time. The principal post-Carboniferous structural events took place in Mesozoic and Cenozoic times. In Mesozoic time, a northwest-trending uplift centered southwest of the present Plateau edge, imparted a shallow northeast dip to the Plateau Paleozoic rocks. Pre-Upper Cretaceous erosion (probably Triassic and Jurassic) beveled Paleozoic strata and Precambrian rocks southward, thus exposing Carboniferous strata for the first time since original burial by Permian deposits. South of the structural high point, structural lowness may have prevailed as indicated by the preservation of much of the Paleozoic sequence in certain localities. However, local zones are present in this (once structurally low) terrain where Cretaceous deposits rest depositionally upon Precambrian crystalline rocks (Empire mountains), thus demonstrating the existence of at least local pre-Laramide deformation and erosion.

In a Late Cretaceous-early Cenozoic time, the entire State was affected by the so-called Laramide orogeny, at which time the present Rocky Mountains evolved. Plateau strata locally were folded and faulted along north- to northwest-trending fold axes of considerable length. In the south, igneous activity was widespread, ore deposits associated with much fracturing were emplaced, and, according to some, thrusting was widespread (Drewes, 1977). Carboniferous rocks were important hosts for base-metal deposits in the south, whereas, to the north, some petroleum accumulations in Carboniferous rocks were controlled by structures derived at this time.

Post-Laramide Cenozoic structural history is still being unraveled. Allochthonous blocks of Carboniferous rocks are well known in the Basin and Range province. To the extent that these blocks are associated with allochthonous Cenozoic rocks, the latest causal event must be Cenozoic in age. Preliminary evidence suggests: (1) a dislocative event during the lower Miocene between 13 and 20 m.y. ago followed by (2) the classic Basin and Range rifting event during late Miocene-Pliocene less than about 15 m.y. ago (Peirce, 1976). This latter event defined, for the most part, the structural setting that controls the broader characteristics of the contemporary landscape south of the Plateau.

GENERAL STRATIGRAPHY AND LITHOLOGY

MISSISSIPPIAN

Mississippian strata range in thickness from 0 to 380 m (1,250 ft). The greater thicknesses are in the northwest and southeast corners of Arizona; thinning to wedge-out beneath Pennsylvanian rocks (fig. 2) is found in east-central Arizona on the Defiance positive area of McKee (1951; Peirce and others, 1977, fig. 6).

In the Plateau section of Arizona, the Mississippian strata are known by the name Redwall Limestone. A formal fourfold division into members is in general use. From the base upward, these units are

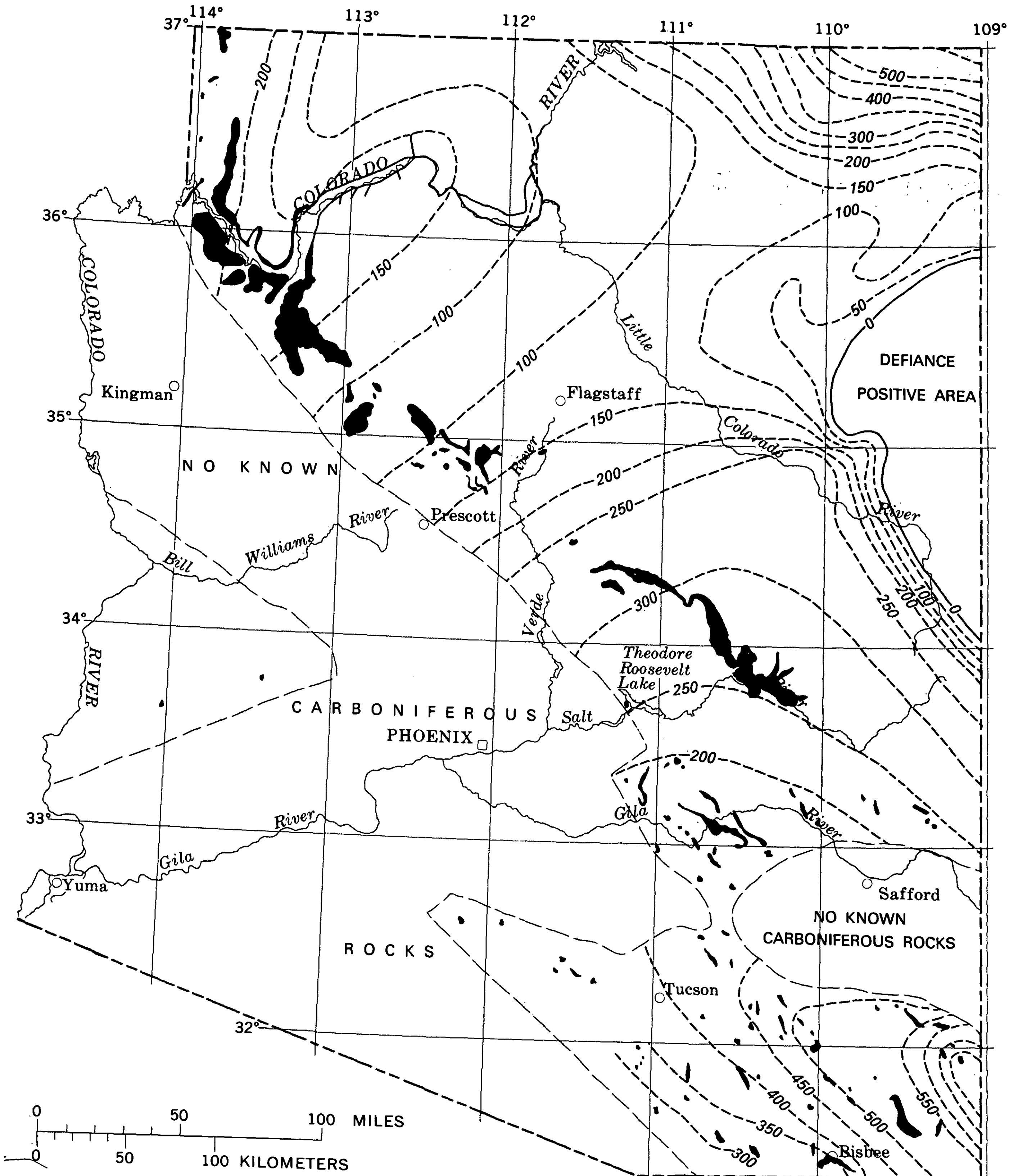

FIGURE 3.—Thickness (in meters) of Pennsylvanian rocks as inferred from present distribution.

known as: (1) Whitmore Wash, (2) Thunder Springs, (3) Mooney Falls, and (4) Horseshoe Mesa. Generally, these members are recognizable in subsurface drill tests as well as in outcrop.

The Redwall Limestone consists almost wholly of carbonate varieties and is generally free from insoluble residues, except chert. Granular carbonate rocks, especially crinoidal beds, predominate. Aphanitic limestone, dolomite, and chert are found in certain beds.

According to McKee (1969), the first transgression began in western Grand Canyon in late Kinderhookian time; the second, in Osagean time, was followed by regression in Meramecian time. Evidence of a third transgression is indicated by isolated remnants of Chesterian-age rocks preserved locally at the top of the Redwall Limestone. An unknown thickness of this formation was removed before initial deposition of the overlying Supai Group in Grand Canyon, the Naco Formation in central Arizona, and the Black Prince-Horquilla limestones of southern Arizona.

The Mississippian rocks of southern Arizona have not been studied regionally in any detail by any one worker. Effort has been concentrated largely in southeastern Arizona where the rocks of this system are thickest (fig. 2).

Thomas (1959) gave a brief summary of southeastern Mississippian highlights as then understood. Armstrong (1962) further defined lithologic and paleontological character found in extreme southeastern Arizona. He raised "Escabrosa" to group status and named two new formations: the Keating Formation below and the Hachita Formation above. Thus far, this terminology has not generally been used in southern Arizona because it does not appear to have regional application, especially to the west. Norby (1971) examined the conodont characteristics of eight widely distributed sections and provided additional lithologic detail.

All workers emphasize the granular crinoidal fragments of the Escabrosa Limestone. Some lithographic limestone, dolomite, chert, and oolitic units are also present. More acid-insoluble material seems to be near the base of the unit than is generally reported in the Grand Canyon country. This may be from the reworking of siliceous components of underlying Devonian strata.

The youngest representative of Mississippian strata in Arizona is found in extreme southeastern Arizona in the Chiricahua Mountains. It is known as the Paradise Formation and was first recognized by Stoyanow (1926, 1936). He described a 40-m-thick sequence assigned to the Paradise Formation as being both distinctive from, and above, the Escabrosa Limestone. This unit is thinner bedded and contains more siliceous components than does the Escabrosa Limestone, and is distinctly yellow where viewed from a distance. He considered it to be late Meramecian and early Chesterian in age. Norby (1971), on the basis of conodonts, considered the Paradise Formation to be late early Chesterian and the upper beds of the Escabrosa Limestone to be early late Meramecian in age. Armstrong (1962) noted that during Chesterian time the crustal instability, so well reflected in Pennsylvanian strata, was initiated. He referred to a paleogeographic feature in northern New Mexico and east-central Arizona as the Penasco dome. This same feature had earlier been referred to as the Defiance positive area (McKee, 1951), a name that seems firmly entrenched in contemporary literature dealing with the Arizona geologic framework. The evidence contained in the Paradise Formation, as pointed out by Armstrong, probably reflects instability of this more positive area and southerly withdrawal of Mississippian seas.

Norby (1971) assigned a late Kinderhookian age for the basal Escabrosa Limestone. Earlier, G. H. Girty (in Ransome, 1904) also had suggested that the lower Escabrosa Limestone is Kinderhookian in age.

The following comments about the depositional environments reflected in Mississippian strata of southern Arizona are taken from Norby (1971, p. 24), unless otherwise noted.

Mississippian strata, for the most part, were deposited in a warm shallow sea of normal salinity. Micritic limestone may represent isolation from currents, whereas oolitic and pelletoid limestones suggest intermittent currents related to wave and (or) tidal action. The crinoidal calcarenites that are very characteristic of the upper half of the formation suggest widespread stability of conditions suitable for the growth of crinoids. Also, this debris is well sorted, which suggests significant current activity, perhaps near or at wave base. The Paradise Formation, containing a variety of shale, sandstone, and carbonate, indicates fluctuations of sea level and near-shore environments. Armstrong (1962) noted that plant fossils in some of the shales are indicative of the proximity to land.

Norby suggested a rapid northwestward transgression in Kinderhookian time as well as a rapid regression in the late Meramecian. Seas reappeared in early Chesterian time, as indicated in the Paradise Formation. Apparently a second transgression during Osagean time, analogous to that recognized by McKee in the Grand Canyon region, has not been defined.

The relationship between Mississippian strata of the northern and the southern regions is not clear. In outcrop, near the geographical center of Arizona, a natural geographic basis is found for dividing these two domains. Stoyanow (1936), noting the thinning of Mississippian strata toward this region (fig. 2), as well as facies changes and local onlap of older Paleozoic strata, referred to the paleogeographic feature as Mazatzal land. Its original extent to the southwest is not known because of the general removal of Paleozoic rocks in that direction. However, on the basis of paleontological evidence, faunal intermingling may have been restricted; this restriction would indicate that a southwest-extending barrier may have existed. Figure 2 suggests that the thinning in central Arizona is related to activity on the Defiance positive area.

PENNSYLVANIAN

Pennsylvanian strata are thicker and more variable than Mississippian strata (fig. 3). They range in thickness from 0 to more than 725 m (2,370 ft) and include both continental and basinal deposits. McKee (1975b) gave an excellent discussion of statewide Pennsylvanian rocks. Ross (1973) provided a thorough discussion of Pennsylvanian depositional environments that prevailed in southern Arizona.

Although the early workers lumped Mississippian and Pennsylvanian strata together, it is now generally recognized that these two systems are separated by a hiatus that includes the extremes of each period. Evidence of exposure and solutioning of Mississippian carbonate rocks to produce a karstic surface is widespread. An insoluble chert rubble is present in many outcrop localities, and much of it has been reworked into basal Pennsylvanian beds. Red coloration of this zone is characteristic where not swept clean by erosion. The contrast between red beds at the base of Pennsylvanian strata and the clean light-colored carbonate rocks of the Mississippian strata makes an important marker horizon in wells drilled in northeastern Arizona.

The Pennsylvanian system contains more marine carbonate rocks in each corner of the State than in the center, except in the southwest, where there is no known record. In central-northern Arizona (much of the Plateau), the amount of carbonate rocks decreases and the amount of red-bed clastic rocks deficient in fossils increases; therefore, the upper limit of Pennsylvanian strata, for the most part, has not been effectively defined, especially in the subsurface. The clastic red beds of much of the Plateau region are lumped into the Supai Formation, which consists largely of Permian strata in east-central Arizona near the Plateau margin.

McKee (1975a) suggested a Pennsylvanian-Permian boundary in a part of the Grand Canyon, and Winters (1963) suggested a boundary in east-central Arizona. McKee recommended that the Supai be raised to group status and that new formations be recognized. Perhaps, as a result of further study, a method for establishing regionally correlatable units within the Supai Formation will be found. However, stratigraphic keys of regional extent will not be easily defined or accepted.

In each of the corners where Pennsylvanian marine carbonate rocks exist, a Permian boundary usually is described as occurring somewhere within a conformable stratal sequence. In central Arizona, along the Plateau edge escarpment, channel-fill and related deposits contain plant fossils that have yet to be studied. Regional stratigraphic considerations suggest that these plants are near the Pennsylvanian-Permian boundary, but on which side is not yet certain. Blazey (1971) did the initial work on plant forms from one fossiliferous locality.

The approximate thickness extremes of Pennsylvanian strata in the three corners, as recorded by McKee (1975b), are: northwest, 335 m (1,100 ft); northeast, 550 m (1,800 ft); and southeast, 730 m (2,400 ft). Pennsylvanian strata are absent beneath Permian rocks on the Defiance positive area, and, in the center of the State where clastic rocks prevail, they are 90–180 m (300–600 ft) thick (fig. 3).

The northwestern area probably contains the eastern marine edge of the Cordilleran geosyncline. The northeastern region is the edge of the Pennsylvanian Paradox basin of southeastern Utah and southwestern Colorado. The marine rocks of the southeastern region, the thickest of Pennsylvanian age in Arizona, are considered a part of the Sonoran geosyncline. This relatively thick sequence in southeastern Arizona and New Mexico originally accumulated in what frequently is called the Pedregosa basin (Kottlowski, 1960). Kottlowski provided an excellent summary of the Pennsylvanian stratigraphy of the southern Arizona region.

The recognition of three different marine sections, separated by a red-bed clastic section in the northern region, has led to four different sets of nomenclature. Neither the eastern Nevada nor the southern Utah sections is reviewed here.

NORTHERN REGION

The marine section of eastern Nevada thins into northwestern Arizona, where red beds assigned to the Supai Formation dominate. McKee (1969, p. 88) suggested that the top of the Pennsylvanian System is marked by a distinctive limestone-siltstone pebble conglomerate "throughout Grand Canyon." Using this reference surface, McKee (1975a) elevated the Supai to group status and defined three new formations. He also raised the Permian Esplanade Sandstone from member to formational rank. The newly delineated Pennsylvanian units, from oldest to youngest, are: (1) Watahomigi Formation, (2) Manakacha Formation, and (3) Wescogame Formation. The last is overlain by the conglomerate thought to be Wolfcampian in age. The type sections for these units are in Havasu Canyon.

The Watahomigi Formation unconformably overlies the Mississippian Redwall Limestone, is 65 m thick, contains largely limestone and mudstone in about equal parts, and is thought by McKee to be Morrowan and Atokan in age. The overlying Manakacha Formation is 77 m thick, consists of sandstone and minor mudstone, and is either Atokan or Desmoinesian in age. The Wescogame Formation is 61 m thick, consists largely of massive crossbedded sandstone or sandy limestone, is probably of Virgilian age, and contains a basal conglomerate. Overlying the Wescogame Formation is the basal conglomerate of the Permian Esplande Sandstone, a prominent cliff-forming sandstone in Grand Canyon. Noble (1922) had divided the Supai Formation into three members lettered A, B, C from top to bottom. His measured section at Bass Canyon includes a conglomerate high in member B that probably is the conglomerate marker of McKee.

East of Grand Canyon, Carboniferous rocks do not again crop out in Arizona. In the subsurface, these rocks are the prime objective in petroleum exploration in the Four Corners region. The nomenclature for Pennsylvanian strata in this region seems to be in a state of flux. In Utah (see chapter on Utah), Pennsylvanian rocks are thickest in the subsurface. They change facies and thin to the southwest in northeastern Arizona. The change southwestward is to undifferentiated red beds of the Supai Formation.

CENTRAL REGION

Pennsylvanian rocks are exposed in canyons and cliffs associated with the southern edge of the Plateau—the Mogollon Rim escarpment (fig. 1). The exposure of Mississippian-Pennsylvanian(?) rocks to the south is in the walls of Oak Creek Canyon about 161 km (100 mi) from Grand Canyon. Stratigraphic differences between the Grand Canyon to the north and the main Mogollon Rim segment farther south and east emphasize the stratigraphic importance of the Oak Creek Canyon locality. No recognized diagnostic fossils are above Mississippian exposures; correlations must be based on lithology. Whereas McKee's Supai Group in Grand Canyon (Havasu Canyon) is about 335 m (1,100 ft) thick, 204 m of which is Pennsylvanian, the Supai Formation in Oak Creek Canyon is nearly 780 m (1,600 ft) thick (Twenter and Metzger, 1963). It is readily divisible into three principal lithologic units: upper, middle, and lower, or A, B, and C, respectively. Although fossils are absent, the Supai Formation generally has been considered Pennsylvanian-Permian in age. Limestone-siltstone pebble conglomerates in the middle (B) slope-forming unit (84 m thick) might possibly bear a significant relationship to the Pennsylvanian conglomerates of McKee at Grand Canyon. If so, then the lower 122–183 m (400-600 ft) of the Supai Formation could be Pennsylvanian in age; thickness of the Pennsylvanian rocks in the Supai Group at Grand Canyon is similar. This lowest part contrasts with the remaining Supai in that it forms cliffs and, in addition, certain beds contain vertically and horizontally oriented tubular chert phenomena.

Forty-eight km (30 mi) farther south is Fossil Creek Canyon, an important stratigraphic reference point at the west-northwest end of the main Mogollon Rim of central Arizona. Along or beneath the rim, Carboniferous rocks are variably exposed for 160 km (100 mi). At Fossil Creek, the stratigraphic section above the Mississippian limestone contains some fossiliferous limestone in the lowest 76 m (250 ft); the fossils are probably of Demoinesian age (Huddle and Dobrovolny, 1945). These rocks, combined with the Supai Formation, total about 550 m (1,800 ft) in thickness. Again, there is no fully accepted Pennsylvanian-Permian boundary. However, beneath a conglomerate about 244 m (800 ft) above the top of the Mississippian, is a coaly zone from which spore-pollen has been recovered and examined. Several years ago I submitted a sample from this zone to Norman O. Frederiksen, then of the Socony Mobil Oil Co. He reported that the materials were probably lower Wolfcampian (Hueco) but that they could be as old as Upper Cisco (Virgilian). This coaly zone (this and the overlying conglomerate were first reported by Ransome in 1916; Noble, in 1922, while discussing a Supai conglomerate in Grand Canyon, referred to Ransome's Supai conglomerate at Fossil Creek) and its stratigraphic setting, have been discussed by Peirce and others (1977). The zone is within a stratigraphic interval about 152 m (500 ft) thick that contains interbedded limestone-siltstone pebble conglomerates, the lowest of which is about 76 m (250 ft) below the coaly bed. This lowest conglomerate, in turn, is about 135 m (450 ft) above the limestone that contains Desmoinesian fossils. It seems possible, therefore, that the 213-m (700-ft) interval between the Desmoinesian limestone and the possible lower Wolfcampian coaly zone includes all Missourian and either all or part of Virgilian time.

Lithologic correlations can probably be made between Oak Creek Canyon and Fossil Creek Canyon even though the latter section is thicker. Except for the fossiliferous limestone near the base of the Fossil Creek section, the sections appear analogous although not identical. A threefold division seems reasonable at both localities, the conglomerates falling into the middle or B unit. Most likely the B unit contains the Pennsylvanian-Permian boundary in both locations. Brew (1965) suggested that the Pennsylvanian strata of central Arizona are no older than Desmoinesian. If Brew is correct in his age assignment, then the Watahomigi (Morrowan) of McKee in Grand Canyon is not represented in central Arizona sections. Because of the Atokan and (or) Desmoinesian age range assigned to the Manakacha, it may or may not be represented in central Arizona. However, the Wescogame of "probable" Virgilian age, which has a basal conglomerate, could have representation in the middle, or B, unit. A time and lithostratigraphic correlation with the Permian Esplanade Sandstone, and its Wolfcampian basal conglomerate, theoretically could fall within the A unit (largely sandstone) of Oak Creek Canyon.

McKee's data indicate that Missourian time in the Grand Canyon region is represented by an hiatus. Brew suggested that the conglomerate-bearing unit (middle, or B) in central Arizona is Missourian in age. This is a projection from 80 km to the east and is not supported by actual paleontological data in or near Fossil Creek. I suggest that the data from the coaly bed limit the amount of section, if any, that should be assigned to the Missourian. The conglomerates represent erosional hiatuses in which unknown quantities of section were removed. At present, it seems more appropriate to consider them a part of a section that is probably Virgilian-Wolfcampian rather than Missourian. On this basis, the Missourian hiatus postulated by McKee in Grand Canyon may extend, at least in part, to the Oak Creek and Fossil Creek localities.

If correlation with the Grand Canyon section of McKee is attempted, the following possibilities should be considered: Watahomigi—not represented; Manakacha—lower, or C; Wescogame—middle, or B; Esplanade Sandstone—upper, or A and, possibly, part of middle, or B.

The Pennsylvanian strata of central Arizona, including the subsurface, are referred to the Pennsylvanian-Permian Supai Formation, the Naco Formation of Pennsylvanian age, or both. The Naco Formation was defined in southern Arizona and originally included Permian rocks. Whereas the southern Arizona section now includes a Naco Group, a Naco Formation is still used in central Arizona in both the northern part of the Basin and Range province and the southern Plateau province.

Eastward and southward from Fossil Creek, the thickness of sections containing non-red Pennsylvanian age strata increases. The thickness of the non-red section, dominated by limestone and gray shale, increases from a few meters at Fossil Creek to more than 370 m in the Carrizo Creek-Salt River Canyon area 130 km distant. Although this change usually is described as gradual throughout this distance, a significant change occurs across a zone near Canyon Creek, only 48 km from the Carrizo Creek-Salt River region (Peirce and others, 1977). The contrast in sections at the extremes, that is, Oak Creek to the north and the Salt River Canyon region to the south, traditionally has been explained as a northwesterly gradation to Supai Formation as if there were meter-by-meter replacement. Using the top of the Mississippian limestone (Redwall) and the bottom of the Permian Fort Apache Member of the Supai Formation as marker horizons, we find that this Pennsylvanian-Permian interval in the Carrizo-Salt River area is 610 m (2,000 ft) thick. At Fossil Creek, the thickness is 503 m (1650 ft) and at Oak Creek Canyon, about 380 m (1,250 ft). Most of this thinning appears to be in rocks of Pennsylvanian age. In addition to the simple thinning, a lateral facies change is represented by loss of limestone west of Canyon Creek. These changes are found beneath the lowest bed in which conglomerates appear all across the region. The presence of Desmoinesian fauna in a thin marine limestone near the base of the Pennsylvanian section at Fossil Creek, coupled with a Desmoinesian through Virgilian marine fauna at Garrizo-Salt River, suggests an offlap relationship. This helps to explain the contrasts in both thickness and facies that are seen mostly within Pennsylvanian rocks. In central Arizona, these changes might be explained more naturally by a waning Naco Formation history than by the onset of a contrasting complex Supai depositional history.

Brew (1965) studied the stratigraphy of the Naco Formation in outcrop along the Mogollon Rim southeastward to the point where Paleozoic rocks pass beneath the volcanic rocks of the White Mountain region. He divided this formation into Alpha, Beta, and Gamma members from the base upward. The Alpha Member, representing the red clastic zone related to the karstic surface that is extensive above the Redwall Limestone in this region (Huddle and Dobrovolny, 1952), ranges from 12 to 27 m in thickness. The Beta Member makes up most of the Naco Formation in the rim region. The thickest complete section of Beta Member measured by Brew is 210 m. The member may be about 256 m thick in the southeast where complete continuous sections are not known. Brew (1965, p. 50) wrote: ——"the member is a richly fossiliferous succession of ledge-forming gray, brownish-gray, and olive gray limestone with interspersed intervals of slope-forming calcareous shale, shaly mudstone, mudstone, and less common siltstone * * * *."

According to Brew, the most complete depositional record preserved in the Beta Member is to the southeast. He suggested that a cliff-making section 55 m thick near Black River but not present to the northwest, probably represents the northernmost extension of the typical Pennsylvanian Horquilla Limestone, the oldest formation of the Naco Group (Gilluly and others, 1954) of southeast Arizona. At Black River, this Horquilla-like part of the Naco section is indicated by Brew to be middle and upper Desmoinesian in age.

Brew (1965, p. 57) made the following observation of the higher Missourian part of the Beta Member at Black River: "—one horizon contains a lenticular bed of conglomerate with red quartzite pebbles and quartz sand in the matrix. The source area of the quartzite is as yet undetermined, but the presence of this bed clearly indicates a departure from the prevailing marine conditions."

The highest member of the Naco Formation of Brew in this central Arizona region is the Gamma Member. It is, in parts of the region, readily distinguished from the underlying Beta Member because of the presence of more red and brown clastic rocks and less limestone, which leads to longer slopes. Brew suggested that this unit is a transition between the marine conditions of the Naco Formation and the nonmarine conditions of the lower Supai Formation and that it ranges in thickness from 21 m at Fossil Creek Canyon, the northwesternmost locality, to about 91 m toward the southeast. Brew stated that this member is difficult to define objectively in the northwestern sections because fossiliferous limestone is absent, red beds of the Naco Formation are present below the member, and red beds of the Supai Formation are above. Brew noted, but did not emphasize, conglomerates to the southeast in the Gamma Member. It seems surprising that at Fossil Creek Canyon he drew an upper boundary of the Gamma—therefore, the top of the Naco Formation—below the entire section that contains conglomerates, the best exposed set of conglomerates along the rim. He also suggested that the lower contact of the Gamma Member is time-transgressive, being older to the west and younger to the east. More specifically, he visualized this contact to the east to be very latest Pennsylvanian in age (possibly even lowest Permian) and Desmoinesian at Fossil Creek to the west. I think that the conglomerate and floral data previously mentioned may have significant correlative value. At Fossil Creek, the coaly unit (not recognized by Brew), probably either latest Virgilian or earliest Wolfcampian in age, is estimated to be approximately 76 m above Brew's Gamma Member top in what he calls Supai Formation of Missourian age. Brew's age designations and definitions of a Gamma Member at Fossil Creek, as he himself hints, are probably subject to revision.

As pointed out, Brew's attention was attracted to a conglomerate in his Missourian (Beta) section to the southeast because it contained quartzite clasts and quartz sand. He described "intraformational" conglomerates in several sections of the Gamma Member. In a later study, Peirce and others (1977) recognized Gamma Member conglomerates that contain large feldspar and quartz grits, as well as quartzite, chert, and calcarenite clasts. These are not "intraformational" conglomerates. They are evidence of tectonic activity along the Defiance positive area that was to the east and northeast. They are not directly related to a Supai Formation delta prograding towards the southeast, Brew's explanation for the seeming younging of the Gamma Member to the southeast. More likely, transgressive-regressive cycles, responsive to regional tectonic or climatic activity, are the major cause for the shallow-water to subaerial environments that alternate vertically. That land areas supporting lush vegetation existed from time to time is suggested by plant fossils and carbonaceous zones that are closely related to conglomerates. In the transition phase (Gamma Member) of Brew, little evidence is seen of deposition in a classic deltaic environment, which is frequently invoked to explain Supai Formation-Naco Formation relationships. The hiatuses are probably longer than usually is recognized. Havenor (*in* Kottlowski and Havenor, 1962, p. 79) is credited with suggesting: "—that deltaic beds are only a minor part of the Supai sequence, believing that the lower Supai red beds and carbonate rocks (Gamma Member of Brew) are predominately shallow-water marine deposits laid down in a shallow ephemeral sea—where a lowering of sea level by only a few feet may have exposed several hundred square miles of the sea bottom."

The entire Naco Formation in the Mogollon Rim outcrop region is as much as 360 m (1,180 ft) thick in the eastern region. To the west, both in outcrop and in the subsurface, it thins and changes to strata referred to the Supai Formation. To the east in the subsurface, it may thicken before it rather abruptly wedges against the Precambrian rocks of the Defiance positive area paleogeographic feature. The environments that prevailed along the interface of the Naco seaway and the positive area in which sand-producing granite was exposed are not known and have not been tested by drilling. The subsurface onlap and pinchout of Pennsylvanian strata to the northeast has been discussed by Lokke (1962) and Peirce (1970; 1976; 1977).

To the south, but still in central Arizona, isolated exposures of Pennsylvanian strata are found in various mountain ranges where they are everywhere truncated beneath Cretaceous (Clifton-Morenci, Deer Creek, Christmas) or Tertiary (Superior, Mescal Mountains, and so forth) rocks.

Although the name Naco Formation is used in the Mogollon Rim region, the Naco Limestone of Stoyanow (1936) is in use farther south to describe Pennsylvanian rocks. However, still farther south, where Permian marine strata overlie Pennsylvanian strata, the nomenclature is again changed, and the names Black Prince Limestone, Horquilla Limestone, and Earp Formation are used to describe the Pennsylvanian and Pennsylvanian-Permian section.

Occasionally, attempts are made to carry or extend the southern terminology northward. Recall that Brew noted a possible representative of the Horquilla Limestone at Black River. Also, the transitional aspect of the Gamma Member is analogous in some ways to the Earp Formation farther south (Ross, 1973). The thickest of the partial Pennsylvanian sections approximates 427 m (1,400 ft) at Coolidge Dam to the east and at Superior to the west. Kottlowski and Havenor (1962) noted that, although limestone and siliciclastic rocks are present in about equal amounts at Superior, the limestone at the Coolidge Dam locality is more abundant. This west-to-east increase in limestone may bear a direct relationship to the west-to-east increase in limestone along the rim 100 km to the north. This implies a lithologic variance trend that would lie in the northeast-southwest quadrants. Evidence, which is pointed out later, suggests that such a trend might well have extended into southwestern Arizona.

SOUTHEASTERN REGION

In southeastern Arizona, Pennsylvanian as well as Permian strata are contained within the Naco Group. Units now considered to be Pennsylvanian in age are the basal Black Prince Limestone and overlying Horquilla Limestone. The Earp Formation overlies the latter and is believed to contain an undefined Pennsylvanian-Permian boundary.

The Black Prince Limestone was named by Gilluly and others (1954) and originally was considered Mississippian in age, but a Pennsylvanian age is not totally discounted. Although thinner, the base of the Black Prince in several sections appears analogous to the Alpha Member of the Naco Formation of Brew to the north. The similarity is derived from the erosion (solution) of the underlying cherty carbonate rocks of the Mississippian Escabrosa Limestone and the production of red beds and associated, often reworked, residual chert. The unit, ranging in thickness from 36 to 85 m, contains limestones not unlike the Escabrosa and Horquilla Limestones. Later, Nations (1963) concluded that this formation is lowermost Pennsylvanian (Morrowan) in age, and that the Mississippian fossils were reworked from below. The Black Prince Limestone is not everywhere recognized in southeastern Arizona; where it is absent, the Horquilla Limestone (Naco limestone as restricted by Stoyanow) overlies the Mississippian Escabrosa Limestone or, in extreme southeastern Arizona, the Paradise Formation.

The Horquilla Limestone ranges from 305 to 400 m (1,000 to 1,600 ft) in thickness and contains some beds of red and green mudstone that impart a topographic character of steep slopes containing many ledges. The Horquilla Limestone is more abundantly fossiliferous than the underlying Mississippian strata and contains fusulinids, which the latter does not (Bryant, 1968). Most of the Pennsylvanian strata in southern Arizona are contained in the Horquilla Limestone, and most of the Pennsylvanian section of Brew is contained in the Beta Member of the Naco Formation.

Ross (1973) did a detailed study of the Pennsylvanian and early Permian depositional history of southeastern Arizona. He included the Mogollon Rim region as far west as Fossil Creek Canyon. Although this work is much too detailed to be included here, the regional integration accomplished is commendable. Among his conclusions are: (1) the Black Prince Limestone is recognizable as far northwest as the Superior section, (2) the conglomerate-bearing section at Fossil Creek contains the Pennsylvanian-Permian boundary, (3) the Naco Formation of Brew in central Arizona can be treated as the Horquilla Limestone and Earp Formation, and (4) depositional history is complex and involves concurrent faulting as well as many hiatuses.

The westernmost exposure of Pennsylvanian strata of the southeastern type occurs in the Vekol Mountains in the southwestern corner of Pinal County. Ross (1973) suggested that the remaining 183 m (600 ft) of limestone represents the lower part of the Horquilla Limestone and that the section consists of about 80 percent carbonate rocks, a percentage known elsewhere only in extreme southeastern Arizona. About 161 km (100 mi) to the northwest, in the Harquahala Mountains, Varga (1977, p. 6) reported that Supai Formation was found above Redwall Limestone. He described the Supai Formation as consisting dominantly of "quartzite interbedded with minor limestone and phyllite layers," the section, though folded and metamorphosed, approximating 365 m (1,200 ft) in thickness.

The contrast between the Vekol and Harquahala sections demands explanation, whether forthcoming or not. These sections might not now have the same geographic relationships, one to the other, as when originally deposited. However, a northeast-trending line between the Supai and Naco formations in central Arizona can be projected between these sections. Perhaps a regional major northeast depositional strike is involved that could bear some relationship to Stoyanow's "Mazatzal land" trend.

GENERAL PALEONTOLOGY

The following remarks are directed more toward the contributions to stratigraphic understanding made through paleontological study than to a listing of fossils.

The paleontology of Arizona Carboniferous rocks has not been exhaustively treated in any single published work. Whereas the Grand Canyon and Mogollon Rim of northern Arizona offer outcrop continuity, the Basin and Range province of southern Arizona presents a disconcerting discontinuity, and a plethora of local studies is the result. Only some very general highlights are offered here.

Most of the fossils attributed to the Arizona Carboniferous lived in marine environments. Although plant remains are relatively scarce, they may offer hope for making progress on the Pennsylvanian-Permian boundary question in central Arizona. The recognition of an extensive plant-fossil community in that region is such a new development that opportunities for original research still remain. Peirce and others (1977, p. 49) noted an unidentifiable bone fragment in a conglomerate from the Gamma Member of the Naco Formation of Brew (1965), thus hinting at another potential source of historical information.

The volume of paleontological literature seems to be weighted in favor of Pennsylvanian rocks, where Fusulinacea offer a special attraction. However, this and the study of microfossils is a late development because the earlier workers focused attention on the megafossil groups. Today, research effort seems to be directed to the study of conodonts as a possible means for refined zonation.

The earliest reference to Carboniferous strata in northern Arizona is attributed to Marcou (1856). Gilbert (1875) suspected that certain fossils in Grand Canyon represented "Lower Carboniferous" and other fossils, a "Coal Measures" fauna. Darton (1925) presented the first general summary of Arizona geology and included a section on the "Carboniferous System," where Carboniferous fossils are listed. Early studies clearly show that the general zonation of Mississippian strata was well outlined in the early 1900's.

MISSISSIPPIAN

Much of the pioneer paleontological work was done by scientists of the U.S. Geological Survey. G. H. Girty, in particular, studied many of the collections made by the eminent early field geologists such as N. H. Darton and F. L. Ransome. It was

Girty, as reported in Ransome (1904), who recognized Mississippian and Pennsylvanian fossils in the Bisbee quadrangle. In Girty's opinion, both Kinderhookian and Osagean time were represented, but Chesterian time was not. Furthermore, he thought that certain forms represented a slightly younger age than Osagean. These conclusions have not been seriously changed by subsequent research.

The absence of Chesterian strata in southern Arizona, except in the Chiricahua Mountains (Paradise Formation), now is recognized throughout the Basin and Range province wherever Mississippian rocks have been studied. Girty's deductions appear to have stemmed largely from his knowledge of brachiopods.

Stoyanow (1926, 1936) briefly reviewed Paleozoic correlations in Arizona by using paleontology as a major tool. He emphasized the paleontological separation of Mississippian and Pennsylvanian limestones where they had been lumped together as a map unit. He recognized that Carboniferous geologic history could not properly be unraveled until the two systems were mapped separately. He extended to central Arizona the region in which upper Mississippian strata are absent. At the same time, he defined the upper Mississippian Paradise Formation in the Chiricahua Mountains of extreme southeastern Arizona, recognizing these strata as being late Meramecian and early Chesterian in age. Stoyanow apparently used crinoids, brachiopods, and bryozoans to advantage.

At the time of Stoyanow's writing (1936), the fossils of the Grand Canyon Redwall Limestone had not been studied in any detail. However, a section of Redwall Limestone at Jerome, on the Grand Canyon side of Stoyanow's Mazatzal land, had been studied by C. E. Wooddell (1927), a Stoyanow student. From this work, Stoyanow noted forms in the Redwall Limestone not present in the Escabrosa Limestone and vice versa (at a much later date, Sando (1964) was to study corals of the Redwall Limestone and draw a similar conclusion). Stoyanow (1936) also said that the Redwall Limestone section at Jerome was late Kinderhookian and Osagean in age, a conclusion that apparently remains valid.

McKee and Gutschick (1969) collected extensively from the Redwall Limestone of the Plateau region. Their collections contain 17 animal groups and 1 plant group. Many of these groups were studied extensively by specialists, and the results were reported by McKee and Gutschick (1969).

Elsewhere, McKee (1969) wrote that the larger Redwall fossil groups are in distinctive associations, the most important being the coral-brachiopod-crinoid, foraminifer-brachiopod, and brachiopod-bryozoan. Many of the Redwall fossils are either local or long ranging, and thus of little stratigraphic value. However, the foraminifers, brachiopods, and corals are useful zone indicators.

According to Betty Skipp (*in* McKee and Gutschick, 1969, p. 173–256), the foraminiferal succession in the Redwall Limestone ranges from late Kinderhookian to middle Meramecian in age. McKee and Gutschick, referring to brachiopod studies then in progress by J. T. Dutro, Jr., said that these forms range in age from late Kinderhookian into Meramecian, and that Chesterian is represented in one section at Bright Angel Trail. In summarizing paleontological studies, these workers concluded that all other fossil groups represent lesser parts of the stratigraphic section and are of ages between the extremes indicated.

W. J. Sando (also *in* McKee and Gutschick, 1969, p. 257–343) provided an interesting discussion of corals. Among other things, he suggested that the Redwall Limestone is linked more closely to the Madison Group of the northern Cordilleran region than to the Escabrosa Limestone of the southern region.

Some preliminary conodont studies by students at Arizona State University (Norby, 1971; Racey, 1974; Walter, 1976) suggest that detailed conodont work might assist in determining close zonation of Mississipian rocks. Norby recorded a more complete conodont zonation for the Escabrosa Limestone of southern Arizona than did Racey in central Arizona, or Walter to the northwest in Redwall country. Currently, W. J. Purves, a University of Arizona doctoral candidate, is making a comprehensive study of the Mississippian of Arizona with emphasis on conodont zonation and detailed carbonate petrology.

PENNSYLVANIAN

The recognition of Pennsylvanian fossils, quite naturally, was contemporaneous with the studies that led to the recognition of juxtaposed Mississippian forms. Noble (1922) pointed out that it was F. B. Meek who identified Pennsylvanian forms for Gilbert (1875) in western Grand Canyon in beds that Noble later correlated with the lowest unit of Noble's redefined Supai Formation (C) and that McKee (1975a) called the Watahomigi Formation. Although McKee stated that Pennsylvanian fossils bear direct relationships to conglomerate beds, his type section descriptions do not reflect the presence of fossils. McKee (1975b, p. 297) later said: "The discovery * * * of invertebrate fossils, including fusulinids and brachiopods, extensively distributed in the Grand Canyon region throughout the rocks called Supai has greatly clarified the age relations of various rock units in that formation * * *" We hope that future clarification of this statement will shed additional light on the specific fossil localities that for so long have escaped the attention of many geologists.

Darton (1925) cited many lists of fossils collected from various localities within the Basin and Range province that were identified, largely by G. H. Girty, as Pennsylvanian. Girty's studies also led to the recognition that strata lumped into Ransome's (1904) original Naco limestone contained Permian (Hueco) forms.

Gilluly, Cooper, and Williams (1954) raised the Naco to group status and differentiated six formations in central Cochise County. Their Pennsylvanian representatives were the Horquilla Limestone (Atokan to mid-Virgilian?) and Earp Formation (Virgilian and basal Wolfcampian). Nations (1963) determined that the Black Prince Limestone beneath the Horquilla Limestone also is Pennsylvanian in age (Morrowan). Bryant (1955) gave considerable attention to the Earp Formation, and Rea and Bryant (1968) discussed an important marker conglomerate within the Earp Formation.

Both Winters (1963) and Brew (1965) provided essential paleontological data relevant to Pennsylvanian stratigraphy in the Mogollon Rim region. Lokke (1962) provided a surface to subsurface correlation of Pennsylvanian strata along the southern edge of the plateau. He said that fossiliferous surface rocks tend to change to an unfossiliferous redbed sequence northward, and this change complicates dating and correlation of strata northward. He emphasized this point by noting that whereas 32 fusulinid-bearing intervals were recognized in the surface section (Salt River-Carrizo area), only one interval was observed in the subsurface about 70 km to the north. Fortunately, this one interval afforded a time horizon that enabled him to conclude that, northward (p. 85): "* * * significant thinning of Pennsylvanian sediments must be recognized in addition to the previously described (by other workers) interfingering of red clastics with Naco subsurface equivalents."

Blazey (1971) investigated a plant fossil locality in a newly opened uranium prospect (Peirce and others, 1977) under Promontory Butte along the rim, and studied both macrofossils and microfossils. He identified 21 species in 18 genera of macrofossils, and 41 species in 29 genera of microfossils. This is a significant locality because it is within a sequence of "Supai Formation" clastic rocks that otherwise are not fossiliferous. According to Blazey (p. 30):

> The floral composition, although not unusual, has two rather striking features in contrast to other floras of late Paleozoic age. These are: (1) the complete lack of lycopsids, which are almost always associated with form genera of sphenopsids and pteridophylls, and (2) the association of younger typically Permian types with older typically Pennsylvanian types. Together these features present a transitional character to the floral composition.

Blazey concluded that the plant zone is near the Pennsylvanian-Permian boundary, and he seemed to thin that it was probably Permian (Wolfcampian). This zone probably correlates with the coaly bed at Fossil Creek 40 km to the west.

Brew (1965) described his Gamma Member of the Naco Formation at Carrizo Creek as "Post-Virgilian," but not Wolfcampian, in age. One wonders if both these workers saw transitional floras and faunas. Peirce and others (1977), on the basis of conglomerates, correlated these two zones. The Promontory Butte locality does not contain the limestone that is present at sections to the east, but this difference is attributed to lateral changes influenced by mild tectonic activity west of Canyon Creek.

Ross (1973) reported the results of the most detailed stratigraphic zonation of Pennsylvanian strata ever undertaken in the Arizona Basin and Range province, including the Mogollon Rim as far west as Fossil Creek Canyon. This work is an excellent example of the kind of stratigraphic detail that is made possible by meticulous paleontological zonation. In this study, stratigraphic refinement evolved, in large part, from the study of fusulinids. However, in the absence of fossil data, especially westward along the Mogollon Rim, geologists must make subjective decisions, and I think that Ross' correlations at Fossil Creek should seriously be questioned.

Whereas Pennsylvanian seas made few inroads into the northwestern and northeastern corners of Arizona, they transgressed extensively in the southeastern corner. Ross depicts northwesterly transgression and divides the depositional framework into basinal and flanking-shelf facies. Pennsylvanian marine strata of Desmoinesian age extended as far north as Fossil Creek in central Arizona. However, the seas here were ephemeral and regressed south and east for the remainder of Pennsylvanian time. True basinal facies formed only in the Pedregosa basin of extreme southeastern Arizona where, according to Ross, the combined Pennsylvanian-lower Permian sequence attains a maximum thickness of

TABLE 1.—*Principal users of Carboniferous rocks and products in 1977*

Location (Number shown in fig. 1)	Producer (Jan. 1, 1978)	Formation source, products, and facilities
	Apache County	
Field (oil and minor gas)		
1. Teec Nos Pos, Navajo Indian Reservation.	Energy Reserves Group	Pennsylvanian Paradox Formation; 3 wells.
2. East Boundary Butte, Navajo Indian Reservation.	Merrion and Bayless	Pennsylvanian Paradox Formation; 4 wells.
3. Dry Mesa, Navajo Indian Reservation.	Monsanto Co	Mississippian Redwall Limestone; 3 wells.
4. Dineh-bi-Keyah, Navajo Indian Reservation.	Kerr-McGee Corp	Igneous sill in Pennsylvanian Paradox Formation; 18 wells.
Total oil production 1954–77	16.2 million barrels	
	Yavapai County	
Cave (commercial)		
5. Highway 66, Grand Canyon Caverns.		In: Mississippian Redwall Limestone.
Limestone products		
6. Nelson siding, Santa Fe Railroad.	U.S. Lime Division, The Flintkote Co.	Mississippian Redwall Limestone; quarry and kilns; lime products.
7. Clarkdale	Amcord, Inc., Phoenix Cement Co.	Mississippian Redwall Limestone; quarry and plant; cement.
	Gila County	
8. Near Miami	Inspiration Consolidated Copper Co.	Mississippian Escabrosa Limestone; quarry; metallurgical flux and lime for copper processing.
	Greenlee County	
9. Morenci	Phelps Dodge Corp	Mississippian Modoc Limestone; quarry; metallurgical flux and lime for copper processing.
	Pinal County	
10. Hayden	(1) McFarland-Hullinger for ASARCO; (2) Kennecott Copper Corp.	Mississippian Escabrosa Limestone; 2 mills and smelters; 2 quarries; metallurgical flux and lime for copper processing.
11. 6 miles southeast of San Manuel	Magma Copper Corp	Pennsylvanian Horquilla Limestone; quarry; metallurgical flux and lime for copper processing.
	Pima County	
12. Rillito	Arizona Portland Cement Co., Div. Calif. Portland Cement Co.	Mississippian Redwall Limestone and Pennsylvanian Horquilla Limestone; quarry and plant; cement.
13. North end Santa Rita Mountains	Andrada Marble Co	Mississippian Escabrosa Limestone (marble); roofing granules, feed additive, landscaping.

Total Carboniferous quarried (1976): estimated 2.5 million short tons.
Quicklime and cement produced (1976): estimated 1.2 million short tons.
Value of quicklime and cement produced (1976): estimated $42 million.

about 1,800 m (6,000 ft). According to Ross, faulting during Pennsylvanian time influenced lithofacies distribution.

UTILIZATION OF CARBONIFEROUS ROCKS

Carboniferous rocks in Arizona are used both directly (materials quarried and processed) and indirectly (commercialized caverns; sources of petroleum). Table 1 presents a summary of principal exploitation sites, producers, uses, and some quantitative estimates of production and value. We estimate that the value level for all uses of Carboniferous rocks in Arizona has reached $50 million per year. Cement and quicklime are the two major products sold in interstate commerce, constituting about 80 percent of this estimated total dollar value. Much of the remainder is produced by internal operations managed by copper producers in connection with the processing of copper ores and concentrates, principally lime for pH control in mill flotation circuits and as limestone flux in smelters. Use in acid neutralization may increase in the future. The varied uses of limestone and additional data about limestone in Arizona have been discussed by Keith (1969).

The prime contributors to the limestone-derived products in Arizona are the two Mississippian carbonate units. The Redwall Limestone in and near the Plateau province supports two major limestone products operations (Nos. 6 and 7 in fig. 1 and table 1), and the Escabrosa Limestone in the Basin and Range province supports one large cement plant west of Tucson (no. 12).

Petroleum production in Arizona is miniscule, amounting to slightly more than 16 million barrels since the first discovery in 1954. Most of this has been produced since 1967 when the largest field, Dineh-bi-Keyah (Navajo for "The People's Field"), was discovered. Although much of the petroleum production in the Four Corners region is closely related to Carboniferous (Pennsylvanian) rocks, the reservoir at Dineh-bi-Keyah is an igneous sill in Pennsylvanian strata. The sill is believed to be early Tertiary in age and occupies the crestal part of an anticlinal structure. In general, the petroleum potential in northern Arizona is limited by the large region of thin nonmarine Pennsylvanian strata (fig. 3). Although petroleum occurrences of significance are not known in southern Arizona, effort is being made to evaluate deep structural attributes in that area.

A solution cave, such as the well-patronized Grand Canyon Caverns (no. 5) west of Flagstaff on busy Route 66, is an example of indirect usage. This cave extends downward from the present surface, which happens to be the exhumed Redwall Limestone-Supai Formation contact. The present cave may be related to the karst surface that formed in Late Mississippian-Early Pennsylvanian time. The cave seems devoid of speleothems and therefore appears inactive. Perhaps the original dissolving action is a Paleozoic feature, and the more recent history has been devoted to washing out in-filled debris.

Carboniferous rocks were important to southern Arizona's mineralization and ore-development history. Many of the larger metallic mineral districts began with the mining of rich replacement deposits that formed in altered Carboniferous limestone. The close association of these limestones with mineralization explains their early study by geologists of the U.S. Geological Survey. Even today, the underground mining operation by the Magma Copper Co. at Superior is exploiting a copper-ore replacement body in the Mississippian Escabrosa Limestone.

In many southern Arizona localities, the Escabrosa and other limestones have been metamorphosed; one product is white marble, from which granules are used to make a reflective roof covering (no. 13).

Although there is no Arizona history of actual uranium extraction from Pennsylvanian rocks, anomalous uranium content is associated with possible Pennsylvanian carbonized plant zones along the Mogollon Rim. Certain petroleum tests north of the Mogollon Rim indicate that carbonaceous debris is widespread in the subsurface. A potential uranium source is the Precambrian granitic rock against which Pennsylvanian strata abut. Arkosic sandstone could exist near this unconformity and constitute a possible host for petroleum and/or uraniferous occurrences.

Carboniferous rocks constitute a vital part of Arizona's mineral resource base. Growth of the Southwestern United States assures a continuing and expanding demand for such fundamental products as cement and lime. Almost every building in Arizona is partly constructed from products derived from the processing of Carboniferous rocks.

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Idaho

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Historical review and summary of
areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in Idaho

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—IDAHO

By Betty Skipp, W. J. Sando, and W. E. Hall

ABSTRACT

Pennsylvanian and Mississippian rocks of inner and outer cratonic-platform origins are present in south-central and southeastern Idaho. They are in elongate, generally north- or northwest-trending linear fault-block mountains on both sides of the Snake River Plain, an east-northeast-trending upper Cenozoic volcano-tectonic depression that separates outcrops in south-central Idaho from those of the southeastern part of the State.

North of the plain, in south-central Idaho, formation of a Lower Mississippian linear north-trending flysch trough east of the Antler highland is recorded in the lower part of the Copper Basin Formation. The Copper Basin comprises at least 3,500 m of conglomerate, sandstone, limestone, and argillite, which spread eastward across the shelf (McGowan Creek Formation) to the inner craton margin, the site of accumulation of a coeval carbonate-bank complex (Madison Group). South of the plain, a Lower Mississippian carbonate complex (Madison Group) also formed on the inner craton margin, but time-equivalent strata west of the carbonate-bank margin include relatively deep water carbonate rocks (Lodgepole Limestone) overlain by a deep-water starved-basin-facies complex (lower part of Little Flat Formation and others). Orogenic detritus is largely absent.

In Late Mississippian time, north of the plain, the flysch trough gradually shoaled. Shallow marine limestone, interbedded with bouldery mudstone, conglomerate, and minor coaly beds (upper part of Copper Basin Formation), is limited mostly to the area of the trough. A few conglomerate and sandstone lenses are present in the westernmost exposures (White Knob Limestone) of the thick Upper Mississippian carbonate-bank complex (Middle Canyon, Scott Peak, Surrett Canyon Formations), which occupied the outer cratonic platform from the inner cratonic margin west to the flysch trough. The inner cratonic platform was the site of deposition of thin sequences of shallow marine to restricted marine shale, sandstone, and limestone (Big Snowy Formation). In latest Mississippian time, a sheet of cratonic mudstone and sandstone (Big Snowy, Arco Hills, and Bluebird Mountain Formations) spread westward across the carbonate-bank complex.

South of the plain, in Late Mississippian time, the inner cratonic platform was the site of accumulation of the shallow and restricted marine sediments of the Amsden Formation. West of the inner cratonic margin, bank limestone interbedded with dark marine shale (Deep Creek and Monroe Canyon Formations, Great Blue Limestone) prograded westward, possibly as far as the Black Pine Range. Metamorphosed Mississippian rocks west of this, in the Albion Range (View Formation of Armstrong), have an orogenic aspect. In latest Mississippian time, mudstone and sandstone, probably of inner cratonic origin, spread across the carbonate-bank complex around the low broad Bannock highland, in much the same way as did similar deposits north of the plain.

Early Pennsylvanian time, both north and south of the plain, records a continuation of the conditions established in Late Mississippian time. In Middle Pennsylvanian time, however, the setting changed. North of the plain, the Copper Basin highland emerged, separating the Wood River basin on the site of the former Antler highland on the west, from a shallow-water carbonate bank (Snaky Canyon Formation) to the east. The limestone and sandstone of the carbonate bank give way both north and east to thick sandstone of the Quadrant Formation. South of the plain, the Bannock highland, which had no sediment accumulation in either Late Mississippian or Early Pennsylvanian time, became submerged, and thin carbonate and sandy carbonate deposits (Wells Formation) spread across it. To the west, a northern arm of the Oquirrh basin, the Sublett basin, subsided and received more than 2,300 m of sandy carbonate and calcareous sandstone (lower part of Oquirrh Formation); these deposits are present as far west as the Albion Range.

Though the paleotectonic settings of Carboniferous sedimentation differ on the two sides of the Snake River Plain, there are enough similarities to indicate that they formed in response to the same continent-margin tectonics. Two previously proposed plate-tectonic models seem feasible. The first is that of a gradually closing inner-arc basin between the edge of the continent and an island arc above an east-dipping subduction zone. The Antler orogeny would have taken place in response to a period of increased plate motion along the subduction zone. Because the Antler orogeny never underwent a molasse phase but was succeeded, in Pennsylvanian time, by a return to shelf conditions, an alternate model has been proposed. In this second model, the Antler orogeny is the product of an arc-continent collision followed by a period of arc withdrawal and tensional tectonics on the continent that produced the highlands and basins characteristic of Pennsylvanian time.

The Carboniferous rocks of eastern Idaho have a potential for hydrocarbons, though none had been produced through 1977. Metalliferous deposits in Carboniferous rocks north of the plain have produced a moderate recovery of almost $20 million in lead-silver, molybdenum, copper, zinc, barite, gold, and tungsten. The thick high-calcium limestones

of the Upper Mississippian carbonate-bank complex remain an untapped potential resource, as do vast resources of less pure limestones.

INTRODUCTION

Carboniferous (Mississippian and Pennsylvanian) rocks are known to crop out only in the central, southern, and eastern parts of Idaho, east of the granitic rocks of the Upper Cretaceous Idaho batholith (fig. 1). Within this part of the State, the outcrops are separated into two distinct areas by the Snake River Plain, a late Cenozoic volcano-tectonic depression, 80 to 90 km wide, which is filled with basalt, related rhyolite and rhyolitic ash, and margin-derived sediments. Carboniferous rocks north and south of the Snake River Plain have different names, markedly different stratigraphic successions, and related plate-tectonic histories. Both north and south of the plain, outcrops are present along the crests and flanks of narrow northwest-trending, north-trending, or arcuate, fault-bounded mountains (fig. 2), which rise to a maximum height of 3,859 m. Structural relations and several deep wells indicate that Carboniferous rocks are present at depth in many, but not all, of the intervening valleys. There is no information at this time on the presence, absence, or extent of Paleozoic rocks under the thick lavas and sediments of the eastern Snake River Plain, though geophysical studies and deep drilling programs are in progress. The nature of the bedrock/basalt contact along the northern margin of the plain suggests that Carboniferous rocks, which dip gently under, or are foundered in, basalt, probably extend under the plain a kilometer or so (fig. 3).

No rocks of definite Carboniferous age have been reported from Idaho west of the Idaho batholith, though deepwater stratified sedimentary and volcanic rocks in northeastern Oregon contain Pennsylvanian fusulinids (D. L. Bostwick *in* Vallier, 1977, p. 7), and correlative strata may be present in Idaho near lat 45° on the western edge of the State (fig. 1; Vallier, 1974; Vallier and others, 1977).

Several areas of scenic outcrops of Carboniferous rocks are present within Idaho. In almost every area the strata are folded, faulted, or sheared, but exposures are continuous enough to work out successions. Steeply dipping conglomerate, sandstone, argillite, and limestone of the Mississippian Copper Basin Formation are spectacularly exposed along the crest of the Pioneer Mountains (figs. 4, 5). Massive to thin-bedded folded limestones of the Upper Mississippian Scott Peak, South Creek, Surrett Canyon, and Bluebird Mountain Formations are well exposed along Antelope Creek in the White Knob Mountains (fig. 6) and make up the crests of the Lost River Range (fig. 7), the Lemhi Range, and the Beaverhead Mountains in several areas. The same formations are particularly well exposed on the southwest corner of the Lemhi Range in the general area of several type localities (Huh, 1967). The Snaky Canyon Formation is well exposed on the east sides of the southern Lost River and Lemhi Ranges and the southern Beaverhead Mountains (fig. 4).

South of the plain, Mississippian rocks of the Madison Group and Pennsylvanian rocks of the Wells Formation make up the ridge crests of the northern Snake River Range. Farther west, Mississippian rocks are exposed low on the flanks of the Portneuf and Chesterfield Ranges, the Deep Creek Mountains, and the Sublett and Albion Ranges; Pennsylvanian and Permian rocks of the Oquirrh Formation make up the crests of the ranges where the strata are moderately well exposed (see fig. 15).

Mean average rainfall within the area of outcrop of Carboniferous rocks (fig. 2) ranges from less than 20 cm per year in the Snake River Plain to more than 100 cm per year in the mountains bordering the plain (Travis and others, 1964, fig. 56). Frost action has affected Carboniferous rocks in all the mountainous areas, resulting in extensive colluvial cover in many outcrop areas. Large landslides of probable Pleistocene age are concentrated locally in shale and shaly limestone interbedded with thick-bedded Upper Mississippian limestone in the ranges north of the Snake River Plain (Radbruch-Hall and others, 1976).

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Idaho Bureau of Mines and Geology.

HISTORY

The first scientific explorers to pass through Idaho probably were Lewis and Clark in 1805 (De Voto, 1953). Geologic mapping, which included study of Carboniferous rocks, was undertaken in 1877 in southeastern Idaho by parties of the Hayden Survey. The work resulted in publications by A. C. Peale and Orestes St. John in 1879 (Mansfield, 1927). After these early investigations, several workers published significant reports in the early decades of the 1900's that included studies of Car-

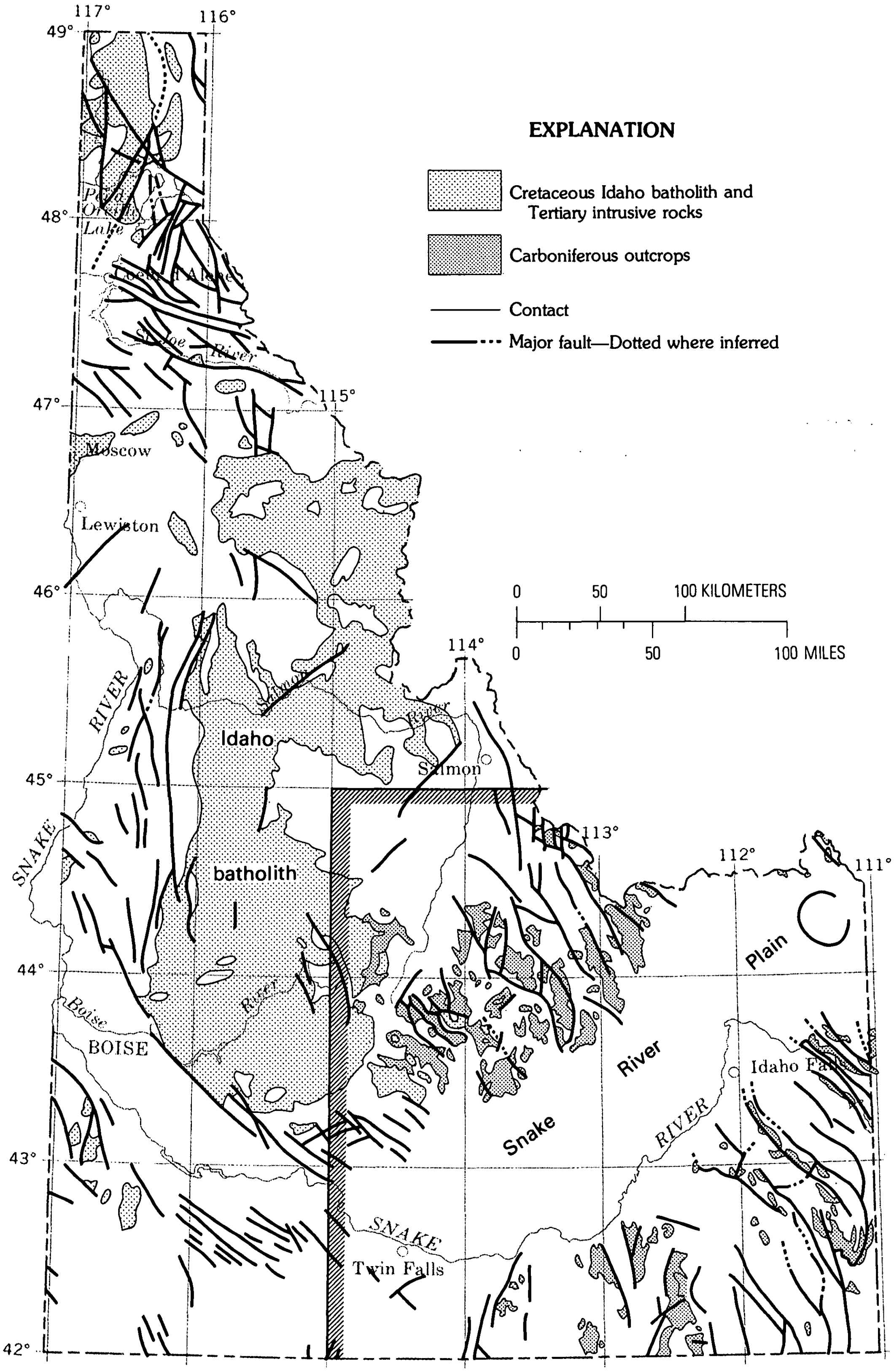


FIGURE 1.—Index map of Idaho showing outcrops of Carboniferous rocks and the location of major faults, the Snake River Plain, the Cretaceous Idaho batholith, and Tertiary intrusive rocks. Area of figures 2, 4, 9, 10, 12, and 13 shown by hachures. Modified from King and Beikman (1974).

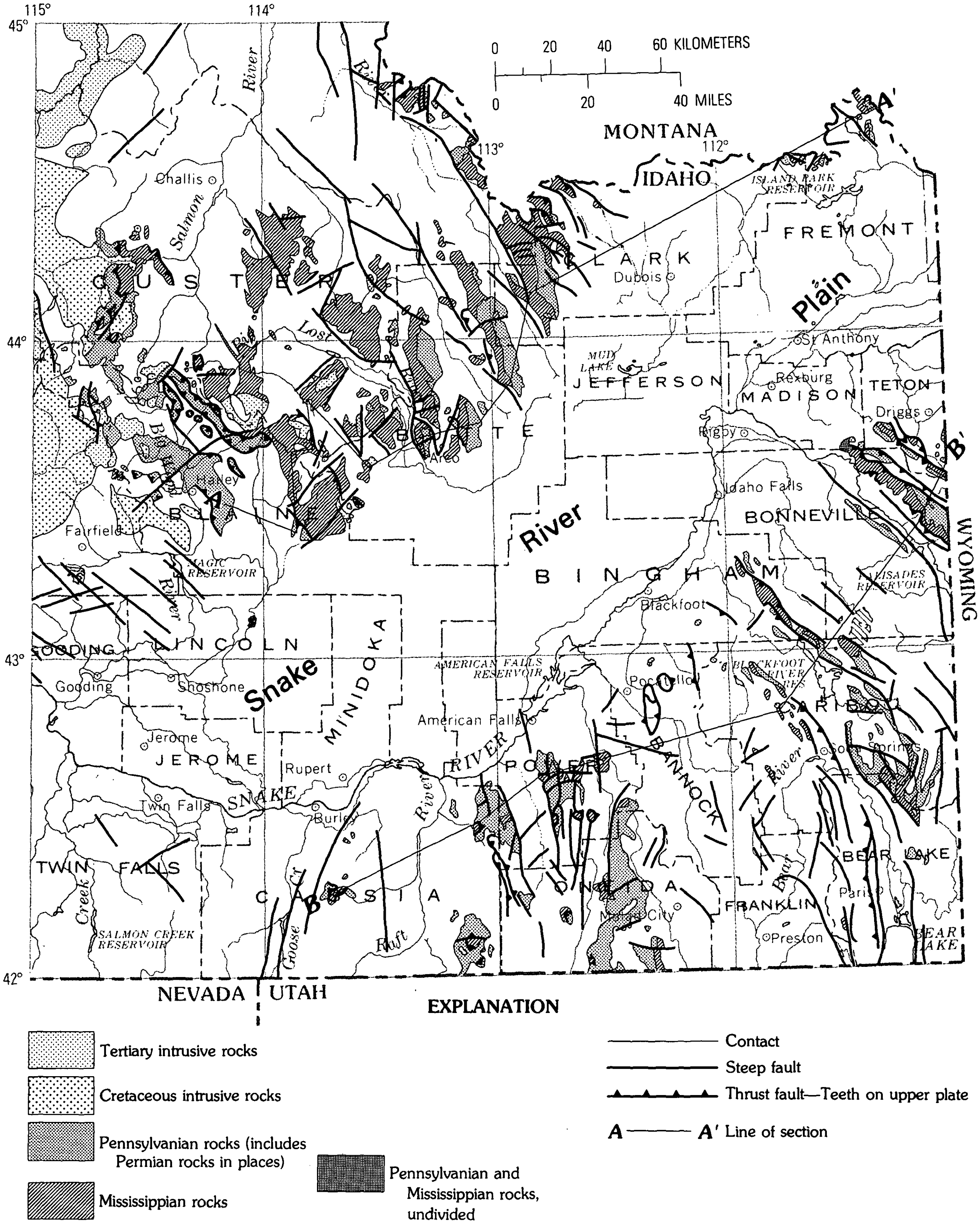


FIGURE 2.—(For caption see facing page.)

FIGURE 3.—Oblique aerial photograph of north edge of Snake River Plain showing outcrop of sandstone, conglomerate, and argillite in lower part of Copper Basin Formation (foreground) foundered in late Quaternary basalt of the Snake River Group. Outcrop of Mississippian rocks (about 100 m wide at bottom edge of photo) is just west of Craters-of-the-Moon National Monument in sec. 32, T. 1 N., R. 23 E., Blaine County. Photograph by W. B. Hall.

FIGURE 2.—Map showing outcrops of Carboniferous (Mississippian and Pennsylvanian) rocks in central and southeastern Idaho, location of Cretaceous and Tertiary intrusive rocks, major faults that affect distribution of the Carboniferous rocks, and locations of cross sections *A–A'* and *B–B'* of figures 11 and 14. The map was compiled from the following sources:

Armstrong, R. L., unpub. data
Bond, J. G. (1978)
Cress, L. D., unpub. data
Dover, Hall, Hobbs, Tschanz, Batchelder, and Simons (1976)
Hait, M. H., Jr., unpub. data
Hall, Rye, and Doe (1978)
Hall, W. E., and Batchelder, John, unpub. data
Hays, W. H., unpub. data
Hobbs, Hays, and McIntyre (1975)
King and Beikman (1974)
Lucchitta, B. K. (1966)
Mapel, Read, and Smith (1965)
Mapel and Shropshire (1973)
Mapel, W. J., unpub. data
Platt, L. B. (1977), and unpub. data
Prostka and Hackman (1974)
Pruit, J. D. (1971)
Ross, C. P. (1937)
Ruppel, E. T. (1968)
Skipp, Betty, unpub. data
Skipp and Hait (1977)
Smith, J. F., Jr., unpub. data
Thompson, M. E. (1977)
Trimble and Carr (1976)
Tschanz, Kiilsgaard, and Seeland, in Tschanz and others (1974)
Tschanz, C. M., unpub. data
Witkind, I. J. (1972, 1976)

LOCALITIES NORTH OF THE SNAKE RIVER PLAIN

1. Big Wood River area. T. 2 N., R. 19 E. Blaine Co. Hall, Batchelder and Douglass (1974); W. E. Hall, unpub. data.

1a. Wilson Creek Ridge. T. 5 N., R. 19 E. Blaine Co. W. E. Hall and J. N. Batchelder, unpub. data.

1b. Dollarhide Summit area. T. 3 N., R. 15 E. Blaine and Camas Cos. W. E. Hall and J. N. Batchelder, unpub. data.

2. Fish Creek Reservoir area. T. 1 N., R. 22 E. Blaine Co. Skipp and Hall (1975a).

3. Fish Creek Reservoir area. T. 1 N., R. 22 E. Blaine Co. Skipp and Hall (1975a).

4. Cottonwood Creek area. T. 2 N., R. 23 E. Blaine Co. Larson (1974).

5. Iron Mine Creek area. T. 2 N., R. 23 E. Blaine Co. Bollmann (1971); Betty Skipp, unpub. data.

6. Central Pioneer Mountains. T. 4 N., R. 21 E.; T. 4 N., R. 22 E. Custer Co. Paull, Wolbrink, Volkman, and Grover (1972); Paull and Gruber (1977); Nilsen (1977); J. H. Dover, unpub. data.

7. Salmon River-Clayton area. T. 11 N., R. 16 E. Custer Co. Hobbs, Hays, McIntyre (1975); S. W. Hobbs, unpub. data.

8. Iron Bog Creek area. T. 4 N., R. 23 E. Custer Co. R. A. Paull and Betty Skipp, unpub. data.

9. Cabin Creek area. T. 6 N., R. 22 E. Custer Co. Skipp (1961a); Skipp and Mamet (1970); R. A. Paull, unpub. data.

10. Hurst Canyon. T. 6 N., R. 28 E. Butte Co. Sandberg, Mapel, and Huddle (1967).

11. Timbered Dome area. T. 3 N., R. 25 E. Butte Co. Skipp (1961b); Betty Skipp, unpub. data.

12. Antelope Creek-Wood Canyon areas composite. T. 4 N., R. 24 E. and T. 5 N., R. 25 E. Custer Co. Betty Skipp, unpub. data.

13. Howe Peak-Arco Hills, T. 4 N., R. 28 E.; T. 5 N., R. 29 E. Butte Co. Shannon (1961); Breuninger (1976); Betty Skipp and R. D. Hoggan, unpub. data.

14. Arco Hills. T. 4 N., R. 26 E. Butte Co. Wornardt (1958); Roberts (1979); Betty Skipp, unpub. data.

15. Doublespring area. T. 12 N., R. 21 E. and T. 12 N., R. 22 E. Custer Co. Mapel, Read, and Smith (1965); Mamet, Skipp, Sando, and Mapel (1971).

16. McGowan Creek. T. 12 N., R. 21 E. Custer Co. Mapel, Read, and Smith (1965); Sandberg (1975).

17. Slate Creek. T. 10 N., R. 15 E. Custer Co. Thomasson (1959); Tschanz, Kiilsgaard, Seeland (*in* Tschanz and others, 1974).

18. Upper Pahsimeroi area. T. 9 N., R. 23 E. Custer Co. Huh (1967); Roberts (1979).

19. Donkey Hills. T. 10 N., R. 25 E. Custer Co. Mapel and Shropshire (1973; Roberts (1979).

20. Hawley Mountain. T. 9 N., R. 26 E. Butte Co. Mamet, Skipp, Sando, and Mapel (1971); Mapel and Shropshire (1973).

21. Hawley Mountain. T. 8 N., R. 26 E. Butte Co. Mamet, Skipp, Sando, and Mapel (1971); Mapel and Shropshire (1973).

22. Lower Cedar Creek. T. 7 N., R. 24 E. Custer Co. Sandberg, Mapel, and Huddle (1967); Sandberg (1975).

23. East and Box Canyons-southern Lemhi Range. T. 6 N., R. 30 E. Butte Co. Shannon (1961); Huh (1967); Sandberg (1975); Skipp, Hoggan, Schleicher, and Douglass (1979).

24. Copper Mountain area. T. 10 N., R. 30 E. and T. 10 N., R. 31 E. Clark Co. Huh (1967); Embree, Hoggan, Williams, and Skipp (1975); Skipp, Hoggan, Schleicher, and Douglass (1979), and unpub. data.

25. Snaky Canyon area. T. 9 N., R. 32 E. Clark Co. Skipp, Hoggan, Schleicher, and Douglass (1979).

26. Deadman Lake area. T. 16 S., R. 10 W. Beaverhead Co., Montana. Huh (1967).

27. Leadore area. T. 16 N., R. 26 E., and T. 16 N., R. 27 E. Lemhi Co. Ruppel (1968); E. T. Ruppel, unpub. data.

28. Hawley Creek area (generalized). T. 16 N., R. 27 E.; T. 15 N., R. 28 E.; T. 16 N., R. 28 E. Lemhi Co. Lucchitta (1966); B. K. Lucchitta, unpub. data.

29. Centennial Mountains. T. 14 N., R. 41 E. Fremont and Clark Cos. Witkind (1976).

30. Henrys Lake Mountains. T. 16 N., R. 44 E. Fremont Co. Witkind (1972); E. K. Maughan, unpub. data.

30a. Centennial Mountains. T. 14 N., R. 42 E. Fremont Co. Witkind (1972).

LOCALITIES SOUTH OF THE SNAKE RIVER PLAIN

31. Albion Range. T. 13 S., R. 23 E.; T. 13 S., R. 24 E. Cassia Co. Armstrong (1968).

31a. Albion Range. T. 11 S., R. 24 E. Cassia Co. Armstrong (1968).

32. Black Pine Range. T. 15 S., R. 28 E., T. 15 S., R. 29 E. Cassia Co. J. F. Smith, Jr., unpub data

33. Northern Sublett Range. T. 11 S., R. 29 E., approx. Cassia Co. R. L. Armstrong, unpub. data.

34. Northern Deep Creek Mountains, Hunter Canyon vicinity. T. 9 S., R. 31 E. and T. 9 S., R. 32 E. Power Co. Trimble and Carr (1976); Sando, Dutro, Sandberg, and Mamet (1976).

34a. Northern Deep Creek Mountains, Bannock Peak. T. 9 S., R. 32 E. Power Co. Trimble and Carr (1976).

34b. Hills east of Bannock Creek, T. 9 S., R. 33 E. Power Co. Trimble and Carr (1976).

35. Southern Deep Creek Mountains. T. 11 S., R. 32 E. Power Co. L. D. Cress, unpub. data.

36. Samaria Mountain. T. 16 S., R. 35 E. Oneida Co. Beus (1968), Platt (1977).

37. Chesterfield Range. T. 7 S., R. 40 E. Caribou Co. Dutro and Sando (1963), Sando, Dutro, Sandberg, and Mamet (1976).

37a. Soda Springs quadrangle, northeast corner. T. 8 S., R. 42 E. and T. 8 S., R. 43 E. Caribou Co. Armstrong (1953, 1969).

38. Portneuf Range. T. 9 S., R. 39 E. Caribou Co. Oriel (1968), S. S. Oriel, unpub. data.

39. Big Hole Mountains. T. 4 N., R. 43 E. Teton Co. Staatz and Albee (1966).

40. Black Mountain, Snake River Range. T. 3 N., R. 43 E. Bonneville Co. Sando (1977), Sando and Dutro *in* Roberts (1979).

41. Sheep Creek, Snake River Range. T. 1 N., R. 45 E. Bonneville Co. Jobin and Soister (1964), Sando (1977), Sando and Dutro *in* Roberts (1979).

42. Hell Creek Canyon. T. 1 S., R. 46 E., T. 1 N., R. 46 E. Bonneville Co. Gardner (1944); Roberts (1979).

43. Hot Springs mine. T. 13 S., R. 44 E., Bear Lake Co. Mallory (1975).

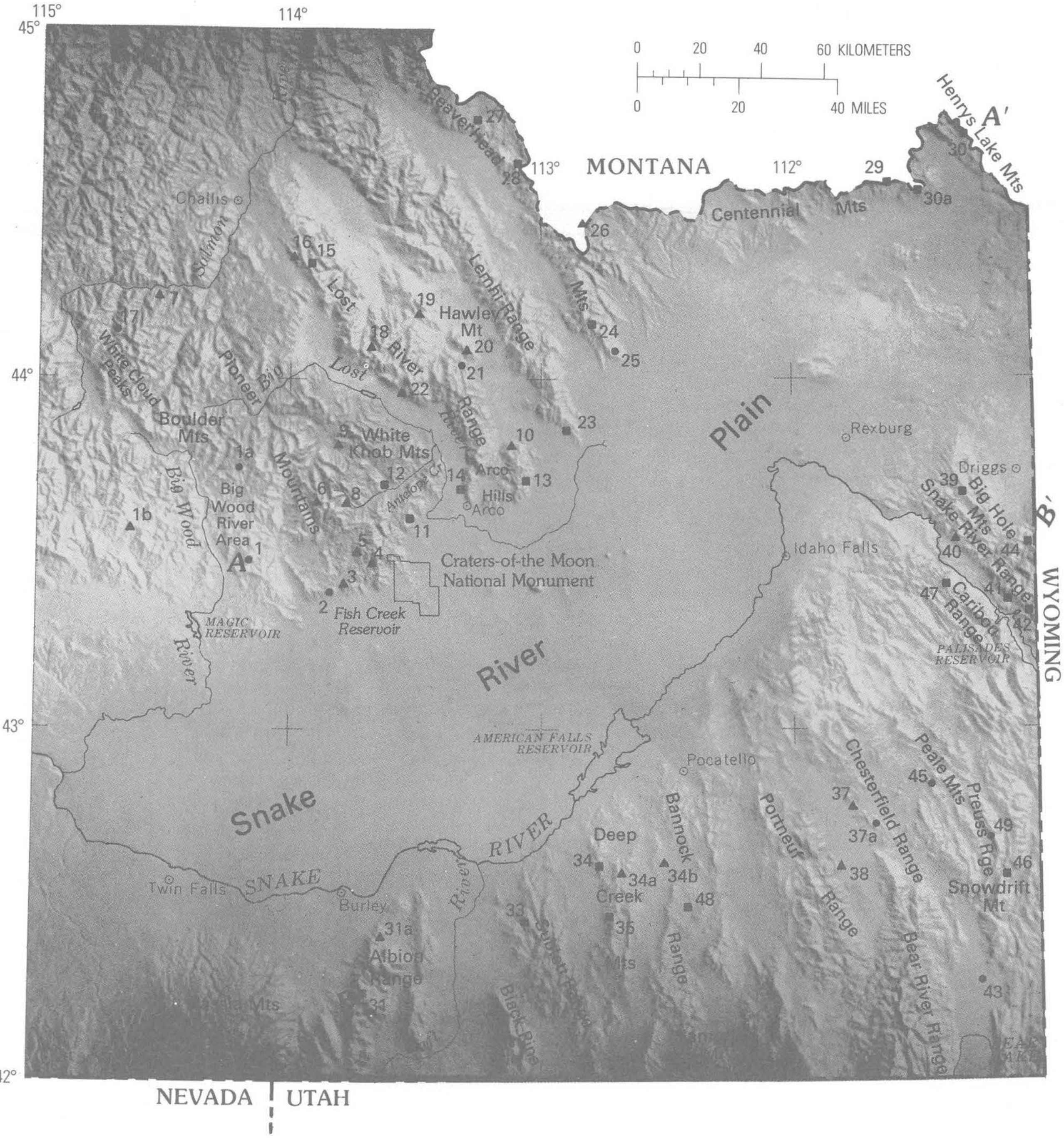


FIGURE 4.—Map showing section localities used in construction of the isopach and lithofacies maps of figures 9, 10, 12, and 13. Triangles indicate Mississippian rocks only; circles, Pennsylvanian rocks only; and squares, both Pennsylvanian and Mississippian rocks. Numbers indicate data sources. Major geographic features referred to in text are shown; also lines of sections *A–A'* and *B–B'* (figs. 11 and 14).

44. Snake River Range. T. 2 N., R. 46 E.; T. 3 N., R. 46 E. Bonneville and Teton Cos. Pampeyan, Schroeder, Schell and Cressman (1967).
45. Wooley Range, Peale Mountains. T. 6 S., R. 42 E., T. 6 S., R. 43 E., Caribou Co. McKelvey *in* Cressman (1964), Mallory (1975).
46. Snowdrift Mountain. T. 9 S., R. 45 E., T. 10 S., R. 45 E. Caribou Co. Gulbrandsen, McLaughlin, Honkala, and Clabaugh (1956); Cressman (1964); Mallory (1975).
47. Caribou Range. T. 1 N., R. 43 E. Bonneville Co. Jobin and Schroeder (1964).
48. Bannock Range. T. 10 S., R. 34 E., T. 11 S., R. 35 E. Bannock Co. Murk (1968).
49. Preuss Range, Peale Mountains. R. 8 S., R. 45 E. (generalized). Caribou Co. Cressman and Gulbrandsen (1955); Montgomery and Cheney (1967).

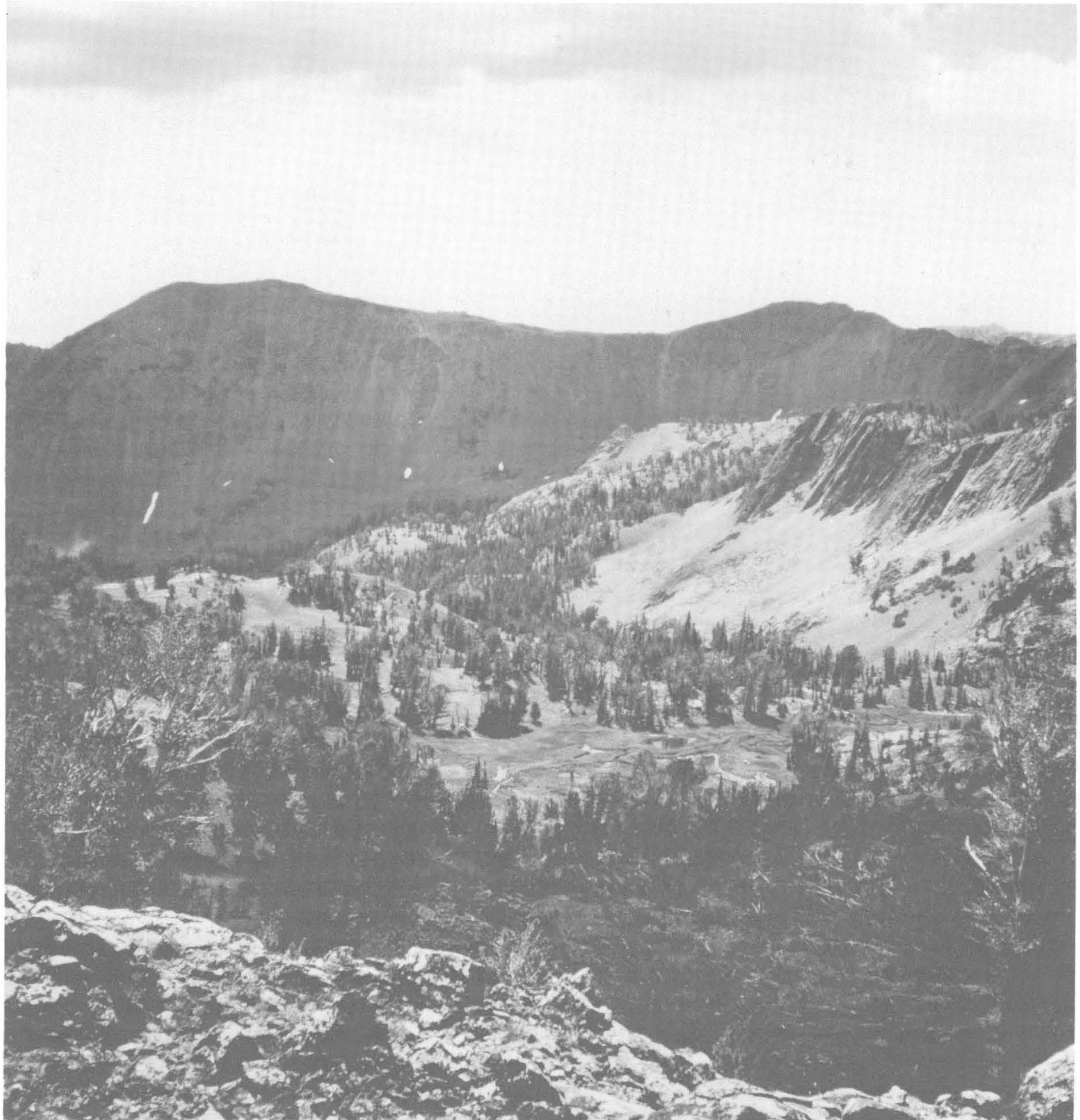


FIGURE 5.—Crest of Pioneer Mountain made up of steeply dipping quartzite, conglomerate, and argillite beds of lower part of Copper Basin Formation. View is taken from above Brockie Lake near locality 6 looking south.

boniferous rocks. Representative reports of this kind are: Schultz and Richards (1913), Umpleby (1913, 1917), Kirkham (1922, 1924, 1927), Mansfield (1927), Anderson (1929, 1931), Umpleby, Westgate, and Ross (1930), and Ross (1934, 1937). Several of the above reports were published by the Idaho Bureau of Mines and Geology. In 1947, the Idaho Bureau, in cooperation with the U.S. Geological Survey, published the "Geologic Map of the State of Idaho" at a scale of 1:500,000 (Ross and Forrester, 1947), which showed the generalized distribution of Carboniferous rocks. A new geologic map of Idaho recently has been published as the result of a similar cooperative effort (Bond, 1978).

In recent years, many workers have published reports dealing primarily with Carboniferous rocks and fossils in various parts of Idaho. These include: Bostwick (1955); Scholten (1957); Thomasson (1959); Ross (1960, 1962a, b, c); Skipp (1961a, b); Carr and Trimble (1961); Churkin (1962); Shannon (1961); Dutro and Sando (1963); Roberts, Crittenden, Tooker, Morris, Hose, and Cheney

FIGURE 6.—Oblique aerial photograph looking north from mouth of Antelope Creek in White Knob Mountains at folded Upper Mississippian and Pennsylvanian beds of the Scott Peak, South Creek, Surrett Canyon, Bluebird Mountain, and Snaky Canyon Formations. Photograph by W. B. Hall.

(1965); Sando (1967, 1976b); Sando, Mamet, and Dutro (1969); Skipp and Mamet (1970); Mamet, Skipp, Sando, and Mapel (1971); Paull, Wolbrink, Volkmann, and Grover (1972); Sandberg (1975); Sando, Gordon, and Dutro (1975); Breuninger (1976); Sando, Dutro, Sandberg, and Mamet (1976); Nilsen (1977); and Paull and Gruber (1977). Regional studies of Carboniferous rocks that have included Idaho have been published by Bissell (1962, 1974); Williams (1962); Schleh (1968); Poole (1974); Mallory (1975); Sando (1976a); Sando, Dutro, Sandberg, and Mamet (1976); Rose (1976); Peterson (1977); Poole and Sandberg (1977); Rich (1977); and Roberts (in press).

POSTDEPOSITIONAL STRUCTURAL SETTING

Some Carboniferous rocks in Idaho are thought to be parautochthonous, but most are allochthonous, having been transported east or northeast along complex thrust- and tear-fault systems; movements have dated from Pennsylvanian(?) to Early Tertiary, possibly extending into Late Tertiary, time (Richards and Mansfield, 1912; Mansfield, 1927; Rubey, 1955; Scholten and others, 1955; Rubey and Hubbert, 1959; Eardley, 1960; Armstrong and Cressman, 1963; Roberts and Thomasson, 1964; Armstrong and Oriel, 1965; Roberts and others, 1965; Armstrong, 1968; Monley, 1971; Beutner, 1972; Scholten, 1973; Hall and others, 1975; Royse and others, 1975; Skipp and Hall, 1975a, b; Ruppel, 1978; and Skipp and Hait, 1977). This list of references dealing with thrust faults in Idaho is representative but not complete.

The thrust faults on which Carboniferous rocks have moved northeast or east are deformed locally by granitic rocks of the Upper Cretaceous Idaho batholith, and are offset by upper Cenozoic basin-and-range extensional structures, steep normal faults that trend generally north or northwest (figs.

FIGURE 7.—Oblique aerial photograph of Arco Hills; crest of Lost River Range (King Mountain) in background. In foreground, Scott Peak Formation forms cliffs at base of hill; South Creek Formation forms slope above; dark cliffs of folded and faulted Surrett Canyon Formation are above the South Creek slope, and, above these, the slopes of the Arco Hills and Bluebird Mountain Formations. Letter "B" on top of ridge is in sandstone of Bluebird Mountain Formation. Photograph by W. B. Hall.

1 and 2). Vertical stratigraphic displacements on basin-and-range structures in Idaho locally are as much as 6,100 m (Skipp and Hait, 1977), though most are much less. The basin-and-range structures, in turn, are cut off or buried beneath the upper Cenozoic lava and sediments of the Snake River Plain (figs. 1 and 2).

The active late Paleozoic to Holocene tectonic history of Idaho has complicated the distribution of Carboniferous rocks. Horizontal displacements on individual thrusts north of the plain have been estimated to be as much as 160 km (Ruppel, 1976) and south of the plain, about 25 km (Armstrong and Oriel, 1965). Total extension on late Cenozoic basin-and-range structures north of the plain is estimated to be about 25 percent of that south of the plain, introducing an apparent right-lateral sense of displacement between outcrops on the two sides of the plain (Skipp, 1976). Both left- and right-slip movements since early Paleozoic time have been suggested along an inferred Snake River fault in the general position of the Snake River Plain (Sandberg and Mapel, 1967; Poole and others, 1977: Poole and Sandberg, 1977).

The isopach maps (figs. 9, 10, 12, and 13) are based on the present distribution of Carboniferous outcrops. Because of structural complexities, however, large isopach intervals have been used, and structure-controlled boundaries have been emphasized. Three major concepts have resulted from this synthesis of structural and stratigraphic data:

1. Carboniferous facies, thicknesses, and structural patterns on the north and south sides of the plain for the time-stratigraphic intervals

shown in figures 10, 12, and 13 are different and suggest separate crustal responses to plate-margin tectonics.

2. Total thrust displacements on both sides of the plain seem to be nearly equal, as the positions of the inner cratonic platform margin and the western edge of the Upper Mississippian carbonate-bank complex (Sando, 1976a; Rose, 1976) do not seem to be displaced significantly.
3. No large strike-slip displacements before late Cenozoic time seem necessary along the axis of the Snake River Plain to account for the present distribution of facies and thicknesses of Carboniferous rocks, according to our interpretation of sedimentary trends.

MISSISSIPPIAN SYSTEM

The Mississippian System of the United States correlates with the Lower Carboniferous and lowermost part of the Upper Carboniferous Series in Europe (fig. 8), and includes, from oldest to youngest, the Kinderhookian, Osagean, Meramecian, and Chesterian Provincial Series.

PALEOTECTONIC SETTING—ANTLER OROGENY

Mississippian rocks in Idaho, both north and south of the Snake River Plain, were deposited across a mobile former continental shelf and a cratonic platform that had been relatively stable from late Precambrian through Late Devonian time. In latest Devonian through earliest Mississippian time, the western edge of the North American continent was disrupted during the Antler orogeny, when Devonian and older rocks of continental-slope and ocean-basin origins were obducted onto the continental margin (Poole, 1974; Poole and Sandberg, 1977). The thrust system along which the translation took place in Nevada, the Roberts Mountains thrust, is nowhere identified with certainty in Idaho (Paull, 1976). Instead, indirect evidence of the tremendous magnitude of the disturbance exists in the more than 4,000 m (fig. 9) of Mississippian turbidites and interturbidites (Copper Basin Formation) that filled a deep and narrow flysch trough along the eastern margin of the Antler orogenic highland (Poole, 1974). At present, the flysch trough is represented largely by an incomplete allochthonous sequence, the Copper Basin Formation, which itself has been folded, faulted, and thrust eastward (Skipp and Hall, 1975b; Skipp and Hait, 1977; Nilsen, 1977). Undated parautochthonous strata, which lithologically resemble parts of the Copper Basin Formation, are present in two areas beneath the Wood River allochthon (structural elements 7 and 8 of fig. 9, and fig. 2).

East of the flysch trough, thick orogenic, thin starved-basin, and thick carbonate sediments were deposited without major interruption in an outer cratonic platform environment. (Sando prefers "foreland basin" usage of Poole and Sandberg (1977) for this feature.) Farther east, in easternmost Idaho, Mississippian rocks were laid down in two depositional cycles (cycles I and III of Sando, 1976a; and Sando, Dutro, Sandberg, and Mamet, 1976) separated by middle and late Meramecian emergence of the inner cratonic platform (Cycle II of Sando, 1976a, and Sando, Dutro, Sandberg, and Mamet, 1976) [1].

UNDERLYING ROCKS

Lower Mississippian rocks rest unconformably on Upper Devonian rocks throughout most of Idaho (fig. 8; Poole and others, 1977, Sandberg and Poole, 1977), except in the White Knob and Pioneer Mountains, where the McGowan Creek and Copper Basin Formations locally unconformably overlie Middle Devonian rocks of the Carey Formation (Skipp and Sandberg, 1975; C. A. Sandberg, written commun., 1974). Elsewhere in the Pioneer Mountains region, the Copper Basin Formation has been thrust over rocks as old as Precambrian(?) (Dover, 1969; Dover and others, 1976). In the Albion Range, Mississippian rocks are thrust over strata as old as Cambrian (R. L. Armstrong, written commun., 1976).

THICKNESS

Mississippian rocks thin to the south and east across Idaho. The thickest recognized Mississippian sequences in Idaho are the flysch of the Copper Basin Formation north of the plain where thickness is greater than 3,896 m (fig. 9). This tremendous thickness drops to near zero at Fish Creek Reservoir (Skipp and Hall, 1975a), where the Copper Basin allochthon (Skipp and Hait, 1977) is overridden by the Wood River and Milligen allochthons (Hall and others, 1975; Skipp and Hait, 1977). In the area designated 7 on figure 9, however, parautochthonous undated metamorphosed turbidite siltite and limestone, which resemble the Drummond Mine Lime-

[1] The terms "outer cratonic platform" and "inner cratonic platform" as used in this paper replace, respectively, "miogeosyncline" and "craton" of former usage. The outer cratonic platform was a mobile marginal cratonic area, which, by the end of Paleozoic time, largely had become a part of the stable craton of the North American continent. The outer cratonic platform was the site of flysch-trough, starved-basin, carbonate-bank, and shelf-basin depositional environments in Carboniferous time.

stone of the Copper Basin Group of Paull, Wolbrink, Volkmann, and Grover (1972), may represent the root zone of part of the Copper Basin allochthon (W. E. Hall, unpub. data). The siltite and limestone is about 1,000 m thick. Another parautochthonous sequence of Mississippian(?) argillite turbidites and deep-sea fan conglomerates, more than 500 m thick, is present in the White Cloud Peaks area (fig. 9, structural element 8) beneath the Wood River allochthon (C. M. Tschanz, written commun., 1977). Zero thickness occurs in the Big Wood River area where the Wood River Formation tectonically overlies the Devonian Milligen Formation. The combined thickness of the White Knob Limestone and McGowan Creek Formation is about 3,000 m in the White Knob Mountains (figs. 4 and 9, locality 9). Thicknesses decrease eastward toward the Idaho/Montana border where less than 500 m is reported

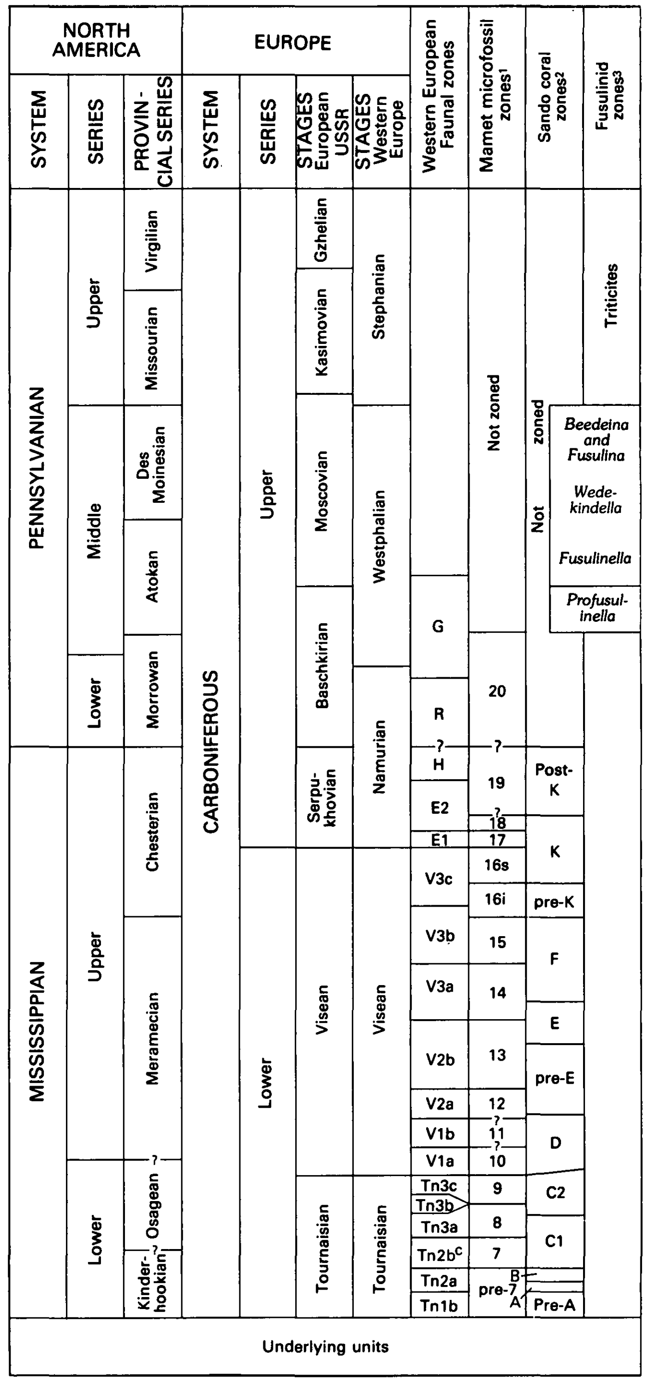


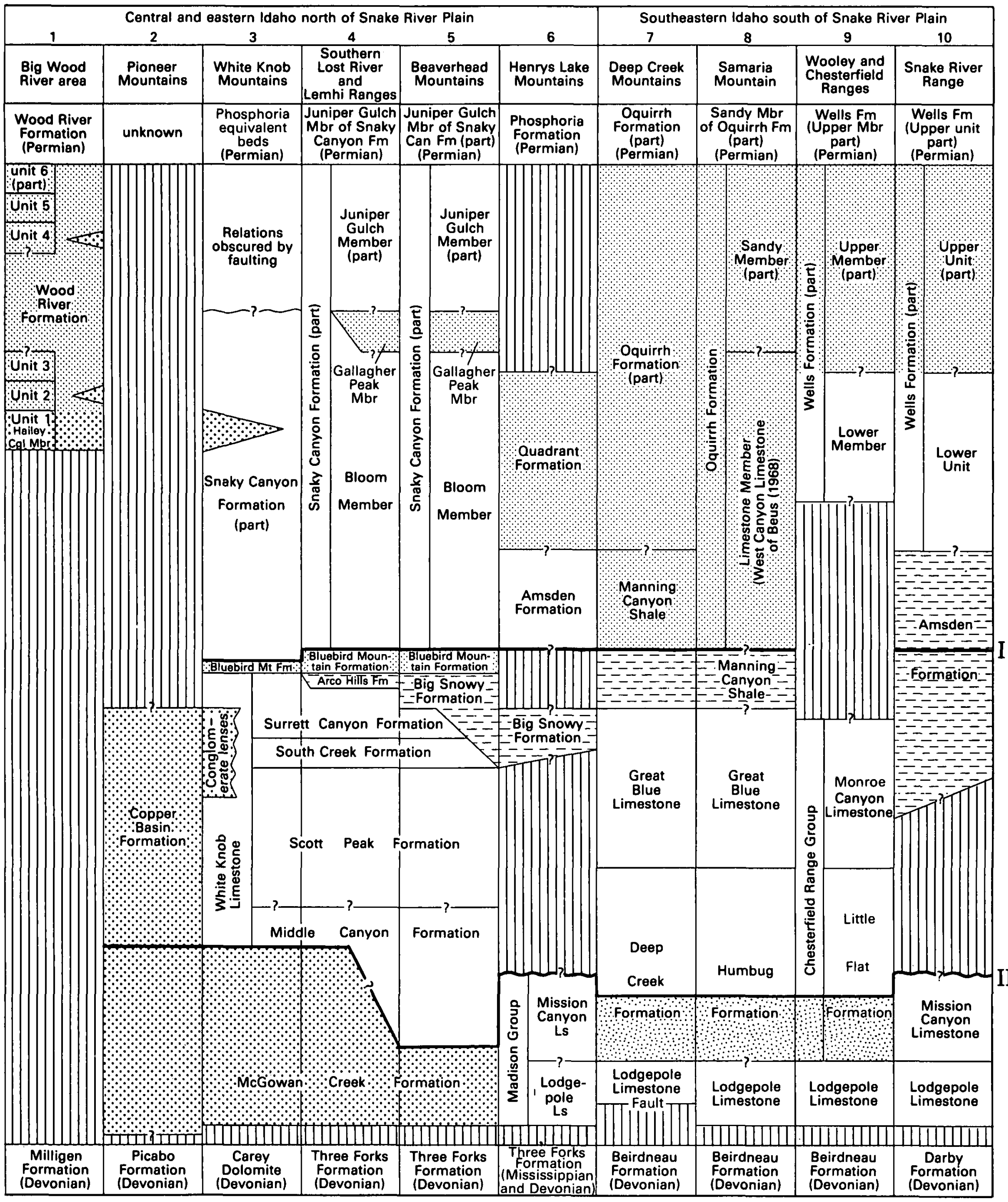
FIGURE 8.—Correlation chart. Numbered column refer to list of sources.

1. Umpleby, Westgate, and Ross (1930); Bostwick (1955); Thomasson (1959); Hall, Batchelder, and Douglass (1974); Skipp and Hall (1975a); Sandberg, Hall, Batchelder, and Axelsen (1975); Hall, W. E. (unpub. data).
2. Ross (1960, 1962a); Paull, Wolbrink, Volkman, and Grover 1972); Yokley (1974); Skipp and Hall (1975a); Skipp and Sandberg (1975); Nilsen (1977); Paull and Gruber (1977); Skipp, Betty (unpub. data).
3. Skipp (1961a); Ross (1962a); Skipp and Mamet (1970); Paull, Wolbrink, Volkman, and Grover (1972); Skipp and Sandberg (1975); Skipp and Hait (1977); Nilsen (1977); Skipp, Hoggan, (Schleicher, and Douglass (1979).
4. Thomasson (1959); Shannon (1961); Sandberg, Mapel, and Huddle (1967); Huh (1967); Mamet, Skipp, Sando, and Mapel (1971); Beutner (1972); Sandberg (1975); Skipp and Hait (1977); Sandberg and Poole (1977); Skipp, Hoggan, Schleicher, and Douglass, 1979; Skipp, Betty (unpub. data).
5. Huh (1967); Scholten and Ramspott (1968); Embree, Hoggan, Williams, and Skipp (1975); Sando, Gordon, and Dutro (1975); Sandberg (1975); Skipp and Hait (1977); Sandberg and Poole (1977); Skipp, Hoggan, Schleicher, and Douglass, 1979.
6. Thompson and Scott (1941); Witkind (1972); Sando, Gordon, and Dutro (1975); Maughan, E. K., (unpub. data).
7. Carr and Trimble (1961); Sando (1967, 1976a); Sando, Mamet, and Dutro (1969); Sando, Dutro, Sandberg, and Mamet (1976); Trimble and Carr (1976); Cress, L. D. (unpub. data, 1978).
8. Beus (1968); Platt (1977).
9. Dutro and Sando (1963); Cressman (1964); Sando (1967, 1977); Sando, Mamet and Dutro (1969); Armstrong (1969); Sando, Dutro, Sandberg, and Mamet (1976).
10. Staatz and Albee (1966); Sando (1967, 1977).

[1] Sando, Mamet, and Dutro (1969); Mamet and Skipp (1970) Brenckle, Lane, and Collinson (1974); Sando, Gordon, and Dutro (1975); Sando, Dutro, Sandberg, and Mamet (1976); Brenckle (1977)

[2] Sando, Mamet, and Dutro (1969); Sando, Dutro, Sandberg, and Mamet (1976); Sando, this volume

[3] Douglass (1977)



EXPLANATION

- Hiatus where rocks are absent because of nondeposition or erosion
- Mostly craton-derived sandstone and sandy carbonate rock
- Mostly craton-derived shale, siltstone, and minor carbonate rock
- Carbonate rocks, inner and outer cratonic platform
- Starved-basin sedimentary rocks
- Orogenic sedimentary rocks derived from Antler, Copper Basin, and Humboldt(?) highlands

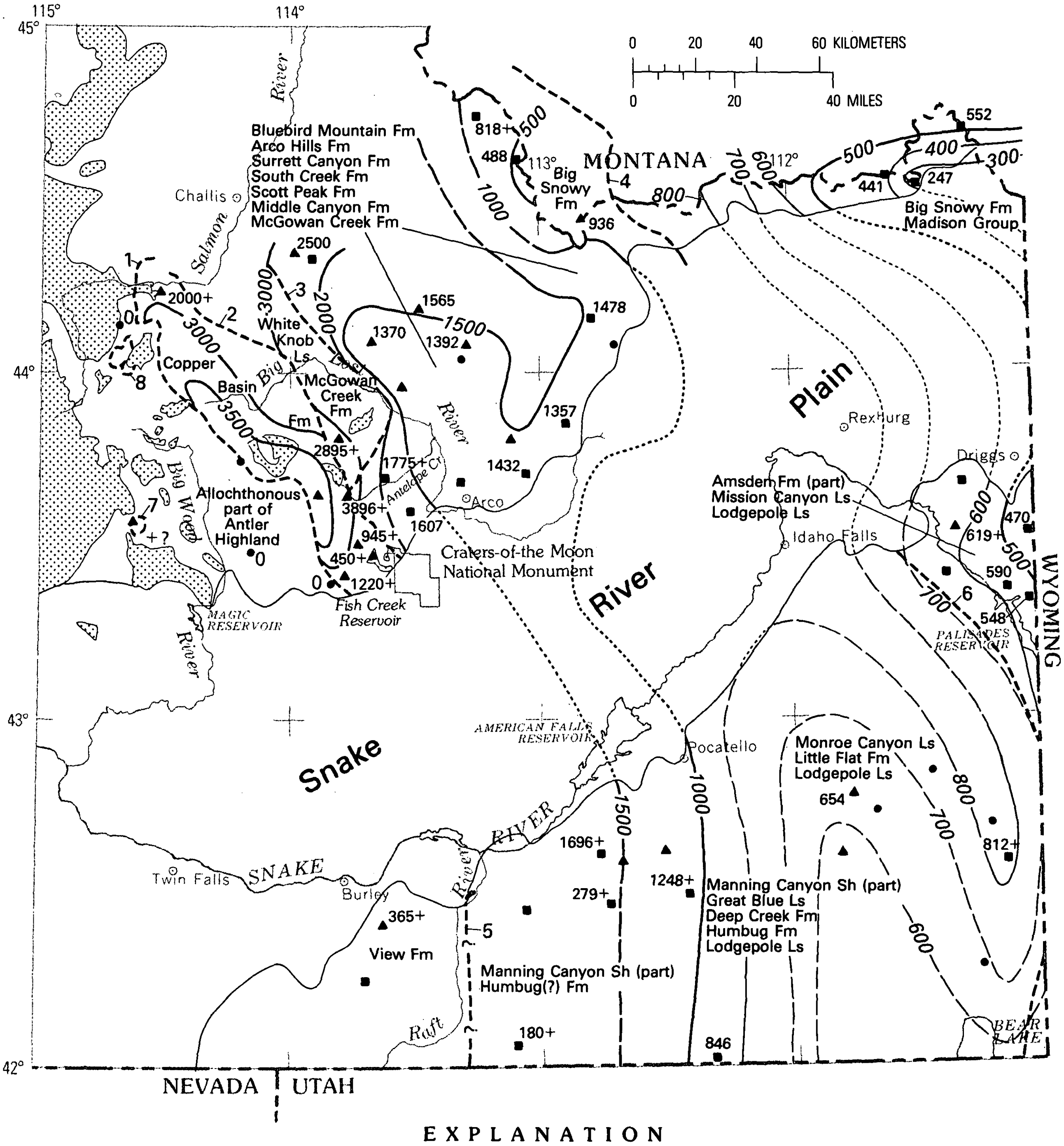
Heavy lines indicate rock-stratigraphic divisions used in construction of figures 9, 10, 12, and 13.

I Separates the rock units included on the Mississippian and Pennsylvanian isopach maps.

II Separates the Lower and lowermost Upper Mississippian rock units from the Upper Mississippian used on figures 10 and 12

—?— Formation or zone boundary—Queried where uncertain

~?~ Erosion surface—Queried where uncertain



EXPLANATION

Tertiary and Cretaceous intrusive rocks

Contact

—500— Isopachs (partly restored) in 500-m and 1000-m intervals—Dashed where inferred; dotted across Snake River Plain

—300— Isopachs (partly restored) in 100-m intervals—Dashed where inferred; dotted across Snake River Plain

1 Structural element or boundary described in caption

▲ ● ■247 Sample localities—Same as on figure 4. Thicknesses in meters (locality 28). The thinnest Mississippian sequences of the cratonic platform in Idaho are in the eastern Centennial Mountains (localities 29 and 30a).

South of the plain, Mississippian rocks gradually thicken westward from 470 m in the Big Hole Mountains (locality 44) to more than 1,696 m in the northern Deep Creek Mountains (locality 34). West of this locality, incomplete and faulted outcrops of Mississippian shale and siltstone are present in parautochthonous sequences beneath thrust plates of Pennsylvanian and Permian rocks in the Sublett, Black Pine, and Albion Ranges (localities 33, 32, and 31).

Mississippian rocks in the western United States were divided into two depositional complexes by Rose (1976), a lower complex encompassing rocks ranging in age from early Kinderhookian to early Meramecian, and an upper complex containing strata of middle Meramecian through middle late Chesterian age. The Manning Canyon Shale and its equivalents were excluded from the concept of the upper depositional complex by Rose (1976). The divisions shown on figures 10 and 12 are based on Rose's depositional-complex concept, but rocks of latest Chesterian age have been included in thickness calculations of Upper Mississippian rocks. The Lower Mississippian complex, thus, generally corresponds to Cycle I, and the Upper Mississippian complex to Cycle III of Sando (1976a) and Sando, Dutro, Sandberg, and Mamet (1976), who divided the cycles into 13 phases and presented a paleogeographic map for each.

LOWER MISSISSIPPIAN COMPLEX

Rocks of Early Mississippian and earliest Late Mississippian age make up the Lower Mississippian complex (fig. 10). Included are those formations between line II and underlying units on the correlation chart (fig. 8). Flysch or foreland-basin, starved-basin and carbonate-platform depositional environments are represented.

North of the plain, most of the Copper Basin Formation, all of the McGowan Creek Formation, and the Madison Group constitute this interval. The Middle Canyon Formation is excluded on the isopach map even though a middle Osagean conodont fauna was recovered about 15 m above the base of the formation in the southern Beaverhead Mountains (Sandberg and Poole, 1977; Poole and Sandberg, 1977). The depositional environment of the Osagean part of the Middle Canyon is considered, however.

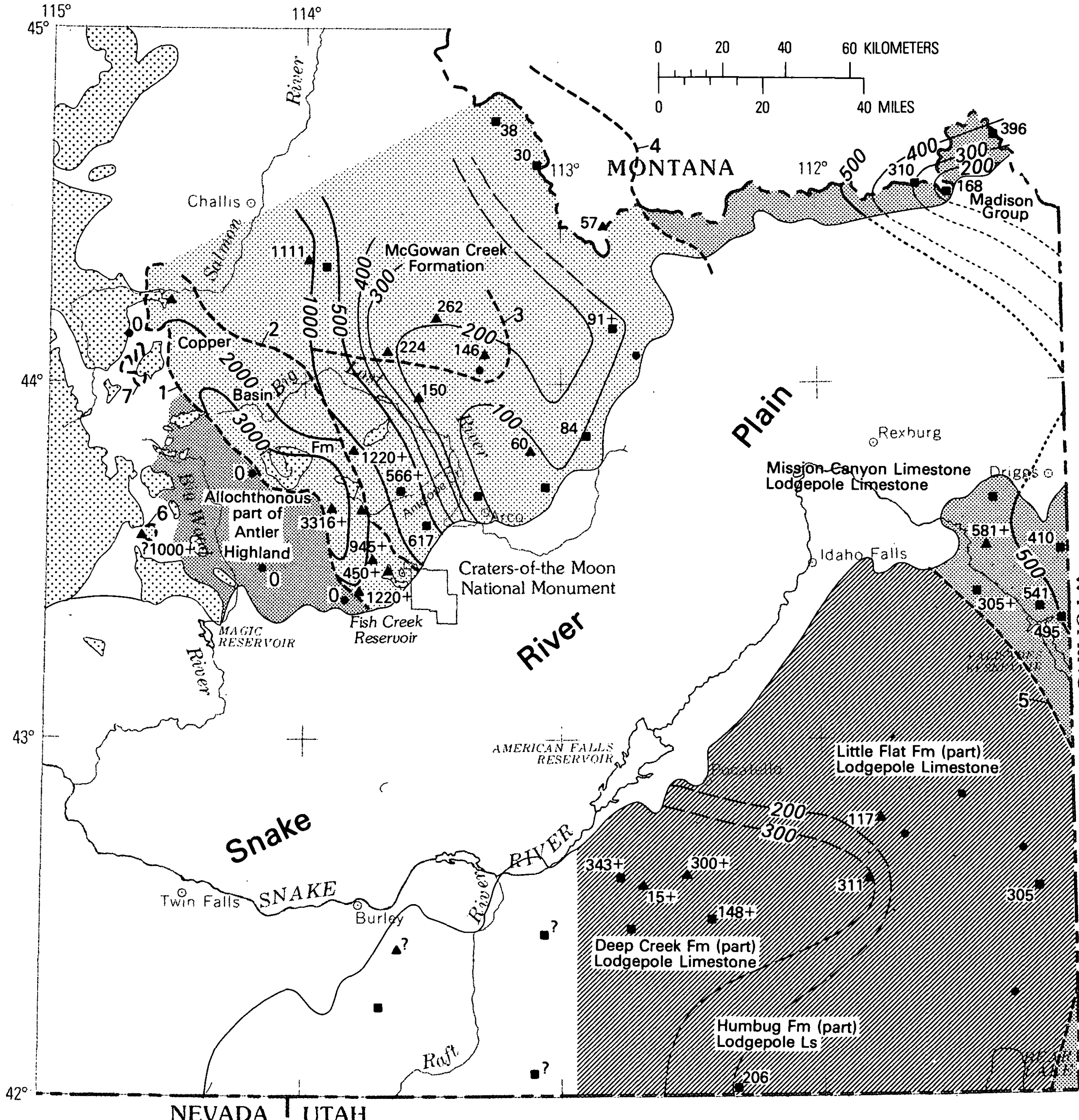
South of the plain, the phosphatic shale member of the Little Flat Formation, an arbitrary lower part of the Humbug Formation (86 m at locality 36), the lower siltstone member of the Deep Creek Formation, the Mission Canyon Limestone, and the Lodgepole Limestone are included.

COPPER BASIN FORMATION (PART)

More than 3,316 m of the Copper Basin Formation, the part of the formation designated the Scorpion plate by Nilsen (1977), is included on figure 10. The Scorpion plate of the Copper Basin allochthon includes, in ascending order, the Little Copper Formation (Paull and Gruber, 1977), the Drummond Mine Limestone, the Scorpion Mountain Formation, and the lower part of the Muldoon Canyon Formation of the Copper Basin Group of Paull, Wolbrink, Volkmann, and Grover (1972). The 18,000 feet (5,500 m) of section reported in 1972 contains a duplicating thrust (Nilsen, 1977; Nilsen, and Skipp, unpub. data); thus, the real thickness is somewhat less, though a complete section has not yet been pieced together. About 4,000 m is shown on figure 11. The Copper Basin Formation is made up of boulder- to granule-conglomerate, sandstone, siltstone, mudstone, and limestone (Paull and others, 1972), interpreted as foreland-basin flysch containing terrigenous turbidites derived from the Antler highland to the west and carbonate turbidites derived in part from cratonic carbonate banks to the

FIGURE 9.—Isopach map of the Mississippian System showing provincial formation names.

1. Structural eastern limit of the Pennsylvanian and Permian Wood River Formation.
2. Structural eastern limit of the Mississippian Copper Basin Formation.
3. Structural eastern limit of Mississippian White Knob Limestone.
4. Structural boundary between the carbonate platform to the west and the inner cratonic platform to the east, and eastern limit of Mississippian McGowan Creek, Middle Canyon, Scott Peak, South Creek, Surrett Canyon, Arco Hills, and Bluebird Mountain Formations. Also boundary between Idaho province and Montana province of Sando (1967).
5. Structural(?) western limit of Upper Mississippian carbonate depositional complex.
6. Boundary between Idaho province (western) and Montana province (eastern) of Sando (1967), in general position of western margin of inner cratonic platform.
7. Structural window of Copper Basin(?) Formation.
8. Structural windows(?) of Copper Basin(?) Formation (C. M. Tschanz, unpub. data).



EXPLANATION

- Tertiary and Cretaceous intrusive rocks
- Antler-derived flysch deposits—Conglomerate, sandstone, siltstone, and mudstone with minor limestone
- Mostly craton-derived deposits—Limestone, phosphatic shale, siltstone, and sandstone
- Cratonic-platform carbonate rocks
- Emergent area
- Contact
- —500— Isopachs (partly restored) in 500-m and 1000-m intervals; dotted across Snake River Plain
- —300— Isopachs (partly restored) in 100-m intervals—Dashed where inferred; dotted across Snake River Plain
- 5 Structural element or boundary described in caption
- ▲ ● ■ 206 Sample localities—Same as on figure 4. Thicknesses in meters

east (Poole, 1974; Poole and Sandberg, 1977; Nilsen, 1977). The age of the Drummond Mine Limestone unit of the Copper Basin Formation is Kinderhookian, as indicated by scattered conodont faunas identified by C. A. Sandberg (Skipp and Hall, 1975a; Paull and Gruber, 1977; Poole and Sandberg, 1977). Sparse ammonoids, trilobites, brachiopods, calcareous foraminifers, and numerous trace fossils also have been recovered from other parts of the flysch sequence considered to be correlative with the Lower Mississippian complex (Skipp and Hall, 1975a; Skipp, 1974; Poole, 1974; Skipp, unpub. data).

McGOWAN CREEK FORMATION

Most of the McGowan Creek Formation, which ranges in thickness from more than 1,220 m in the White Knob Mountains (locality 9) to 30 m in the Beaverhead Mountains (locality 28), is composed of orogenic detritus, recognized as such by Poole (1974) and designated the lower member of the McGowan Creek Formation by Sandberg (1975). This lower member consists of fine-grained, thinly bedded turbidites (Poole, 1974; Sandberg, 1975; Nilsen, 1977; Poole and Sandberg, 1977). The upper member of the McGowan Creek Formation, which ranges in thickness from 60 to 150 m, is present only north and west of Hawley Mountain (figs. 2, 4), where the McGowan Creek is thickest (Sandberg, 1975). This upper member, consisting of calcareous siltstone interbedded with silty micritic limestone, is considered a starved-basin facies (Poole and Sandberg, 1977). The McGowan Creek Formation has been dated by conodonts (Sandberg and others, 1967; Sandberg, 1975). Other fossils are sparse, though scattered brachiopods, mollusks (pelecypods, gastropods, cephalopods), trilobites, corals, and trace fossils have been found.

MADISON GROUP

The Madison Group, consisting of the transgressive Lodgepole Limestone overlain by the regressive Mission Canyon Limestone, is present on both the north and south sides of the plain, where it appears to thicken westward to the inner cratonic platform margin, aggregating more than 581 m in easternmost Idaho (fig. 10; locality 40). The Lodgepole Limestone is divided into the Paine Member and overlying Woodhurst Member. The Paine Member, a sparsely fossiliferous, thin-bedded, fine-grained, silty, and argillaceous limestone, is interpreted as a slope deposit laid down seaward from a shelf-carbonate bank (Sando, 1977). The Woodhurst Member consists of cyclically interbedded fine-grained and coarse-grained carbonate beds representing an alternation of high-energy and low-energy conditions intermediate between those of the Paine Member of the Lodgepole and the overlying Mission Canyon Limestone. The Mission Canyon contains thick beds of crinoidal limestone in the lower part and solution breccias, cherty fine-grained limestone, dolomite, and dolomitic limestone in the upper part. The Mission Canyon Limestone represents a westward-prograding shelf-carbonate envivironment characterized by several episodes of restricted circulation in highly saline lagoons in the upper part (Sando, 1977). Calcareous foraminifers, corals, brachiopods, and encrinitic debris are common in much of the Madison Group; all except the encrinitic debris have been valuable in age determinations.

The Lodgepole Limestone in the ranges west of the inner cratonic platform margin (fig. 8, columns 7, 8, and 9) is represented largely by lithologies like those of the Paine Member overlain by a thin Woodhurst Member. Total thickness in the Chesterfield Range (locality 37) is 79 m (figs. 10 and 11).

FIGURE 10.—Isopach map of the Lower Mississippian and lowermost Upper Mississippian rocks (lower depositional complex of Rose (1976) and cycle I of Sando (1976a), showing Osagean-early Meramecian lithofacies, provincial formation names, and position of allochthonous part of Antler highland.

1. Western structural limit of well-dated Lower Mississippian part of the Copper Basin Formation.
2. Eastern structural limit of the Copper Basin Formation.
3. Southern boundary of starved-basin facies of McGowan Creek Formation (Sandberg, 1975).
4. Eastern structural limit of the McGowan Creek Formation and western limit of lower shelf margin of Rose (1976), and craton margin of Sando (1976a).
5. Approximate western limit of lower carbonate shelf margin of Rose (1976), craton margin of Sando (1976a), and eastern limit of the starved-basin facies (Sandberg and Gutschick, 1977; Poole and Sandberg, 1977).
6. Structural window of Copper Basin(?) Formation.
7. Structural windows of Copper Basin(?) Formation.

Foraminifers, conodonts, and corals have been used to determine its age.

LITTLE FLAT, DEEP CREEK, AND HUMBUG FORMATIONS (PART)

In the Chesterfield Range, the Lodgepole is overlain by the siltstone member of the Little Flat Formation of the Chesterfield Range Group. The siltstone member consists of 38 m of phosphatic siltstone and dark cherty fine-grained limestone representing a starved-basin facies (Rose, 1976; Sando and others, 1976; Poole and Sandberg, 1977). The lower siltstone member of the Deep Creek Formation, 231 m thick at locality 34, has also been interpreted as a starved-basin deposit (fig. 11) (Sando and others, 1976). An arbitrary thickness of the Humbug Formation in the Samaria Mountain area (locality 36), consisting of dark-brown calcareous sandstone and siltstone, is assigned to the starved-basin facies. Conodonts, foraminifers, brachiopods, cephalopods, and fish fragments are present in the siltstone and limestone sequences.

DEPOSITIONAL HISTORY

The depositional histories north and south of the plain during the time frame of the lower depositional complex are somewhat different. The rapid transgression of the Cordilleran sea in late Kinderhookian and early Osagean time north of the plain spread orogenic detritus derived from the Antler highland eastward across the outer cratonic platform to the edge of the inner cratonic platform, and carbonate detritus from the platform spread westward to mingle with deep-water terrigenous turbidites in the flysch trough (Poole, 1974; Sandberg and others, 1975; Nilsen, 1977; Poole and Sandberg, 1977). South of the plain, in the apparent absence of orogenic detritus, carbonate sediments, largely slope and forebank deposits, were laid down in front of a prograding carbonate bank (Sando, 1976a; Rose, 1976; Sandberg and Poole, 1977).

In middle Osagean through early Meramecian time, the sea gradually regressed, forming the carbonate-bank deposits of the Mission Canyon Limestone along the inner cratonic platform margin. West of this area, and south of the plain, a starved basin formed. Relationships between starved-basin and carbonate-bank facies are shown in figure 11.

North of the plain, during the same interval, limestone of the lower part of the Middle Canyon Formation spread westward as forebank deposits in front of the westward-prograding bank limestone of the Mission Canyon Limestone (Sandberg and Poole, 1977; Poole and Sandberg, 1977). The eastward spread of flysch detritus had either ceased by this time or was restricted. Locally, siltstone and limestone of the upper member of the McGowan Creek Formation (fig. 10) grade upward into silty limestone of the overlying Middle Canyon and are interpreted as a starved-basin facies (Sandberg and Poole, 1977; Poole and Sandberg, 1977).

UPPER MISSISSIPPIAN COMPLEX

Rocks of Late Mississippian age make up the Upper Mississippian complex (fig. 12). Included are those formations between lines I and II on the correlation chart (fig. 8). Sediments of a shoaling flysch basin, carbonate bank, lagoon, and/or restricted marine basin are represented.

North of the plain, a poorly defined upper part of the Copper Basin Formation, the White Knob Limestone, the Middle Canyon, South Creek, Scott Peak, Surrett Canyon, Arco Hills, Bluebird Mountain, and Big Snowy Formations, are included. South of the plain, the upper chert-banded limestone member of the Deep Creek Formation, an arbitrary upper part of the Humbug Formation, the cherty limestone, sandstone, and sandy limestone members, in ascending order, of the Little Flat Formation, the Great Blue Limestone, the lower part or all of the Manning Canyon Shale, and the lower part of the Amsden Formation, make up the interval.

COPPER BASIN FORMATION (PART)

An incomplete part of the Copper Basin Formation having estimated minimum thicknesses of 580 and 2,000 m is shown on figure 12; about 1,000 m is estimated on figure 11. In the vicinity of locality 8, this part of the formation includes conglomerate, mudstone, pebbly mudstone, sandstone, scattered lenses of shallow-water fossiliferous marine limestone and sandy dolomite, and thin coal seams deposited in a marginal-marine to shallow-marine environment (Skipp and Hait, 1977; Nilsen, 1977). These shallow-water sedimentary rocks grade downward into turbidites that closely resemble those of the lower part of the formation. The upper part of the Copper Basin Formation is bounded by thrust faults (Skipp and Hait, 1977). The thickness at locality 8 is an estimate, as this part of the sequence was not included in the measured sections of Paull, Wolbrink, Volkmann, and Grover (1972). The thickness at locality 7 is an estimate of thickness by S. W. Hobbs (oral commun., 1976) of an isoclinally folded sequence of relatively fine grained

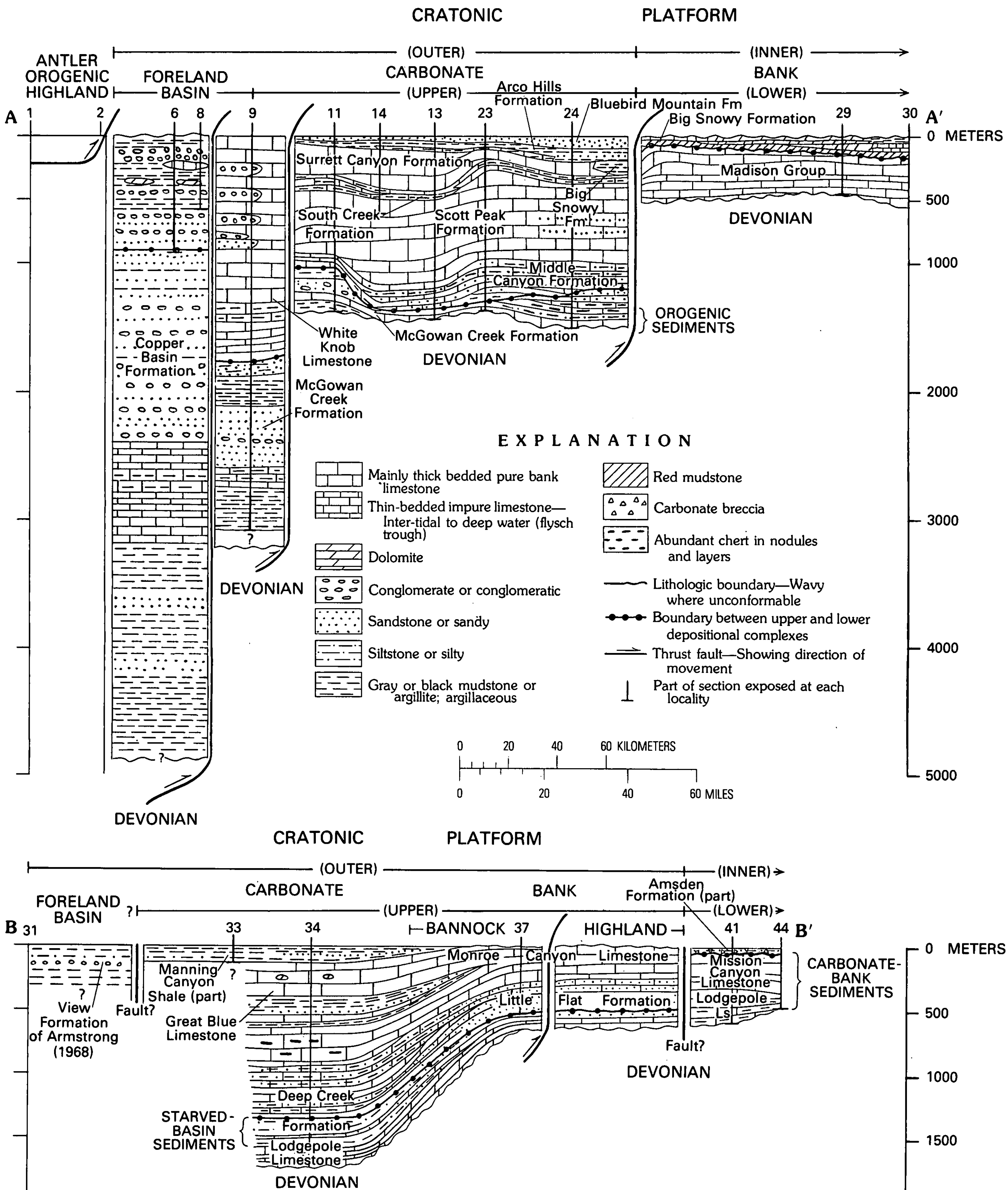
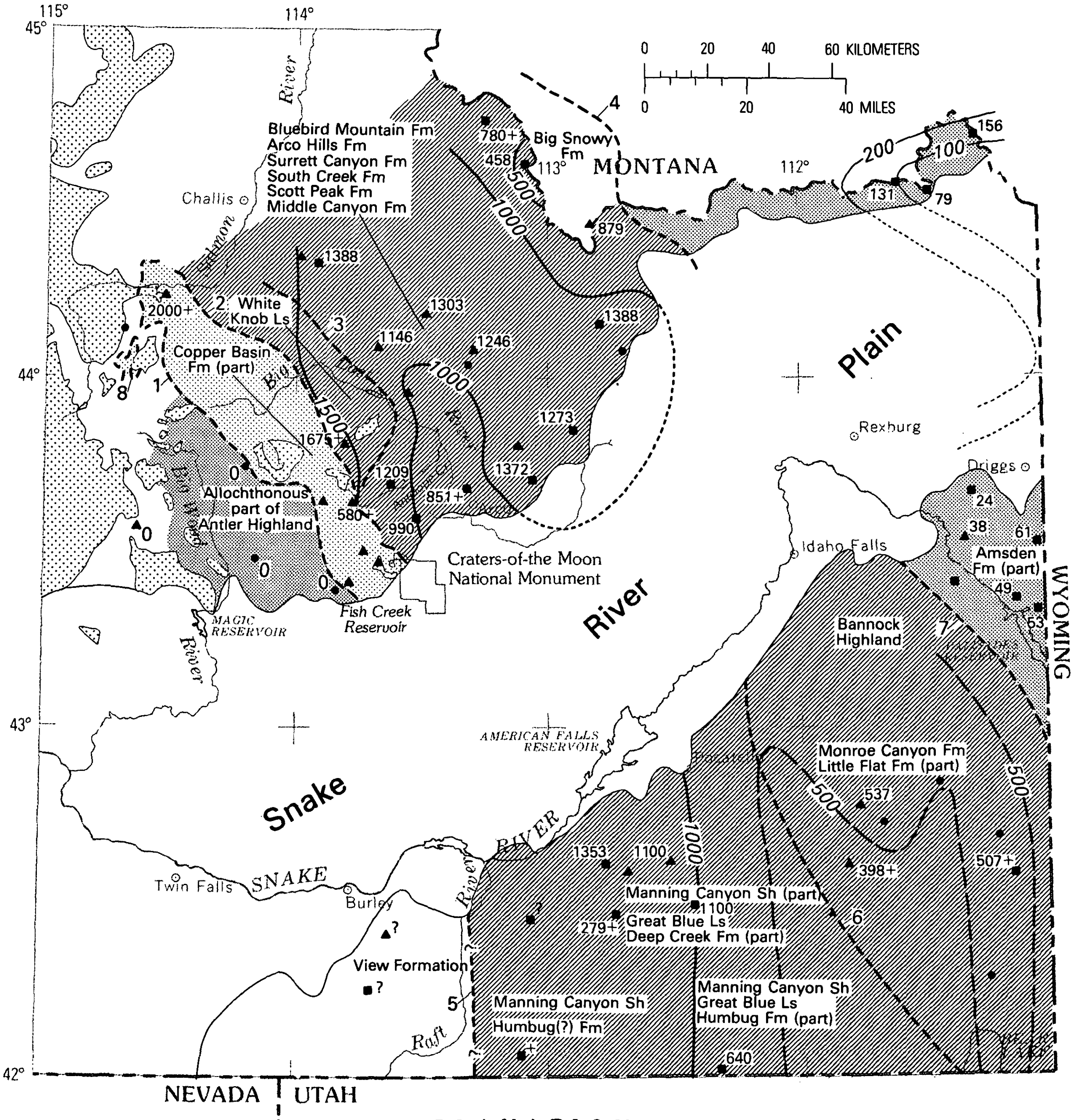


FIGURE 11.—Cross sections showing stratigraphic and structural relations of Mississippian rocks in Idaho north of the Snake River Plain (A–A') and south of the plain (B–B'). Locations of cross sections are shown on figures 2 and 4. Locations of stratigraphic sections are shown on figure 4. Datum is the top of the Mississippian.



EXPLANATION

Tertiary and Cretaceous intrusive rocks

Conglomerate, sandstone, argillite, pebbly mudstone, limestone and sandy dolomite—Much of it of shallow-water origin. Minor coal present. Some deep-water flysch sediments may be present in lower part

Shelf carbonate rocks locally interbedded with Antler-derived conglomerate and sandstone (White Knob Formation)

Craton-derived red and black claystone, limestone, dolomite and sandstone

Emergent area

——— Contact

—500— Isopachs (partly restored) in 500-m and 1000-m intervals—Dashed where inferred; dotted across Snake River Plain

—200— Isopachs (partly restored) in 100-m intervals—Dotted across Snake River Plain

– –5– – – Structural element or boundary described in caption

▲ ● ■640 Sample localities—Same as on figure 4. Thicknesses in meters quartzitic turbidites exposed along the Salmon River (Hobbs and others, 1975a; Nilsen, 1977). Strata exposed in the area outlined by structural element 8 (fig. 12) resemble the Salmon River turbidites (C. M. Tschanz, written commun., 1977). Locally abundant faunas, which include brachiopods, mollusks, bryozoans and calcareous foraminifers date parts of the sequences as late Meramecian to Chesterian.

The variety of lithologies in the upper part of the Copper Basin Formation suggests deposition in a shoaling foreland basin, which, before the end of Mississippian time, became a neutral area and, perhaps locally, an emergent-positive area.

WHITE KNOB LIMESTONE

The White Knob Limestone, at least 1,675 m thick at the type section in the White Knob Mountains, gradationally overlies the McGowan Creek Formation (Skipp, 1961a). It consists of fine-grained argillaceous and laminated silty spiculitic mudstone in the lower part. This mudstone grades upward into clean wackestone, grainstone, and packstone, which in the upper 840 m of the type section, are interbedded with lenses of conglomerate containing gravel fragments as much as 10 cm in diameter, medium- to coarse-grained sandstone, and minor siltstone beds (Skipp, 1961a). Fabrics of limestone, conglomerate, and sandstone in the upper part indicate deposition mostly in a subtidal to intertidal turbulent marine environment. The fine-grained lower beds are considered to be forebank deposits. The interbedded upper limestone-terrigenous clastic sequence of Chesterian age (Skipp and Mamet, 1970) is the westernmost representative of the prograding Upper Mississippian carbonate bank described by Rose (1976) north of the plain. Interbedded clastic rocks were derived from the Antler highland, though the mechanisms that produced this phenomenon remain obscure. Uplift of Copper Basin deposits to the west or north seems necessary for this interpretation. Mississippian-age thrusting has been postulated by Nilsen (1977). Shallow-water carbonate rocks of late Chesterian age are interbedded locally with terrigenous clastic rocks of the Copper Basin Formation.

Though much of the White Knob is recrystallized or altered by contact metamorphism, locally abundant faunas composed of bryozoans, mollusks, brachiopods, ostracodes, corals, trilobites, algae, foraminifers, crinoidal debris, and sparse fish plates are present in the upper part and were used to date the formation as middle Meramecian to late Chesterian (Skipp and Mamet, 1970; Yokley, 1974).

MIDDLE CANYON, SCOTT PEAK, SOUTH CREEK, AND SURRETT CANYON FORMATIONS

The carbonate-bank and forebank deposits of the Middle Canyon, Scott Peak, South Creek, and Surrett Canyon Formations are present from the White Knob Mountains east to the Idaho-Montana border and beyond. Thicknesses range from 1,288 m at localities 15 and 24, the northern Lost River Range and the southern Beaverhead Mountains, respectively, to 646 m in the Howe Peak area (locality 13) and about 390 m in the central Beaverhead Mountains (locality 28). These formations constitute a prograding carbonate-bank complex (Huh, 1967; Rose, 1976), which generally thickens westward (figs. 11, 12). The Middle Canyon, consisting of sandy, silty, and cherty, thin- to medium-bedded carbonate mudstone, is, in part, the forebank deposit formed in front of the prograding bank (Huh, 1967; Sandberg and Poole, 1977). The sandy

FIGURE 12.—Isopach map of Upper Mississippian rocks showing late Meramecian and Chesterian lithofacies and provincial formation names, and position of allochthonous part of Antler highland.

1. Structural western limit of well-dated Mississippian Copper Basin Formation, some of which is known to be of Late Mississippian age, and structural eastern limit of Wood River Formation.
2. Structural eastern limit of Copper Basin Formation and western limit of carbonate-bank complex.
3. Structural and/or depositional eastern limit of coarse detritus derived from Antler highland interbedded with shallow-water limestones (White Knob Limestone).
4. Eastern limit of Upper Mississippian carbonate-bank complex and structural boundary between the outer cratonic platform to the west and the inner cratonic platform to the east (craton margin of Sando (1976a), and upper carbonate-platform margin of Rose (1976)).
5. Structural western limit(?) of carbonate-bank complex.
6. Western limit of Bannock highland.
7. Structural boundary between outer cratonic platform and inner cratonic platform (craton margin of Sando (1976a) and upper carbonate-bank margin of Rose (1976)). Approximate eastern limit of Bannock highland.
8. Structural windows(?) of Copper Basin(?) Formation.

and silty parts of the Middle Canyon contain inner cratonic-platform-derived detritus deposited on the outer cratonic platform after Madison deposition; thus, they are correlative with the Kibbey Formation of the Big Snowy Group in Montana (Sando and others, 1975, fig. 13, pl. 11). Thick-bedded, variably cherty, fossiliferous bank limestones of the Scott Peak and Surrett Canyon Limestones, which form massive cliffs in several ranges, are separated by thin-bedded silty and argillaceous limestone of the South Creek Formation, which probably represents a period of relatively deep water marine circulation that temporarily interrupted carbonate-bank buildup. In the Beaverhead Mountains, the Scott Peak Formation contains sandy zones that are missing farther west. The Surrett Canyon Formation is thin or missing in the Beaverhead Mountains but is about 330 m thick in the northern Lost River Range (locality 15). Fossils are varied and abundant, particularly in the Scott Peak and Surrett Canyon bank limestones. They include corals, brachiopods, mollusks, bryozoans, algae, calcareous foraminifers, trilobites, ostracodes, and shark teeth, which indicate that the formations range in age from middle Meramecian to late Chesterian (Mamet and others, 1971). A valuable field guide to diagnostic rugose and colonial corals from these formations is available (Sando, 1976b).

BIG SNOWY, ARCO HILLS, AND BLUEBIRD MOUNTAIN FORMATIONS

In the Beaverhead Mountains, the South Creek and Surrett Canyon Formations are thin or absent and are replaced by black shale, siltstone, fine-grained calcareous sandstone, and phosphatic carbonate conglomerate of the Big Snowy Formation (fig. 8). These Big Snowy sediments mark the end of carbonate-bank buildup, the beginning of a restricted marine environment, and the westward spreading of inner cratonic-platform detritus onto the outer cratonic platform in late Meramecian to middle Chesterian time. Farther west, very fine grained silty limestone, shale, and siltstone of the Arco Hills Formation indicate the same restriction of circulation and shoaling of the seas near the site of the foundered flysch trough in late Chesterian time. The limestone, siltstone, and shale of the Big Snowy and Arco Hills Formations were succeeded by a very late Chesterian flood of very fine grained inner-craton-derived sand (the Bluebird Mountain Formation), which transgressed from east to west. The Bluebird Mountain Formation ranges in thickness from 99 m in the Beaverhead Mountains (locality 24) to about 5 m in the White Knob Mountains (locality 12). The Arco Hills Formation contains a fauna dominated by brachiopods (many phosphatic) and mollusks. Calcareous foraminifers, conodonts, and a few small rugose corals are present. The Big Snowy also contains numerous brachiopods, some mollusks, bryozoans (including *Archimedes* sp.), conodonts, algae, and calcareous foraminifers. The Bluebird Mountain Formation contains sparse foraminifers, conodonts, ostracodes, bryozoans, echinoderms, and brachiopods (Crawford, 1976; Skipp and others, 1979).

East of the inner cratonic platform margin, the Big Snowy in the Centennial Mountains (localities 29 and 30a) and Henrys Lake Mountains (locality 30) is less than 200 m thick and consists of sandstone, siltstone, red, green, and black claystone, thin-bedded limestone and thick-bedded, coarsely crystalline dolomite. The limestone contains corals, brachiopods, blastoids, bryozoans, foraminifers, and algae of late Chesterian age.

AMSDEN FORMATION (PART)

South of the plain, quartzite (Pampeyan and others, 1967; locality 44), sandstone, variegated shale, limestone (Staatz and Albee, 1966; locality 39), sandstone, siltstone, and shale (Jobin and Soister, 1964) have been assigned variously to the lower part of the Wells Formation, the Big Snowy Formation, or the lower part of the Amsden Formation. All these sequences, herein assigned to the Amsden Formation (fig. 8), are 61 m or less thick (fig. 12) and are of Late Mississippian (latest Meramecian and Chesterian) to Early Pennsylvanian age (Sando, 1976a). They compose the western extension of lagoonal and restricted marine facies of the cratonic platform of Wyoming (fig. 11) described by Sando in a companion paper (this volume). Faunas in the limestone beds consist of foraminifers, bryozoans, brachiopods, ostracodes, gastropods, and corals. The entire thickness of the Amsden Formation south of the plain is included on the isopach map for the Upper Mississippian (fig. 12), because it is not known how much of the interval might be Pennsylvanian. Most of the faunas reported are of Late Mississippian age, and it is concluded that only the uppermost part of the interval is of Pennsylvanian age.

DEEP CREEK, LITTLE FLAT, AND HUMBUG FORMATIONS (PART)

South of the plain and west of the inner craton margin, sandstone and sandy and cherty limestone of the upper member of the Deep Creek Formation, the upper three members of the Little Flat Formation, and an arbitrary upper part of the Humbug Formation were deposited on the outer cratonic platform. Included in this interval is a sequence of limestone, sandy limestone, and quartzite, more than 180 m thick, exposed in the Black Pine Range (locality 32; J. F. Smith, Jr., written commun., 1976). The sequence contains conodonts of probable middle Meramecian age (John Repetski, written commun., 1976) and is questionably assigned to the Humbug Formation. The interval is about 460 m thick in the Deep Creek Mountains (locality 34) and thins eastward to about 250 m in the Chesterfield Range (locality 37). Cratonic-platform-derived sandstone is dominant in the Chesterfield Range, whereas sandy limestone predominates in the Deep Creek Mountains. The sand and silt probably originated on the inner craton and were swept across the exposed and eroding Lower Mississippian carbonate complex (Sando and others, 1975; Rose, 1976); they are equivalent to the Darwin Sandstone Member of the Amsden Formation in Wyoming (Sando and others, 1975, fig. 13, pl. 11). A similar history is postulated for the sandy facies of the Middle Canyon Formation north of the plain. Fossils are sparse in these facies, but conodonts, foraminifers, brachiopods, and corals establish a middle Meramecian age for the sequences.

GREAT BLUE AND MONROE CANYON LIMESTONES

In the Deep Creek Mountains, above the Deep Creek Formation, about 760 m of Great Blue Limestone consists of thick- to thin-bedded carbonate-bank limestone, and a medial black to gray shale interval, 130 m thick. The Great Blue at this locality marks the westernmost outcrop of the upper carbonate shelf south of the plain (localities 34 and 35). The upper carbonate shelf margin was depicted by Rose (1976, fig. 9) and Poole and Sandberg (1977, fig. 7) as just east of the Deep Creek Mountains, whereas Sando (1976a, fig. 6) showed it west of this locality. Recent studies show that the Great Blue in the Deep Creek Mountains is probably part of the carbonate bank (L. D. Cress, unpub. data, 1978). Indirect evidence suggests to the senior author that the limestone of the Great Blue could extend as far west as the Black Pine Range (locality 32), where normally associated strata similar to the overlying Manning Canyon Shale and the upper part of the underlying Deep Creek or Humbug Formation are exposed in separate structural windows (J. F. Smith, Jr., written commun., 1976). To the east, thick- to thin-bedded limestone of the Monroe Canyon Limestone, more than 280 m thick, make up the eastward-thinning complex of shelf-carbonate deposits. Brachiopods, corals, bryozoans, echinoderms, gastropods, trilobites, foraminifers, and algae are abundant in the Monroe Canyon and Great Blue Limestones (Dutro and Sando, 1963; Sando, and others, 1969; Trimble and Carr, 1976).

MANNING CANYON SHALE (PART)

The Manning Canyon Shale is present west of boundary 6 and east of boundary 5 and is not present in the area of the Bannock highland (fig. 12). The formation consists of dark-gray or black to varicolored shale or argillite interbedded with calcareous siltstone and sandstone or quartzite, minor limestone, and oolitic phosphate rock. In the Deep Creek Mountains, 134 m of the lower part of the Manning Canyon is assigned to the Upper Mississippian complex. At Samaria Mountain, the entire interval is only 80 m thick and is assigned to the Mississippian. In the Black Pine Range, more than 600 m of black argillite containing a few lenses of quartzite and limestone has yielded a conodont that has an age range from latest Mississippian through Early Permian; the unit has been mapped as Manning Canyon Shale by J. F. Smith, Jr. (written commun., 1976). Part of the Black Pine sequence probably is Late Mississippian in age. The Manning Canyon Shale in southeasternmost Idaho has been interpreted as a wedge of fine terrigenous material probably derived from the margins of the inner cratonic platform (Rose, 1976; L. D. Cress, written commun., 1978), which spread westward across the upper depositional carbonate complex south of the plain, much as did the Arco Hills and Bluebird Mountain Formations north of the plain. Sando believes that the terrigenous material was derived from the Antler orogenic highland to the west. Black shale and phosphorite indicate a restricted marine environment in part. Fossils are rare, but brachiopods and conodonts of latest Chesterian age (Beus, 1968) have been identified in the lower part of the Manning Canyon Shale.

VIEW FORMATION OF ARMSTRONG (1968)

More than 100 m of dark, graphitic, muscovite-quartz-calcite schist and dolomitic quartz schist overlain by light-gray quartzite and scattered conglomerate lenses occurs in the Albion Range (locality 31) and is tentatively considered part of the Mississippian. Conglomerate in the sequence suggests correlation with the Chainman Shale or Diamond

Peak Formation, Antler highland-derived orogenic sequences present in northeastern Nevada. The sequence has been thrust eastward, and its original depositional site is unknown. The Raft River valley between the Albion and Sublett Ranges may contain the present boundary between the carbonate bank and the foreland basin (figs. 11, 12).

DEPOSITIONAL HISTORY

Regional uplift during latest early Meramecian time drained the inner cratonic platform and caused the sea to be confined to the outer cratonic platform and the foreland basin (Sando, 1976a). Lower Mississippian limestone banks of the Madison Group were exposed both north and south of the plain, and terrigenous detritus was carried westward into the foreland basin to form the sandy limestone of the Middle Canyon Formation north of the plain and the sandstone and sandy limestone of the upper parts of the Deep Creek, Little Flat, and Humbug Formations south of the plain. In middle Meramecian through late Chesterian time, thick-bedded fossiliferous bank limestones, representing several transgressive-regressive cycles, spread westward across the foreland basin (Huh, 1967; Rose, 1976). North of the plain, the bank limestone is interbedded with conglomeratic detritus derived from the shoaling distal flysch basin on the west (fig. 11). South of the plain, in Idaho, the location of the western edge of the carbonate bank is obscured by faulting, lack of outcrops, and metamorphism, but it may extend as far west as the Black Pine Range (fig. 11). In southeasternmost Idaho, the northern part of the Bannock highland was emergent in latest Chesterian time (Sando, 1976a, p. 332).

North of the plain, the inner cratonic platform remained emergent until latest Meramecian time. Black shale, sandstone, and limestone conglomerate (Big Snowy Formation) replaced carbonate-bank deposition in the western part of the inner cratonic platform during late Meramecian into Chesterian time. In latest Chesterian time, a blanket of mudstone and very fine grained sandstone spread westward from the inner platform at least as far west as the present eastern edge of the Copper Basin Formation. The flysch trough in middle Chesterian time was receiving fine- to coarse-grained western-derived detritus in a very shallow, locally restricted marine environment. As there is no evidence of western-derived detritus in the latest Chesterian part of the White Knob Limestone, we suggest that much of the area of the trough ceased to subside, became neutral, and then emergent, possibly before, or shortly after, the end of Mississippian time.

MISSISSIPPIAN-PENNSYLVANIAN CONTACT

The Mississippian-Pennsylvanian contact is gradational over much of the cratonic platform of Idaho, where Mississippian rocks are present. The systemic boundary lies within the Manning Canyon Shale south of the plain and is thought to be within the lowermost beds of the Snaky Canyon Formation near the contact with the underlying Bluebird Mountain Formation north of the plain. The base of the Bluebird Mountain Formation is a sharp boundary and relatively easy to find in the field, whereas the top is gradational and somewhat arbitrary. As the unit is sparsely fossiliferous and has a sharp basal contact, beds correlative with the formation have been considered lowermost Pennsylvanian (Mamet and others, 1971; Sando, 1976a). Calcareous Foraminifera of Mamet zone 19 (fig. 8), however, are present at several horizons within the formation (Skipp and others, 1979); the formation is assigned to the Mississippian in this paper.

Pennsylvanian rocks are absent in the outcrop area of the Copper Basin highland, which was a source area in Middle Pennsylvanian time (Skipp and Hall, 1975a; Skipp and Hait, 1977). The Mississippian-Pennsylvanian boundary is unconformable in the area of the Bannock highland in southeastern Idaho (figs. 13, 14; Williams, 1962; Sando, 1976a) and in the area of Big Snowy deposition on the inner cratonic platform at the Montana-Idaho border. The systemic boundary appears conformable south of the plain near the Wyoming border, where it is within the Amsden Formation (Sando, 1976a).

PENNSYLVANIAN SYSTEM

The Pennsylvanian System of the United States correlates with most of the Upper Carboniferous Series in Europe (fig. 8) and includes, from oldest to youngest, the Morrowan, Atokan, Des Moinesian, Missourian, and Virgilian Provincial Series.

PALEOTECTONIC SETTING—HUMBOLDT OROGENY

Pennsylvanian rocks in Idaho can best be grouped into two depositional frameworks. The early one, representing Early and early Middle Pennsylvanian (Morrowan and Atokan) time reflects the paleotectonic setting inherited from latest Mississippian time. Limestones, many sandy, silty or argillaceous, were deposited in shallow epeiric seas, which spread uniformly across the cratonic platform westward to the edge of the neutral or emergent Copper Basin area, the site of the former Mississippian flysch trough. The seas also flooded the inner cratonic platform along the eastern edge of the State, and both carbonate and terrigenous sediments accumulated in restricted marine and lagoonal environments.

In late Middle Pennsylvanian (Des Moinesian) time, the tectonic setting changed. The Copper Basin highland (figs. 13, 14) rose and contributed minor coarse detritus to a newly formed Wood River basin on the west, to the Snaky Canyon Formation area on the east, and to the Oquirrh Formation on the south. The Bannock highland in southeastern Idaho, which had been emergent through Late Mississippian and Early Pennsylvanian time (Williams, 1962; Sando, 1976a), submerged and received sediment. At the same time, floods of very fine to fine-grained sand were deposited in the Wood River, Sublett, Oquirrh, and Quadrant basins.

Ketner (1977) recently defined the Humboldt orogeny in Nevada to include the late Paleozoic orogenic events that affected parts of the Great Basin from late Middle Pennsylvanian time through Early Permian time. The axis of the north-trending Humboldt highland of western Nevada, which shed detritus both east and west, was west of the axis of the former Antler highland, according to Ketner (1977). Poole and Wardlaw (1978) described the highland as a series of Pennsylvanian flysch basins and emergent ridges or islands in central Nevada. They considered the highland to have been along the western side of the former Antler orogenic highland.

Extension of the Humboldt highland into southern Idaho is tenuous because the Idaho batholith occupies the probable position of this inferred orogenic belt. Paleocurrent evidence of a western or southwestern source terrane for the bulk of Wood River Formation detritus (Thomasson, 1959) is compatible with the idea of a highland in western Idaho. The major problems with this interpretation are (1) the mineralogy of the Wood River sandstones, which suggest a cratonic basement source (Hall and others, 1974), and (2) the similarity between Oquirrh basin and Wood River basin sedimentary rocks, even though the Oquirrh basin formed earlier. The Oquirrh basin probably also was filled with detritus derived from cratonic sources (Bissell, 1974).

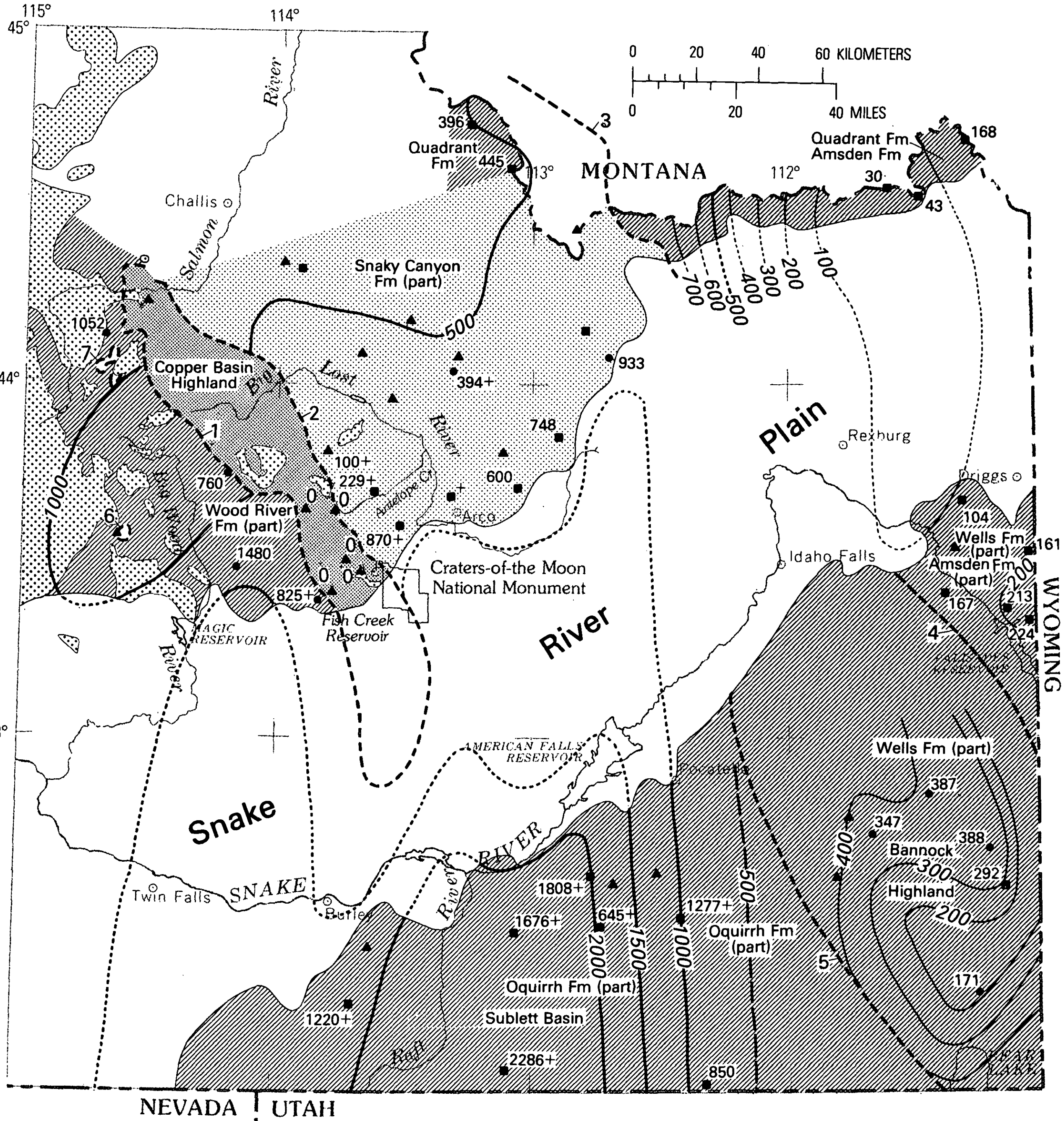
North of the plain, throughout Pennsylvanian time, the Copper Basin highland remained an effective barrier between the Wood River basin on the west and the depositional area of the Snaky Canyon Formation on the east. South of the plain, no such barrier existed; instead, an area of subsidence and basin filling, referred to as the Sublett basin (Cramer, 1971; this report figs. 13, 14), formed as a northern extension of the Oquirrh basin during Pennsylvanian and Permian time. Minor granule and pebble conglomerate in the Pennsylvanian part of the Oquirrh Formation of the Black Pine and Sublett Ranges, and the Deep Creek Mountains (J. F. Smith, Jr., written commun., 1976; R. L. Armstrong, written commun., 1976; Trimble and Carr, 1976), probably was derived from the Copper Basin highland to the north.

The depositional framework established during Middle Pennsylvanian time persisted into the Permian.

UNDERLYING ROCKS

North of the plain, upper Middle Pennsylvanian rocks rest in thrust contact on the Devonian Milligen Formation in the Wood River basin (fig. 8; Sandberg and others, 1975), but the Wood River is interpreted as having been deposited unconformably on the Milligen. The Milligen Formation is part of the complex that constituted the Late Devonian-Early Mississippian Antler orogenic highland (Poole, 1974; Sandberg and others, 1975). East of the Copper Basin highland, the Pennsylvanian-Mississippian contact is gradational, and its position in the stratigraphic sections is defined on the basis of contained faunas. East of the inner cratonic platform margin, the Amsden Formation unconformably overlies Upper Mississippian rocks in the vicinity of the Centennial Mountains and Henrys Lake Mountains (localities 29, 30, and 30a).

South of the plain, the Mississippian-Pennsylvanian contact is considered to be conformable within the Amsden Formation where the Amsden is present. Elsewhere, except for the area of the Bannock highland, where Middle Pennsylvanian rocks rest on Upper Mississippian strata, the lithologic transition from the Mississippian to Pennsylvanian Systems is within the Manning Canyon Shale. Through much of southern Idaho, however, the transition is not actually observed, because uppermost Mississippian shale formed a zone of structural weakness along which Mesozoic thrusting took place (S. S. Oriel and L. B. Platt, unpub. data, 1978). Mississippian rocks are exposed only in structural windows in the Black Pine and Sublett Ranges (figs. 2, 4; J. F. Smith, Jr., M. H. Hait, Jr., and R. L. Armstrong, unpub. data).



EXPLANATION

Tertiary and Cretaceous intrusive rocks

Mostly cherty or sandy limestone, minor sandstone or quartzite

Mostly calcareous or quartzitic sandstone, with sandy limestone and minor granule- to cobble- sized conglomerate and shaly mudstone in places

Emergent area

—500— Isopachs (partly restored) in 500-m and 1000-m intervals—Dashed where inferred; dotted across Snake River Plain

—100— Isopachs (partly restored) in 100-m intervals—Dashed where inferred; dotted across Snake River Plain

--1-- Structural element or boundary described in caption

■ ▲ ●1052 Sample localities—Same as on figure 4. Thicknesses in meters

OVERLYING ROCKS

North of the Snake River Plain and west of structural element 3 (fig. 13), rocks of Pennsylvanian age are overlain gradationally by Permian strata except in the area of the Copper Basin highland where neither Pennsylvanian nor Permian rocks have been recognized. The Pennsylvanian-Permian boundary lies within the Wood River and Snaky Canyon Formations. Because of the gradational nature of the contact, some Permian rocks are shown as combined with Pennsylvanian rocks in figure 2. In the central Beaverhead Mountains (localities 27 and 28), sandstone and limestone assigned to the Quadrant Formation correlate with the lower part of the Snaky Canyon Formation and are overlain gradationally by carbonate rocks assigned to the Grandeur Member of the Park City Formation (Ruppel, 1968), which probably correlates with the upper part of the Snaky Canyon Formation. The Pennsylvanian-Permian boundary may be within the lower beds of the Park City Formation as defined in these areas, though no faunal data are available at this time. Only beds assigned to the Quadrant Formation have been included on the isopach map (fig. 13).

The Quadrant Formation, which crops out in Idaho east of structural element 3, is no younger than Middle Pennsylvanian in age on the basis of fusulinids obtained from the nearby type section (Thompson and Scott, 1941); it is unconformably overlain by the Phosphoria Formation of Permian age (fig. 8).

South of the plain, the Pennsylvanian-Permian boundary is gradational in all areas of outcrop and is present within both the Oquirrh and Wells Formations.

THICKNESS

Pennsylvanian rocks generally thicken to the south and west across Idaho, a very different trend from that of thicknesses within the Mississippian System. The thickest sequences (Oquirrh Formation), more than 2,286 m, are present in the Sublett and Black Pine Ranges—a northern extension of the Oquirrh basin of Utah. The oldest upper Paleozoic rocks found to date in the Cassia Mountains (fig. 4), west of the Albion Range, are of Permian age (Morgan, 1977). The strata so resemble the upper part of the Wood River Formation north of the plain that Pennsylvanian rocks are assumed to be present in the subsurface.

Observed thicknesses in the Wood River area range from 760 to 1,480 m; thinnest sequences are in the center of the area of outcrop (W. E. Hall, unpub. data; fig. 13). No Pennsylvanian rocks have been reported from the area of the Copper Basin highland. East of the highland, the Pennsylvanian part of the Snaky Canyon Formation ranges from 600 to about 1,100 m in thickness. Strata assigned to the Quadrant Formation in the Beaverhead Mountains (localities 27 and 28) are about 400 m thick. East of the outer cratonic platform, Pennsylvanian rocks thin to as little as 30 m (locality 29).

South of the plain and east of the Sublett basin, on the Bannock highland (fig. 13), the thickest sequences assigned to the Wells Formation, 388 m and 387 m, are in the Preuss and Wooley Ranges (Peale Mountains) (localities 49 and 45) of southeasternmost Idaho. In the area of Amsden Formation south of the plain, a maximum of 224 m of Wells Formation is reported in the Snake River Range (locality 42).

FORMATIONS INCLUDED

Rocks of Early to latest Pennsylvanian age are included between line I and rocks above the Pennsylvanian System at the top of the correlation chart (fig. 8). Cratonic-platform basin and restricted-marine or lagoonal depositional environments are represented.

North of the plain, the lower parts of the Wood River and Snaky Canyon Formations make up the interval in most areas. Rocks assigned to the

FIGURE 13.—Total isopach map of Pennsylvanian rocks showing Middle and Upper Pennsylvanian lithofacies, provincial formation names, location of Copper Basin highland, and position of Lower Pennsylvanian Bannock highland.

1. Structural eastern limit of Wood River Formation and western edge of Copper Basin highland.
2. Structural eastern edge of Copper Basin highland and western limit of Snaky Canyon Formation.
3. Structural eastern limit of Snaky Canyon Formation and western limit of Amsden Formation.
4. Structural western limit of Amsden Formation and eastern border of Bannock highland (modified from Williams, 1962, and Sando, 1976a).
5. Western limit of Wells Formation and approximate western border of Bannock highland.
6. Structural window of Mississippian(?) rocks.
7. Structural windows(?) of Mississippian(?) rocks.

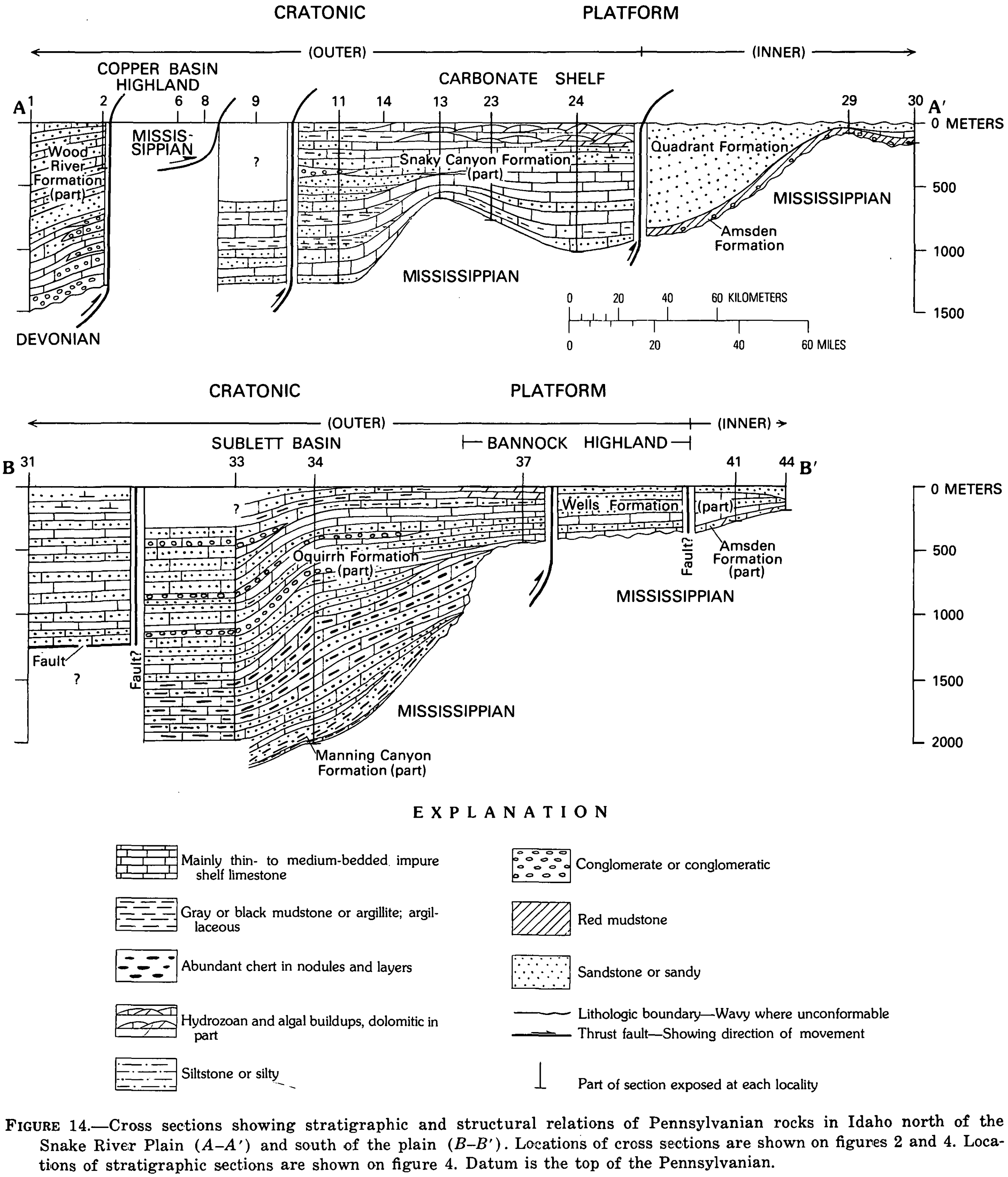


FIGURE 14.—Cross sections showing stratigraphic and structural relations of Pennsylvanian rocks in Idaho north of the Snake River Plain (*A–A′*) and south of the plain (*B–B′*). Locations of cross sections are shown on figures 2 and 4. Locations of stratigraphic sections are shown on figure 4. Datum is the top of the Pennsylvanian.

Quadrant Formation in the Beaverhead Mountains (localities 27 and 28) are correlatives of the Snaky Canyon Formation. The Amsden and Quadrant Formations constitute the Pennsylvanian System east of structural element 3 (fig. 13) on the cratonic platform.

South of the plain, the upper part of the Amsden Formation and the lower part of the Wells Formation constitute the Pennsylvanian System in the eastern part of the outcrop area. To the west, Pennsylvanian strata are represented in the upper part of the Manning Canyon Shale and the lower part of the Oquirrh Formation as far west as the Black Pine Range (locality 32).

WOOD RIVER FORMATION (PART)

The Pennsylvanian part of the Wood River Formation includes the Hailey Conglomerate Member (unit 1) at the base, overlying units 2, 3, 4, 5, and slightly less than half of unit 6 (Hall and others, 1974; fig. 8). This part of the formation is 1,480 m thick at the type section (locality 1) and 760 m thick at Wilson Creek to the north (locality 1a). The base of the Hailey Conglomerate Member is the sole of a major thrust; it is strongly sheared and detached into boudins where it is present (fig. 14). The Wood River rests on the Devonian Milligen Formation through much of its outcrop area and overrides the Mississippian Copper Basin Formation and a sequence of Ordovician through Devonian rocks (Dover and others, 1976) along its eastern outcrop margin (structural element 1, fig. 13). The Wood River Formation may rest on the Copper Basin in the White Cloud Peaks area (structural element 7, fig. 13). The Pennsylvanian part of unit 6 appears to be overlain conformably by beds of Wolfcampian (Early Permian) age.

The Hailey Conglomerate Member (unit 1), which consists of chert- and quartzite-pebble to -cobble conglomerate and quartzite interbedded with brown, partly algal, limestone, is about 120 m thick at locality 1 (fig. 4) and in its type section near the town of Hailey. Faulting and folding in the type locality has duplicated the section, which was reported by Thomasson (1959) to be 986 ft (300 m) thick. Though no studies of clast-size distribution or sedimentary structures within the sheared basal conglomerate have been made, it now seems likely that the chert and quartzite clasts were derived from the emergent eastern Copper Basin highland (fig. 13). The basal conglomerate is overlain by more than 1,000 m of interbedded sandstone, quartzite, sandy and argillaceous limestone, and minor easterly derived turbidite conglomerate (Hall and others, 1974; Skipp and Hall, 1975a; Skipp, 1975).

The bulk of the formation consists of very fine to fine-grained, slightly feldspathic, quartzose limy sandstone and sandy limestone that had a source different from that of the conglomerate. The craton (Hall and others, 1974; Skipp and Hall, 1975a) and the Humboldt highland (Ketner, 1977) have been suggested as possible source terranes. South to southwest and a few northwest current directions were measured in the Wood River Formation by Thomasson (1959), who also noted that sand in the Slate Creek section (locality 17) is coarser, fine to medium grained, and more abundant than in the type section (locality 1) to the south.

The Pennsylvanian part of the formation ranges in age from Middle (Des Moinesian, possibly Atokan) through Late (Virgilian) (Bostwick, 1955; Hall and others, 1974). No Missourian faunas have been recovered from the Wood River Formation. About 61 m of strata occur between fusulinid faunas of Des Moinesian and Virgilian age in the type section, but no fault or unconformity has been recognized in the poorly exposed interval (Hall and others, 1974). Whether the Missourian stage is condensed in this interval or is missing altogether is not known.

Pennsylvanian rocks south of the plain informally have been called Wood River Formation because of their lithologic resemblance to units 2 through 6. These rocks are assigned to the Oquirrh Formation. Permian rocks in the Cassia Mountains (fig. 4) bear a strong resemblance to the Permian part of the Wood River Formation (Morgan, 1977), and it seems obvious that Pennsylvanian and Permian sedimentation took place between these areas now separated by the Snake River Plain.

The Wood River Formation is distinguished by its locally abundant fusulinid faunas (Bostwick, 1955; Hall and others, 1974), particularly in unit 6, but other faunal and floral groups are present, including crinoidal debris, bryozoans, brachiopods, corals, nonfusulinid foraminifers, and algae.

SNAKY CANYON FORMATION (PART)

East of the Copper Basin highland and north of the Snake River plain the Pennsylvanian System is represented by the lower part of the recently defined Snaky Canyon Formation (Skipp and others, 1979 which includes the basal Bloom Member overlain by the Gallagher Peak Sandstone Member and the lower beds of the Juniper Gulch Member. At the type section in the southern Beaverhead Mountains (localities 24 and 25), the Pennsylvanian part of the formation is about 900 to 1,000 m thick. The formation is more than 870 m thick at Timbered Dome (locality 11) and as little as 600 m thick in the Arco Hills-Howe Peak area (locality 13).

The Bloom Member in the Lost River and Lemhi Ranges and the Beaverhead Mountains consists of medium-bedded gray limestone, much of it sandy or silty, interbedded with thin beds of very fine grained, yellowish-brown-weathering sandstone and siltstone. Nodules of incipient chert (rims of brown-weathering silicified material enclosing gray limestone), as well as concentrically laminated stromatolitic mounds, are common. Impure wackestones, packstones, and lesser amounts of calcareous mudstone are the chief carbonate rocks. Fossils are fairly common and include small fusulinids and other foraminifers, brachiopods, corals, bryozoans, crinoid columnals, and abundant encrinitic debris. The base of the Bloom Member east of the White Knob Mountains is probably near the Late Mississippian-Early Pennsylvanian boundary, and the top is probably Missourian in age (fig. 8).

The Gallagher Peak Sandstone Member consists largely of very fine grained calcareous sandstone. It is less than 60 m thick, appears to thin westward, and has not been recognized with certainty in the White Knob Mountains. The member is of probable Missourian age.

The lower part of the Juniper Gulch Member consists of interbedded sandy and cherty, generally light-gray-weathering, thin- to thick-bedded limestone and dolomite. The sand and chert is mainly in the basal 100 m. The Pennsylvanian-Permian boundary is near the middle of the member in hydrozoan(?)-algal carbonate buildups, 100 to more than 200 m thick. The buildups were described in detail by Breuninger (1976), who suggested that lenticular buildups stood topographically above the adjacent sea floor while they formed (fig. 14). Other fauna associated with the buildups include common encrusting foraminifers and bryozoans and less common gastropods, crinoids, and brachiopods. Fusulinid foraminifers are sparse, although they have been found throughout the lower part of the member and date it as mostly of Missourian and Virgilian age (Shannon, 1961; Breuninger, 1976; R. C. Douglass, written commun., 1976; fig. 8). The Juniper Gulch Member also has not been recognized in the White Knob Mountains area.

In the White Knob Mountains (localities 11 and 12), the Snaky Canyon Formation consists of basal thick-bedded sandy limestone overlain by interbedded thin- to medium-bedded, fossiliferous, argillaceous and silty limestone, calcareous shale, siltstone, and very fine grained sandstone. The limestone is mostly impure carbonate mudstone and wackestone containing abundant calcareous sponge spicules and lesser amounts of bryozoan, molluscan, and brachiopod debris. Minor sandy limestone, together with quartz and chert sand of medium grain size, and granule limestone conglomerate are present locally. At Timbered Dome (locality 11), a chert-quartzite pebble- to cobble-conglomerate bed, 14 m thick containing fragments as large as 10 cm in diameter, crops out about 700 m above the base. The conglomerate, which was derived from the Copper Basin highland, is assigned to the Middle Pennsylvanian (fig. 8), on the basis of silicified brachiopods identified by J. T. Dutro, Jr. (written commun., 1978), from beds 15 m below.

Strata that correlate with the Snaky Canyon Formation have been called Quadrant Formation in the central Beaverhead Mountains (localities 27 and 28). These Quadrant sections contain about 50 percent (locality 28) or more (locality 27) quartzitic and calcareous sandstone (fig. 13). The increase in percentage of sand to the north probably indicates a northern or northeastern source for the detritus (Lucchitta, 1966). The sandstone is fine to very fine grained, clean, and thick bedded and is interbedded with locally sandy limestone and dolomite and minor black shale and mudstone (Lucchitta, 1966; Ruppel, 1968). The carbonate rocks contain nodules of incipient chert and stromatolitic mounds resembling those described in the Snaky Canyon Formation, and numerous fragments of brachiopods, bryozoans, corals, crinoids, and a few small fusulinids. The sequences are poorly dated but are thought to represent most of the Pennsylvanian System.

AMSDEN (PART) AND QUADRANT FORMATIONS

East of structural element 3, and north of the Snake River Plain, the Amsden and Quadrant Formations constitute the Pennsylvanian System. In the Henrys Lake Mountains (locality 30), the Amsden Formation is 84 m thick and consists of a basal limestone conglomerate overlain by red claystone, dark-gray limestone, banded red chert, dolomite breccias, sandy dolomite, and dolomitic sandstone at the top (E. K. Maughan, written commun., 1977). The limestones contain bryozoans, echinoderms, mollusks, conodonts, hydrozoans, and calcareous foraminifers. Lithologies of the Amsden represent a shallow-marine environment. Terrigenous detritus in the Amsden may have been derived from local uplifted areas of the Big Snowy Formation within the Montana-Dakota region (Maughan, 1975).

The Amsden Formation is overlain by the Quadrant Formation, which is dominated by thick-bedded fine-grained quartzose sandstone, locally quartzitic; thin lenticular sandy dolomite beds are interleaved in the basal and upper parts. The Quadrant is believed to overlie the Amsden Formation conformably in the Henrys Lake Mountains (E. K. Maughan, written commun., 1976). The Quadrant is thin, 84 m (locality 30) and 24 m (locality 29) thick in easternmost Idaho, but is known to be about 800 m thick in southwestern Montana, just north of the Idaho border and east of structural element 3 shown in figure 13 (Sloss and Moritz, 1951; Scholten and others, 1955). Emergent land sources to the north-northwest and northeast have been suggested for the multicycle deposits of the Quadrant Formation (Mallory, 1967; Maughan, 1975). Fusulinids from the type section in Wyoming, just northeast of the northeast corner of Idaho, indicate a Des Moinesian and possibly Atokan, age for the formation (Thompson and Scott, 1941; fig. 8).

RANCHESTER LIMESTONE MEMBER OF AMSDEN FORMATION

The upper part of the Amsden Formation, which is probably of Early Pennsylvanian age, is included in the discussion of Upper Mississippian rocks south of the plain. The Amsden is recognized east of structural element 5 (fig. 13), where it is composed largely of the Ranchester Limestone Member (Sando and others, 1975; Sando, this volume), which probably is less than 30 m thick. The Ranchester is a sequence of interbedded cherty dolomite and limestone, sandstone, and shale.

WELLS FORMATION (PART)

South of the plain, the lower part of the Wells Formation makes up approximately the eastern half of the outcrop area of Pennsylvanian rocks in Idaho. The Wells Formation was named by Richards and Mansfield (1912) for exposures in Wells Canyon, which is 2 miles south of locality 46 (fig. 4). The Wells is divided into two members, a lower and an upper. Fusulinids from sections near locality 45 in the Wooley Range of the Peale Mountains, date the lower member of the Wells as Middle Pennsylvanian (Atokan and Des Moinesian), and the lower part of the upper member as Des Moinesian or younger. Permian fusulinids have been collected 300–400 ft (100–120 m) above the base of the upper member (Cressman, 1964).

No faunas of Morrowan age have been recovered from these sequences, and, on this basis, an Early Pennsylvanian high or neutral area (the Bannock highland) was proposed by Williams (1962); this area is outlined in modified form on figure 13 (structural element 5).

At present, there is no record of Late Pennsylvanian faunas from the Wells Formation, and several authors have suggested an hiatus at this level (Mallory, 1975; Rich, 1977). A Late Pennsylvanian fauna was reported in the Tensleep Sandstone in central westernmost Wyoming (Love, 1954; Wanless and others, 1955) from sandstone tentatively correlated with beds in the lower part of the upper member of the Wells Formation (Cressman, 1964). This correlation is used in this report in the absence of new data (fig. 8).

The formation ranges in thickness from 104 m (locality 39) in the east to about 405 m (locality 37a) in western exposures.

In the type area, the lower member of the Wells Formation consists of a basal very fine grained to fine-grained quartzose sandstone bed, 30–100 m thick, overlain by medium gray sandy and minor oolitic limestone, interbedded with relatively thin, fine-grained sandstone beds as much as 3 m thick. Chert nodules and layers occur throughout. Phosphatic rocks and a thin black shale are present locally at the top of the member (Cressman, 1964). The lower part of the upper member consists of yellowish-brown and light-gray thick-bedded, very fine to fine-grained calcareous quartzose sandstone containing thin (2–5 m) beds of gray limestone, calcareous sandstone, and dolomite (Cressman, 1964; Montgomery and Cheney, 1967).

In the Snake River Range (locality 44) and Big Hole Mountains (locality 39), the Wells consists of a lower thin to irregularly bedded brownish-gray cherty limestone, minor limestone breccia, and thin sandstone interbeds overlain by fine-grained well-sorted quartzite.

On the basis of published descriptions, the total amount of sandstone and/or quartzite in the Pennsylvanian part of the Wells Formation is extremely variable—50 to 75 percent at localities 45, 46, and 49 near the type section and less than 20 percent at locality 37a in the Chesterfield Range and locality 41 in the Snake River Range. At localities 42, 44, and 47 in the Snake River Range and Caribou Range, the amount is close to 50 percent. An average of about 50 percent is used on figure 13.

Sparse fossils in the Wells include crinoid plates, brachiopods, corals, fusulinids, and other foraminifers.

MANNING CANYON SHALE (PART)

An interval of quartzite and sandstone, limestone, siltstone, and shaly argillite, 220 m thick, assigned to the upper part of the Manning Canyon Shale in the northern Deep Creek Mountains (locality 34), is included in the Pennsylvanian part of the Manning Canyon Shale. A similar thickness (186 m) is present in the southwestern part of the Bannock Range (locality 48). Elsewhere south of the plain, the incomplete Manning Canyon is exposed in structural windows in the southern Deep Creek Mountains (locality 35), the Black Pine Range (locality 32), and the Sublett Range (locality 33), and neither thicknesses nor age determinations are available.

OQUIRRH FORMATION (PART)

Strata assigned to the lower part of the Oquirrh Formation occur at Samaria Mountain (locality 36), in the Deep Creek Mountains (locality 34), and into the Albion Range (locality 31). They also are presumed to be present in the subsurface as far west as the Cassia Mountains (fig. 4; see discussion of Wood River Formation). Thicknesses within the Sublett basin (fig. 13) range from about 850 m at Samaria Mountain to more than 2,286 m in the Black Pine Range (locality 32).

At Samaria Mountain, the limestone member (= West Canyon Limestone Member of Beus, 1968) and the lower 400 m of the overlying sandy member are included within the Pennsylvanian part of the Oquirrh Formation (Platt, 1977). The limestone member consists of interbedded brown to gray quartz sandstone, gray calcareous sandstone, and limestone. The sandy member is made up of sandstone and minor chert-granule to cobble-conglomerate beds in the Upper Pennsylvanian part of the sequence. The sequence is composed of more than 50 percent sandstone. Scattered fusulinid faunas indicate that Lower through Upper Pennsylvanian beds representing the five stages are present (Beus, 1968; Platt, 1977).

In the northern Deep Creek Mountains (locality 34), a nearly complete sequence of the lower part of the Oquirrh Formation (1,710 m thick) is assigned to the Pennsylvanian (fig. 15; Trimble and Carr, 1976). Units A, B, C, and the lower part of D consist of sandy or silty limestone and minor quartzite, bedded chert, and chert-granule to pebble conglomerate representing late Atokan through Virgilian time. The base of the sequence is faulted. Conglomerates occur in Middle and Upper Pennsylvanian beds. Bedded chert is restricted to the Upper Pennsylvanian part. Sandstone and quartzite make up a minor part of the sequence. Fusulinids, corals, brachiopods, gastropods, and bryozoans locally are abundant in the formation.

A thick (more than 1,676 m) incomplete section of the Oquirrh Formation in the northern Sublett Range (locality 33; R. L. Armstrong, written commun., 1977) has been assigned to the Pennsylvanian System. The lower part of the sequence consists of interbedded silty, sandy, or cherty limestone and calcareous sandstone, overlain by quartzite and calcareous sandstone, sandy and cherty limestone, and thin interbeds of chert-granule conglomerate. Chert-granule conglomerate is more common in the northern part of the range than in the central part, suggesting a northern source for the detritus, such as the Copper Basin highland. Fusulinids are abundant locally.

The thickest Pennsylvanian Oquirrh sequence (more than 2,220 m thick) has been reported from the Black Pine Range (locality 32). Limestone and sandy limestone, sandstone or quartzite, sandstone breccia, and local limestone and chert-pebble conglomerate are present in many fault slices (J. F. Smith, Jr., written commun., 1976). Scattered fusulinids, corals, and mollusks date the incomplete sequence.

In the Albion Range (locality 31), an allochthonous sheet of unmetamorphosed sandy limestone and calcareous sandstone assigned to the Oquirrh Formation has been thrust over Paleozoic metasedimentary rocks (Armstrong, 1968). The limestone and sandstone, which are several thousand feet thick, resemble Oquirrh strata in the Sublett Range and are, in part, of Pennsylvanian age.

DEPOSITIONAL HISTORY

During Pennsylvanian time, sandy carbonate rocks together with minor conglomerates were deposited in several cratonic basins across the area of outcrop in southeastern Idaho. In Early and early Middle Pennsylvanian time, shallow-water carbonate deposits and sandstones were laid down on the carbonate shelf east of the site of the Mississippian flysch trough north of the plain and east of the inner cratonic platform margin. The inner cratonic platform may have been emergent part of the time and then was inundated slowly by poorly circulating marine waters by the end of Morrowan time. South of the plain, the low-lying Bannock highland separated the shallow offshore marine depositional site of the Ranchester Limestone Member of the Amsden Formation on the east from the restricted-marine en-

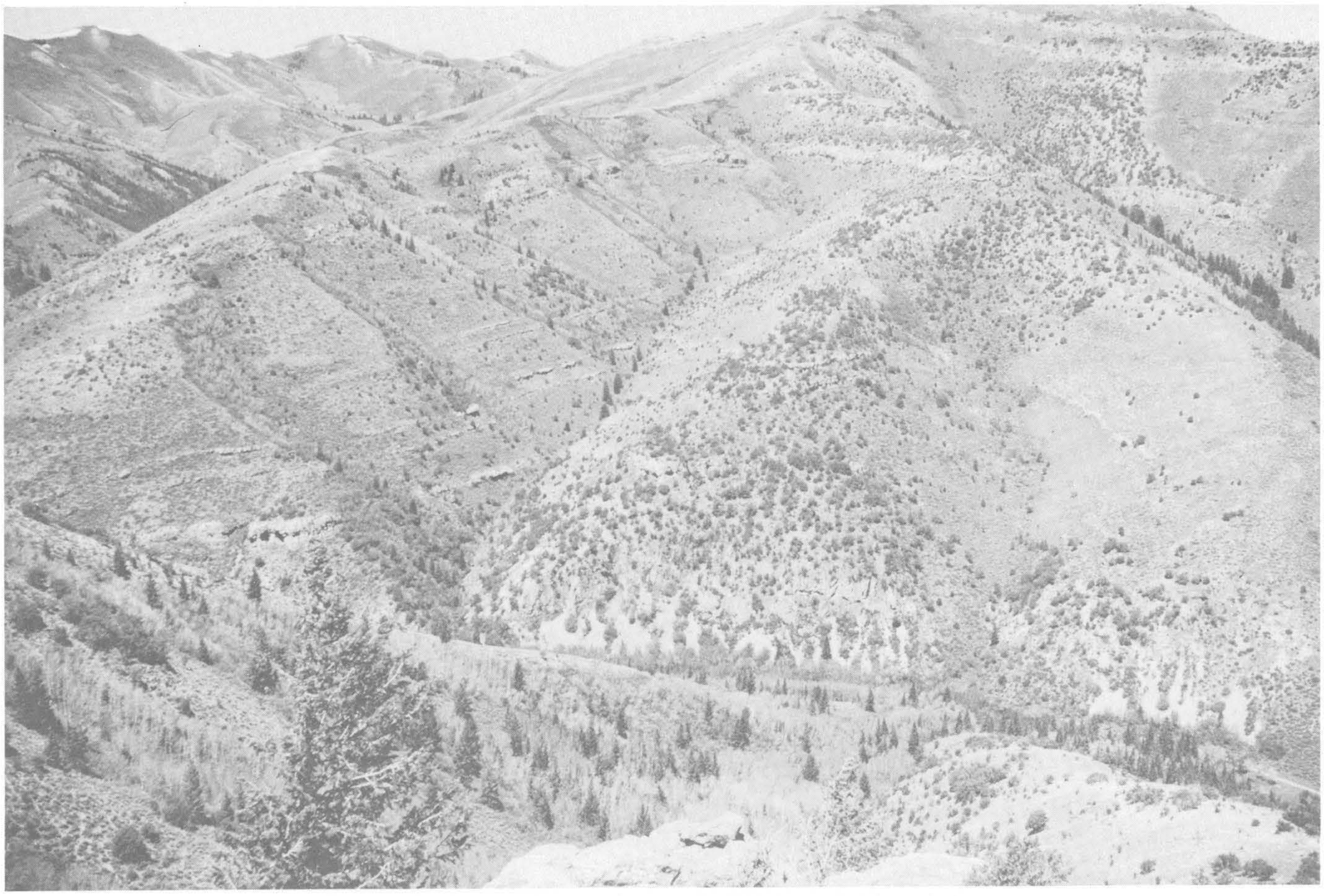


FIGURE 15.—Anticline in Pennsylvanian part of Oquirrh Formation in northern Deep Creek Mountains near locality 34 (fig. 4). Base of unit C is at base of conspicuous ledge that crosses upper part of draw in central part of picture. From Trimble and Carr (1976, p. 31, fig. 8).

vironment of deposition of the upper part of the Manning Canyon Shale. Both sequences contain sand derived principally from a cratonic source (L. D. Cress, unpub. data, 1978).

In late Atokan and/or Des Moinesian time, the tectonic setting that had been inherited from the Mississippian System changed drastically. North of the plain, Wood River sediments began to accumulate on the former site of the Antler orogenic highland. Sediments of the Mississippian flysch trough became emergent, probably because of faulting, and provided chert- and quartz-pebble and granule detritus to areas west, east, and south (Skipp, 1975; Skipp and others, 1979. At the same time, large volumes of fine-grained quartz sand flooded the basins north and south of the plain. Less sand reached the area of accumulation of the Snaky Canyon Formation, in which intermittently emergent carbonate buildups, composed of hydrozoans and algae, formed.

South of the plain, in late Atokan or Des Moinesian time, the Bannock highland was completely submerged, and shallow-water-marine sandy carbonate, calcareous sandstone, and noncalcareous sandstone were deposited in the rapidly subsiding Sublett basin, a northern extension of the Oquirrh basin, and in the area of the Wells Formation. Minor granule to pebble conglomerate reached the area of the Sublett basin from the Copper Basin highland to the north.

Sources for the tremendous volumes of Middle and Upper Pennsylvanian sand in the cratonic-platform basins of Idaho, Montana, and Utah remain elusive. Petrography of the fine-grained sand, of probable second- or third-cycle origin, suggests a cratonic rather than a western orogenic source; hence, the northeasterly, easterly, and southeasterly source areas proposed by Bissell (1974) appear to be the most reasonable for southern Idaho.

The influence of detritus shed eastward from a possible western Humboldt highland (Ketner, 1977) is not clear. Perhaps the record of this material is lost in the area of the Idaho batholith. Possibly, more petrographic work will show that both cratonic and western orogenic source terranes provided material for some of the Pennsylvanian and Permian basins.

PLATE-TECTONIC MODELS

Major changes in the structural settings for deposition of Carboniferous rocks in Idaho took place in latest Devonian to earliest Mississippian time. These periods are coincident with major plate-boundary tectonic events along the western margin of the proto-North American continent.

A postulated plate-tectonic model for the Antler orogeny is that of back-arc thrusting during a period of increased plate motion in which the sediments of an inner-arc basin were obducted onto the continental margin east of an offshore island arc; west of the island arc was an east-dipping subduction zone (fig. 16). The obducted inner-arc basin sediments in this model make up the Antler orogenic belt (Burchfiel and Davis, 1972, 1975; Poole, 1974; Poole and Sandberg, 1977; Dickinson, 1977). In this model, the inner-arc basin continued to close during Pennsylvanian and Permian time, and the island arc (Klamath-North Sierran arc of Poole and Sandberg, 1977, p. 82, fig. 8) collided with the proto-North American continent in Triassic time (Silberling, 1973) and became welded to the continent. This interpretation was used to construct model I of Paleozoic plate-tectonic events (fig. 16).

Dickinson (1977) noted the short duration of the Antler orogeny and pointed out that flysch deposition was not followed by molasse accumulation but by a return to modified shelf conditions. Carboniferous rocks in Idaho record exactly this sequence of events. Dickinson suggested that, perhaps, mid-Carboniferous rifting broke up the Antler orogen and interrupted the maturing process, producing instead a fault-controlled rift topography on which the many basins and intervening local highlands of the Pennsylvanian and Permian Systems formed. In this model, the Humboldt highland in Nevada might be a westward-rifted and uplifted remnant of the old Antler highland. A postulated model (model II) of this concept of plate-tectonic events is presented in figure 17. In this reconstruction, the Antler orogeny is the result of an arc-continent collision along a west-dipping subduction zone (Dickinson, 1977). The Antler collision was followed by a period of rifting during which fault-bounded basins and highlands formed on the cratonic platform. In Permian time, a volcanic arc again approached the proto-North American continent and was welded to it in mid-Triassic time (Dickinson, 1977).

ECONOMIC GEOLOGY

Carboniferous rocks in Idaho had not produced any coal or petroleum to date (1977). Minor coaly beds like those in the upper part of the Copper Basin Formation are present but are neither extensive enough nor of high enough rank to be commercial deposits.

Petroleum exploration in the overthrust belt of eastern Idaho is in progress, and the area is considered potentially productive (Monley, 1971; Royse and others, 1975; Powers, 1977). Mississippian carbonate-shelf complexes (Rose, 1976), and the sandstones and carbonate rocks of the Pennsylvanian cratonic-platform-basin sequences are potential reservoir rocks. The starved-basin facies of the Deep Creek and Little Flat Formations (Poole and Sandberg, 1977; Sandberg and Gutschick, 1977) and the phosphatic shales and interbedded quartzites of the Big Snowy Formation and the Manning Canyon Shale are both potential source and reservoir rocks. The structurally complex setting of these rocks discouraged earlier exploration, but the present need for more domestic hydrocarbons has opened the area to the search for gas and oil.

Metalliferous deposits have a long history of exploitation in Idaho; a description by W. E. Hall of known deposits associated with Carboniferous rocks follows.

Mississippian and Pennsylvanian carbonate rocks are also a valuable resource in Idaho, and a summary of their economic potential is included.

METALLIFEROUS DEPOSITS

Carboniferous rocks in central and south-central Idaho contain many metalliferous deposits that have been exploited sporadically since the 1880's. The mineral resources, listed in approximate decreasing order of economic significance, include deposits of lead, silver, molybdenum, copper, zinc, barite, gold, tungsten, and subeconomic tin. Although the deposits in general are small and the production moderate (estimated at approximately $20 million, fig. 18), several very large low-grade deposits are known, but they have not been developed extensively.

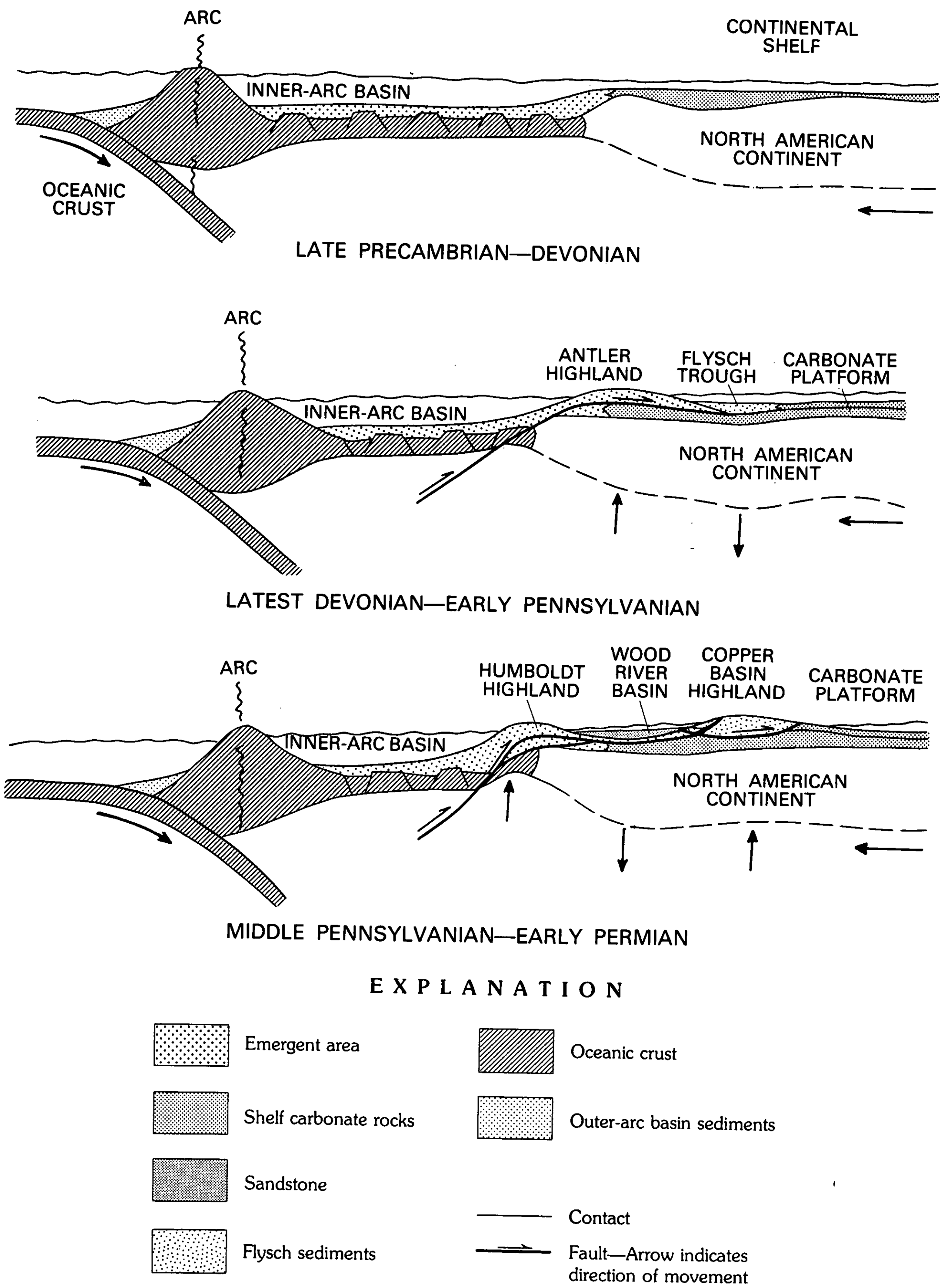


FIGURE 16.—Model I of plate-tectonic events along western margin of proto-North American continent showing an east-dipping subduction zone west of an island arc and a gradually closing inner-arc basin. Modified from Poole (1974) and Poole and Sandberg (1977).

Most favorable sites for ore in the Carboniferous are in contact-metamorphosed calcareous beds beneath thrust faults and near leucocratic granitoid or porphyritic stocks, sills, or dikes of Cretaceous or Eocene age. The Copper Basin Formation of Mississippian age, the White Knob Limestone of Late

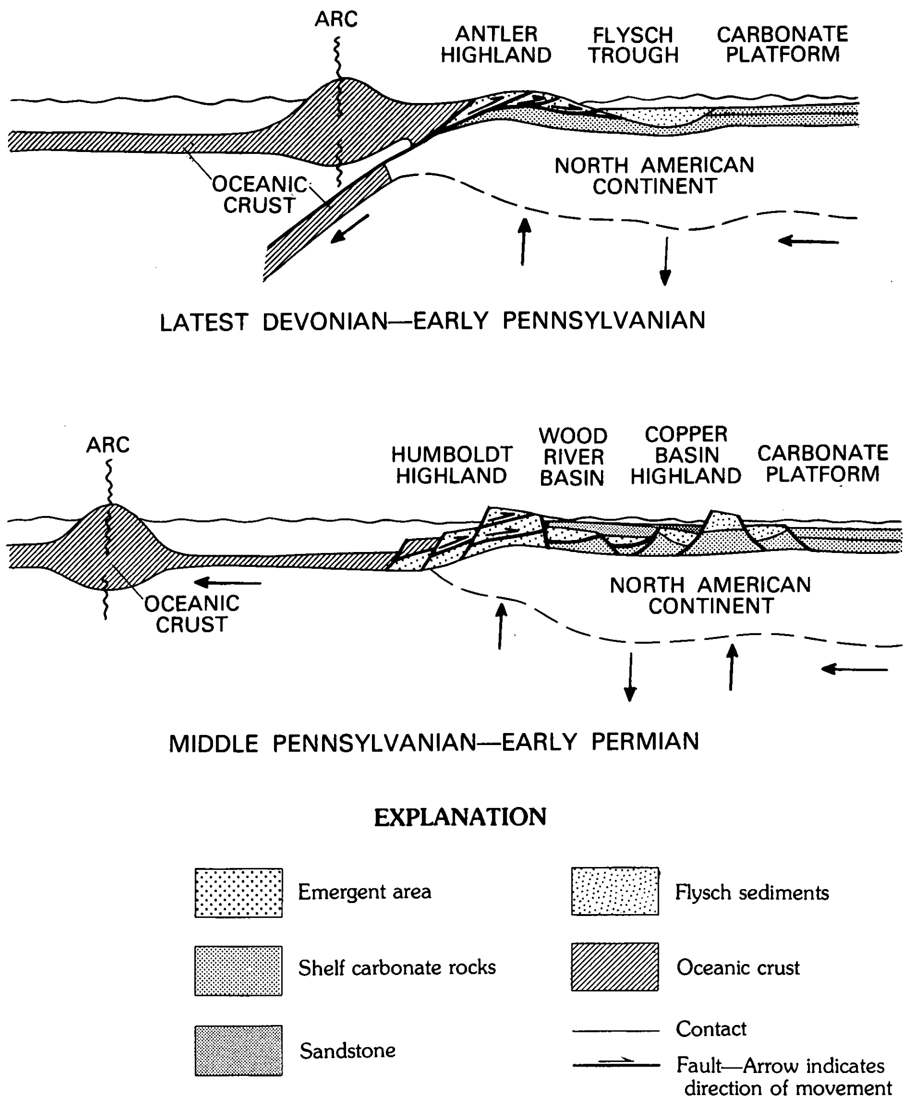


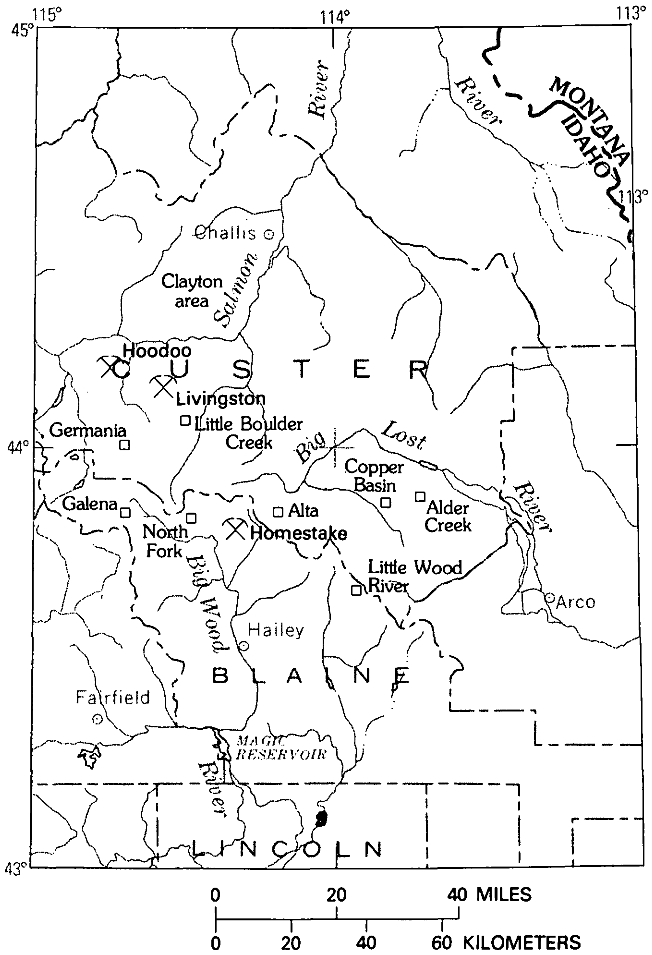
FIGURE 17.—Model II of plate-tectonic events along the western margin of proto-North American continent showing an arc-continent collision along a west-dipping subduction zone, an event that produced the Antler orogeny. This orogeny was followed by mid-Carboniferous rifting (Dickinson, 1977).

Mississippian age, and the Wood River Formation of Pennsylvanian-Permian age each host significant deposits.

Ore occurs in the Copper Basin Formation, in the more calcareous sequences where cut by intrusive masses in the Little Wood River and Alta mining districts and, according to C. M. Tschanz (written commun., 1977), at the Hoodoo and Livingston mines at the north end of the Boulder Mountains and at the Twin Apex and Bruno mines in the Clayton area, where the deposits are controlled by a thrust fault at the base of the Copper Basin Formation (S. W. Hobbs, oral commun., 1977). The Little Wood River district contains small but high-grade lead-silver and some barite and zinc deposits (Anderson and Wagner, 1946), and the Alta district has small lead-silver-gold deposits (Umpleby and others, 1930; U.S. Geol. Survey and U.S. Bur. Mines, in press). The Livingston and Hoodoo mines have produced substantial amounts of zinc as well as some lead, silver, and gold (Tschanz and others, 1974), and the Twin Apex and Bruno mines have produced lead, zinc, and a little silver and gold.

The White Knob Limestone is the host rock in the Alder Creek district near Mackay and at the Copper Basin mine. About 90,000 kg of copper was produced from replacement deposits in quartzite and limestone in the Copper Basin district (Umpleby, 1917; Vhay, 1964, p. 71). The Alder Creek district has yielded 26 million kg of copper as well as appreciable amounts of lead, silver, and gold (Nelson and Ross, 1968, p. A29–A30). The ore deposits are in irregular replacement bodies in granite porphyry and in tactite from contact-metamorphosed White Knob Limestone.

The Wood River Formation at its type locality in the vicinity of the Big Wood River valley is an unfavorable host for metalliferous deposits, and nearly all the ore in that area has a structural control in the Devonian Milligen Formation beneath the Wood River thrust (Anderson and others, 1950, p. 10; Hall and Czamanske, 1972, p. 350). However, in the Boulder Mountains, the Wood River Formation is the host rock for many deposits (Tschanz and others, 1974). These include the small high-grade lead-silver deposits in the North Fork, Galena, and Germania districts and the large low-grade disseminated molybdenum deposit at the southeast end of the White Cloud stock in the Little Boulder Creek district. In addition, tin occurs in subeconomic amounts in the lead-silver deposits in an arcuate north-trending belt about 30 miles long and 4 miles wide along the crest of the Boulder Mountains (Tschanz and others, 1974, p. 251). Locally, the concentrations of tin are high, and the potential tin resource is certainly significant.



Production

	$ Million
Livingston mine	2.3
North Fork and Galena districts	1.4
Hoodoo mine	.6
Little Wood River district	.2
Germania district	.45
Alder Creek district	12.7
Alta district	.1
Homestake mine	.2
Total	17.95 (≈$20M)

FIGURE 18.—Locality map and production figures, in millions of dollars, for major metalliferous deposits related to Carboniferous rocks in Idaho.

CARBONATE ROCKS

Idaho contains large reserves of carbonate rocks; many of these are present in the marine Carboniferous rocks of the south, south-central and southeastern parts of the State. Although Idaho's production of limestone has remained small, the widespread distribution of relatively pure calcium carbonate limestone of Pennsylvanian and Mississippian formations—such as the Scott Peak and Surrett Canyon, the Madison Group, and parts of the Oquirrh and Wells Formations—provides potential resources for cement, agricultural and industrial lime, flux, and the manufacture of paper, vinyl, and glass (Savage, 1969).

Small quantities of limestone and marble are used in construction as both dimension and broken stone.

More carbonate rock could be used in road construction, where strategically located, but roadbuilders in southern Idaho have tended to rely upon basalt aggregate from the vast resources of the Snake River Plain rather than the local limestone. A comprehensive region-by-region review of carbonate rocks and their economic potential in Idaho has been prepared by Savage (1969).

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Nevada

By E. R. LARSON *and* RALPH L. LANGENHEIM, JR.

With a section on Paleontology

By JOSEPH LINTZ, JR.

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in Nevada

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THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—NEVADA

By E. R. LARSON[1], and RALPH L. LANGENHEIM, JR.[2]

ABSTRACT

The entire Carboniferous is represented in discontinuous exposures of clastic and carbonate sedimentary rocks in eastern Nevada, but eugeosynclinal assemblages (shale, chert, turbidites, volcanic rocks) in thrust slices in central Nevada may lack parts of the system.

Mississippian clastic rocks derived from the north-trending Antler orogenic highlands in central Nevada are as much as 2,500 m thick. They are coarsest in central Nevada, but are fine and thin eastward through a starved basin to a carbonate-bank assemblage in central Utah and southeast Nevada. Mississippian eugeosynclinal rocks consist of chert, shale, turbidites, and lesser volcanic rocks thousands of meters thick in thrust slices in central Nevada.

Pennsylvanian rocks are primarily limestone in eastern Nevada and coarse clastic facies near the Antler belt. Thickest accumulations are in the Bird Spring-Ely Basin, farther east than the Mississippian maximum. Widespread disconformities in both Mississippian and Pennsylvanian rocks in eastern Nevada reflect continued instability in the Cordilleran miogeosyncline throughout the Carboniferous.

Pennsylvanian eugeosynclinal rocks deposited in a trough west of the Antler belt are contemporaneous deposits of shale, chert, and volcanic rocks (Pumpernickel Formation) and turbidites together with lesser chert and some limestone (Havallah Formation). These beds are as much as 5,000 m thick in incomplete sequences in western-derived thrust slices in north-central Nevada.

Carboniferous faunas in the carbonate belt have been well studied and allow good paleontologic control. Brachiopods are common throughout the Carboniferous, but mollusks are more abundant in the clastic sequences. Fusulines are abundant and diagnostic in the Pennsylvanian rocks in the carbonate belt but are less common in the lenticular limestone of the eugeosynclines. Conodonts are widespread and distinctive. Their presence in both clastic- and carbonate-facies rocks has allowed reliable interfacies correlations.

INTRODUCTION

Carboniferous beds (figure 1) were recognized in the Spring Mountains in southern Nevada by the Wheeler Survey of 1872–1873 (Wheeler, 1875) and in eastern Nevada by the 40th Parallel Survey (King, 1878), 1870–1878. Study of the Eureka district, begun by Hague and Emmons (1877) as a part of the 40th Parallel Survey, was continued by Hague (1883), whose monograph on the Eureka district (1892) established much of the Paleozoic stratigraphic terminology of east-central Nevada. Modern stratigraphic studies have brought many new subdivisions, but only a few detailed regional studies that synthesized Carboniferous paleontology, sedimentology, and stratigraphic relationships have been made (Bissell, 1964; Dott, 1955; Smith and Ketner, 1975; Steele, 1960). None of these was on a statewide basis. Figure 2 outlines the development of Carboniferous stratigraphic terminology in the State. Apparent duplication of names is due in part to scattered exposures and uncertainties of correlation.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Nevada Bureau of Mines and Geology.

The facies and the original distribution of Carboniferous strata in Nevada were determined by the prior depositional-tectonic history of the State. Beginning in the late Precambrian and continuing to the middle Paleozoic, carbonate sedimentary rocks and lesser shale and orthoquartzite formed a thick wedge on the western margin of the North American craton—the Cordilleran miogeosyncline. This carbonate assemblage graded west into deep-water deposits of dark shale, bedded chert, medium clastic materials, turbidites, and volcanic rock—the Cordilleran eugeosyncline. Eugeosynclinal rocks were carried eastward onto the miogeosyncline by the Roberts Mountain thrust during the late Devonian-early Mississippian Antler orogeny. This orogenic activity formed the north-trending Antler orogenic belt, an uplifted area through central Nevada that separated an eastern foreland basin containing clastic and carbonate rocks from a western shale-

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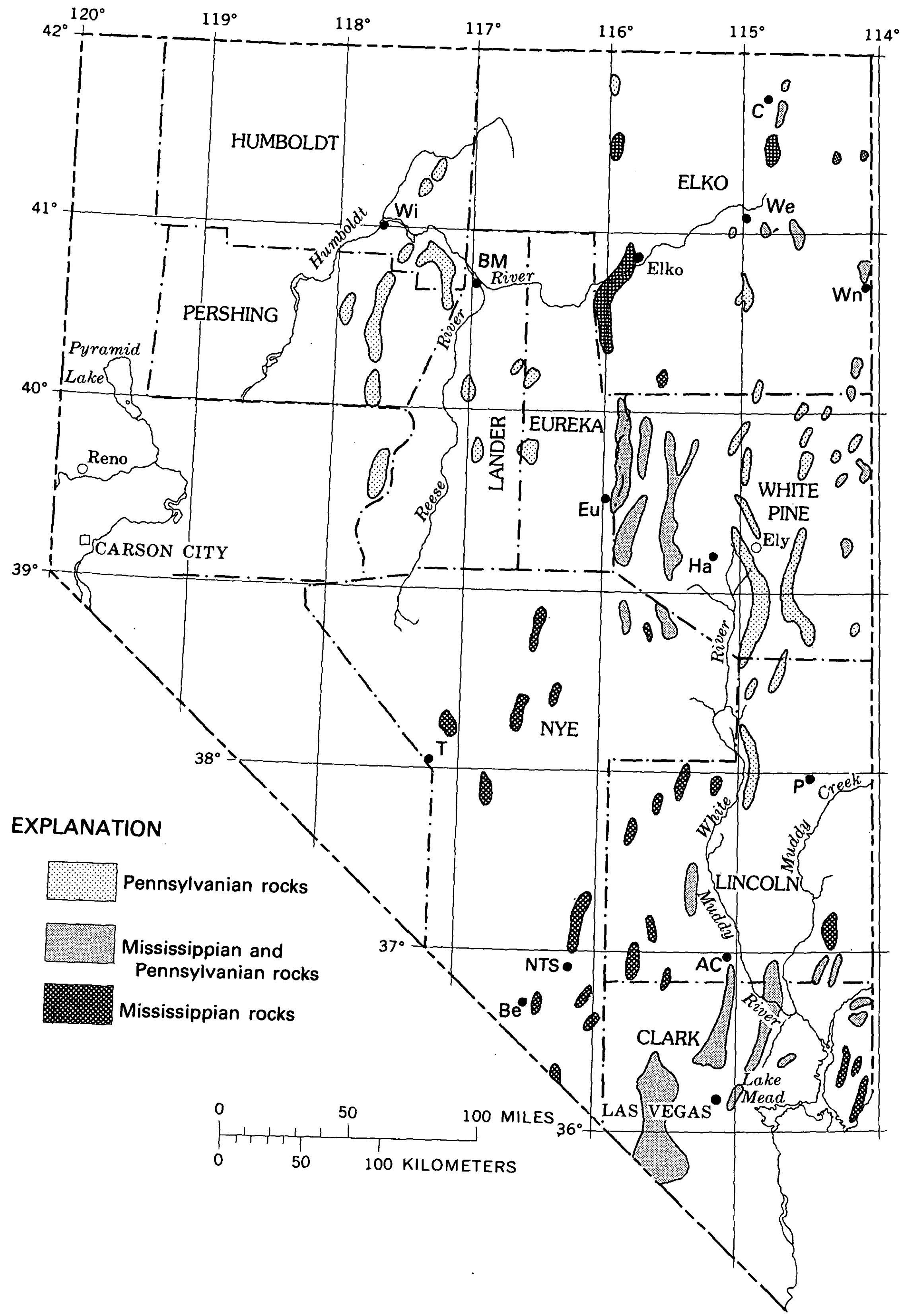
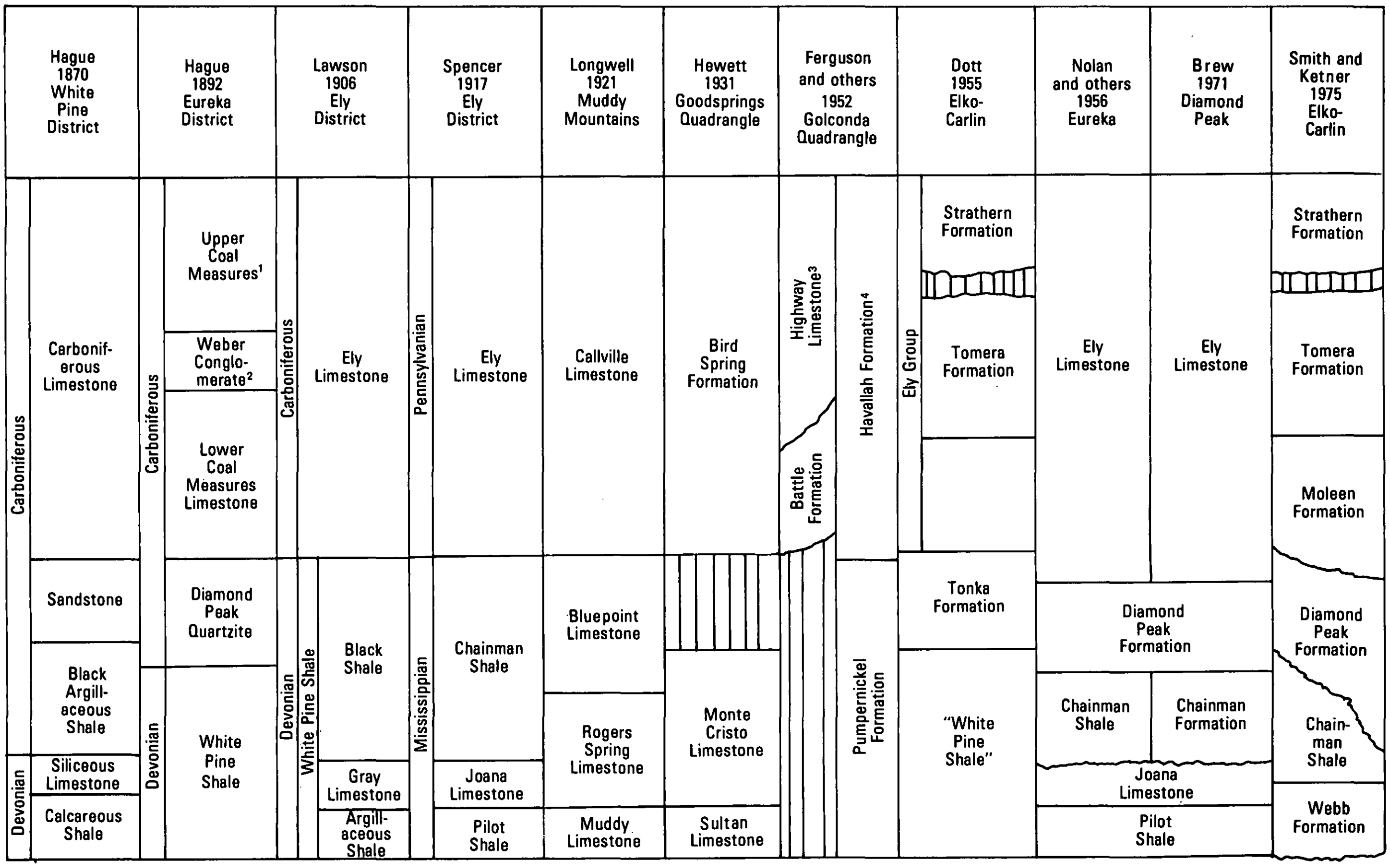


FIGURE 1.—Carboniferous outcrops. Localities: AC, Arrow Canyon Range; Be, Beatty; BM, Battle Mountain; C, Contact; Eu, Eureka; Ha, Hamilton; NTS, Nevada Test Site; P, Pioche; T, Tonopah; We, Wells; Wi, Winnemucca; Wn, Wendover.



[1]Fault block of Lower Pennsylvanian
[2]Cretaceous
[3]Autochthonous
[4]Allochthonous

FIGURE 2.—Development of Carboniferous terminology in Nevada.

volcanic sequence. The Antler belt was an active source of clastic sediments during the Permo-Carboniferous. Figures 3 and 4 show the stratigraphic belts and representative stratigraphic columns from them.

The stratigraphy of the foreland belt is well known but is complicated by facies differences and later tectonic displacements.

The eugeosyncline (shale-volcanic belt) can only be interpreted from strata preserved in klippen that lie on the foreland-basin sequence or on the Antler belt.

The original distribution of Carboniferous rocks has been modified by major west-to-east thrusting during the lower Triassic (Golconda thrust) and middle Cretaceous through lower Tertiary Sevier and Laramide orogenies. Middle Tertiary through Holocene block faulting (Basin and Range province) has segmented all older structures. Major strike-slip faults (Death Valley-Furnace Creek shear zone, Las Vegas shear, Wells fault; the last two features shown on fig. 5) have displaced Carboniferous sedimentary rocks tens of kilometers.

MISSISSIPPIAN SYSTEM

MISSISSIPPIAN FORELAND BASIN

LOWER MISSISSIPPIAN

Lower Mississippian (Kinderhookian, Osagean, lower Meramecian) beds are 1,000 m thick in a north-trending belt from Elko to Beatty (fig. 5). These deposits thin abruptly westward to zero on the eastern edge of the Antler belt and thin gradually to a few hundred meters in eastern Nevada, where they may be locally absent. The area of thickest sedimentary rocks first received western-derived medium- and fine-grained clastic deposits. In the Eureka area (fig. 4), the black Pilot Shale conformably overlies the Nevada Formation of Devonian age. Conodonts indicate that the Devonian-Mississippian boundary is within the shale. To

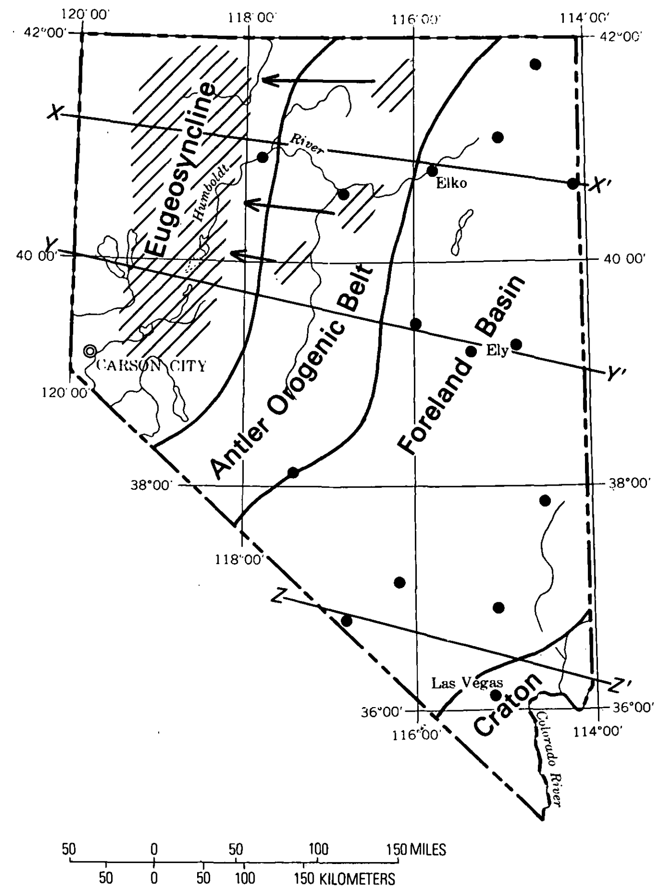


FIGURE 3.—Paleotectonic map of Carboniferous elements in Nevada. Original site of deposition of allochthonous eugeosynclinal units indicated by ruled pattern. *X–X'*, *Y–Y'*, and *Z–Z'* are lines of restored sections (block diagrams) shown in figure 9.

the east near Ely, the Pilot Shale is discontinuous, the result of post-Pilot warping and erosion (Langenheim, 1960, 1961).

Fine arenites in the Pilot Shale near Wendover (figs. 5, 6) have Buoma sequences in which all elements are present. These rocks are interpreted as proximal turbidites (Poole, 1974, p. 66), part of a submarine fan that extended southward from Idaho. Graded beds are again present in the Eleana Formation near Beatty in southern Nevada (Barnes and others, 1963) (fig. 6). These Lower Mississippian turbidites are the early flysch deposits of the Antler orogeny.

The Joana Limestone, which overlies the Pilot Shale in the Eureka-Ely area, is continuous to the east and southeast with the lower part of a carbonate bank that extends southwest from central Utah to southern Nevada (figs. 6, 9). The relationship of the Joana to contiguous beds is complex and is generally disconformable. Near Ely, the Joana lies on the Pilot Shale or directly on the Devonian Guilmette Formation as the result of regional warping and erosion (Langenheim, 1960, 1961). Near Eureka, the Joana lies disconformably (Langenheim, 1961) or conformably (Nolan and others, 1956) on the Pilot. The pattern seems to be one of regional disconformity.

The Joana Limestone thickens from 100 m at Ward Mountain near Ely (Langenheim, 1960) to 300 m near Pioche and is laterally continuous with the Dawn Limestone of the Monte Cristo Group in the Arrow Canyon Range (fig. 6). The contact of the Joana with the overlying Chainman Formation is a regional disconformity according to MacKenzie Gordon, Jr. (*in* Brew, 1971, p. 35), on the basis of limited faunal evidence, but Rose (1976, fig. 3) has interpreted the Joana and Chainman as a continuous sequence that reflects deposition on a carbonate bank and adjacent basin. The Joana is locally absent east of Eureka (Nolan and others, 1956), and changes in thickness from 50 to 150 m in 5 km have been observed (Larson and Riva, 1963; Brew, 1971). Joana time was one of minimum influx of clastic sediments into the foreland basin. The post-Joana hiatus that represents most of the Meramecian in eastern Nevada (Mackenie Gordon, Jr., *in* Brew, 1971, p. 34-77) resulted from uplift in the Antler area and warping in the basin. The upward-coarsening flysch deposits of the overlying Chainman Formation indicate progressive uplift in the Antler source area.

In the Elko-Carlin area, the Kinderhookian Webb Formation (Smith and Ketner, 1975, A38) is a laminated to thin-bedded claystone containing limestone lenses equivalent to the upper Pilot Shale and the Joana Limestone. The Webb lies on both autochthonous Devonian Limestone and allochthonous Ordovician eugeosynclinal deposits (Valmy Formation). The lithology of the Webb indicates that the formation was deposited nearer to an active source of fine clastic materials than were the Pilot and Joana, a condition that is emphasized in the succeeding Chainman and Diamond Peak Formations.

Glauconitic black shale that overlies the Joana Limestone in easternmost Nevada and adjacent Utah was deposited in a starved basin between the toe of the Antler flysch wedge and the central Utah carbonate bank (Poole and Sandberg, 1977) (figs. 5, 6). A carbonate buildup of the Joana Limestone isolated the basin from a western source of sediments, so that the black shale was largely of eastern origin (Poole, 1974).

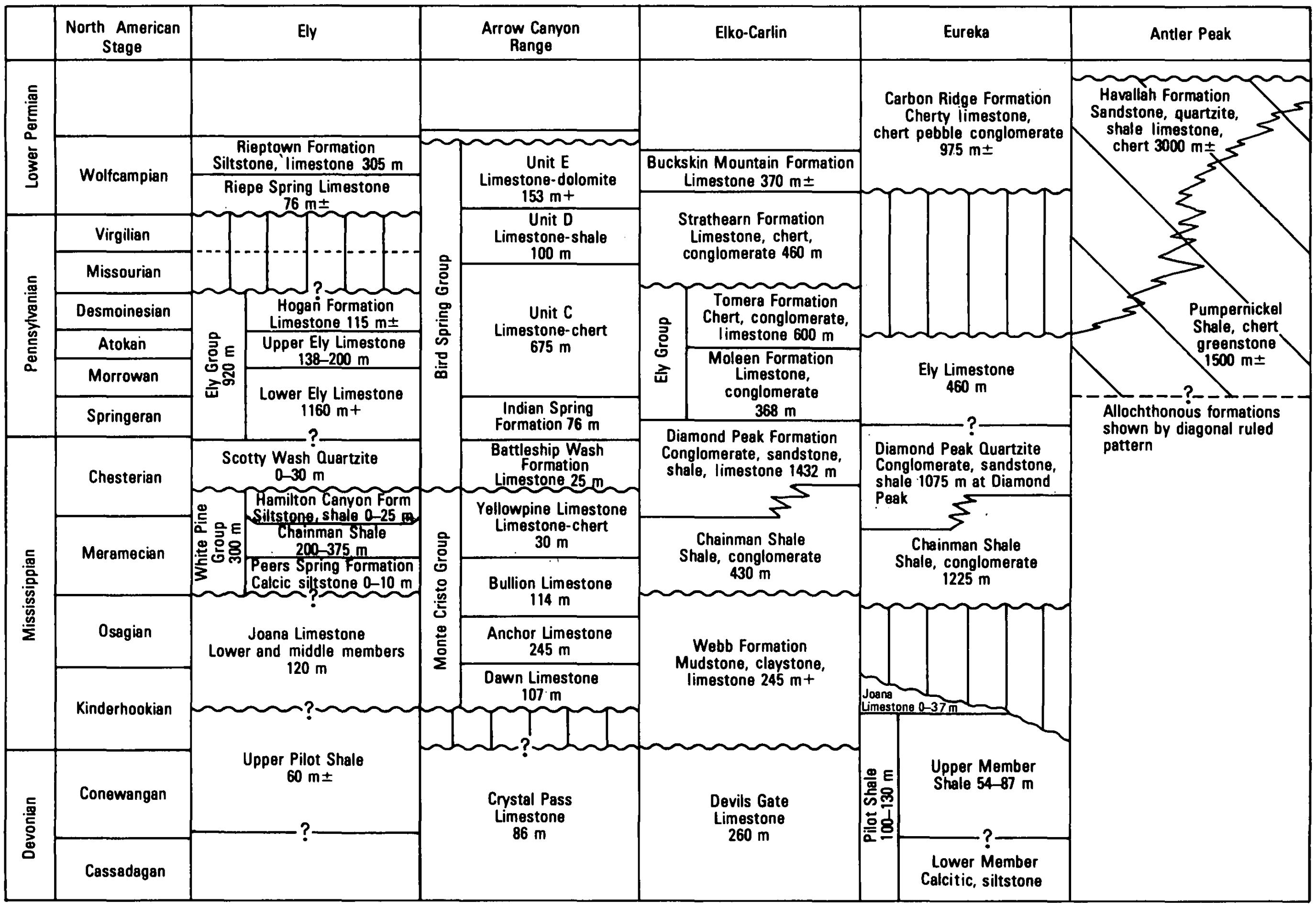


FIGURE 4.—Correlation chart of representative Carboniferous assemblages from stratigraphic belts.

UPPER MISSISSIPPIAN

Upper Meramecian and Chesterian deposits in eastern Nevada, the flysch and molasse of the Antler orogeny, are thickest adjacent to the Antler belt, reaching a thickness of more than 1,500 m near Elko and Beatty (fig. 7). The thickest deposits contain thick conglomerate sequences (Diamond Peak Formation) that overlie and interfinger eastward with a shale-arenite formation (Chainman). The terminology of the upper Mississippian has been the source of much discussion. Because of the eastward transgression of the Diamond Peak, the formation boundaries are not time horizons and are recognized differently by various workers.

The name Illipah has been used for arenites that occupy the position of the Diamond Peak Formation in the Ely area and in exploratory oil wells drilled near Hamilton, but the name is not recognized by the U.S. Geological Survey. Humphrey (1960) applied the name Illipah to Eocene deposits in the White Pine mining district.

The Chainman Shale (Formation) is approximately 100 m of generally gray to greenish-gray silty and sandy shale that disconformably overlies the Joana Limestone near Ely, the type area (Spencer, 1917). Westward, in the Eureka-Hamilton area, the Chainman overlies the Joana Limestone (locally on the Pilot); near Elko and Carlin, it overlies the Webb.

Graded sandstone turbidites in the Chainman were deposited in deep water, attested to by meandering trails (*Nereites*) on bedding surfaces. Pebbly mudstones near Elko (Smith and Ketner, 1975) contain well-rounded chert and quartzite cobbles, indicating proximity to the source, but those near Eureka have only chert pebbles, indicating a more distal deposit.

DIAMOND PEAK FORMATION

Maximum elevation of the Antler belt is marked by conglomerate of the Diamond Peak Formation, which is the molasse of the Antler orogeny. This

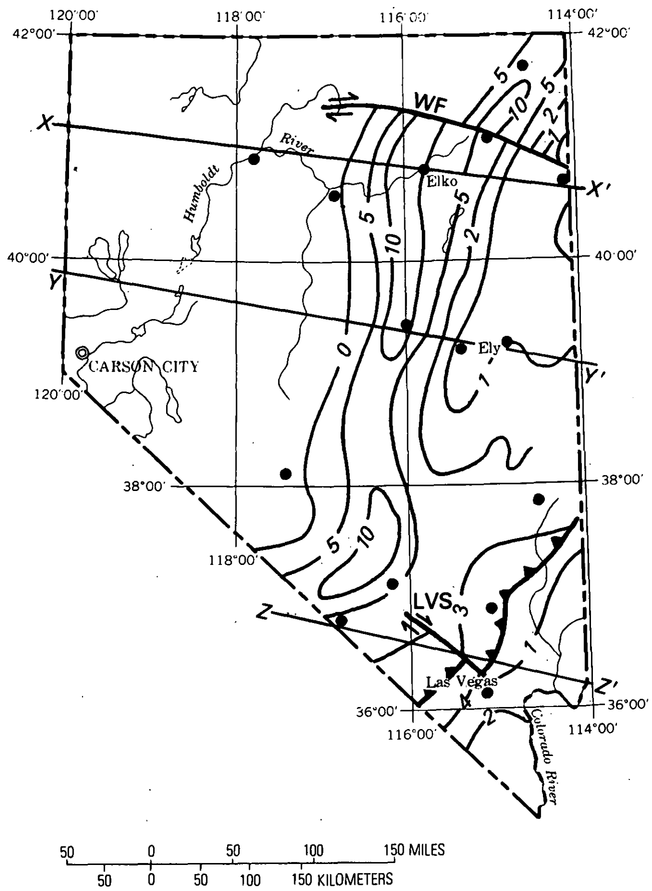


FIGURE 5.—Thickness of Lower Mississippian (Kinderhookian, Osagean, lower Meramecian) rocks. Isopachs in hundred of meters. Allochthonous rocks are restored to the site of original deposition. Faults: LVS, Las Vegas shear; WF, Wells fault. *X–X'*, *Y–Y'*, and *Z–Z'* are lines of restored sections (fig. 9). Modified from Poole and Sandberg (1977).

assemblage of coarse to fine clastic deposits and some limestone is more than 1,000 m thick in the Eureka-Elko-Carlin belt, but it thins rapidly eastward and cannot be recognized with certainty east of Hamilton. The proportion of conglomerate in the upper Mississippian decreases eastward and southeastward from near 50 percent near Elko to less than 2 percent near Ely (fig. 8).

The Diamond Peak is Meramecian and Chesterian and contains the Mississippian-Pennsylvanian boundary near Elko (Mackenzie Gordon, Jr., *in* Brew, 1971, p. 34–55; fig. 11). It is entirely Meramecian and Chesterian near Eureka, but is only upper Chesterian near Hamilton (Mackenzie Gordon, Jr., *in* Brew, 1971, p. 34–77). Upper Chesterian arenites near Pioche (Scotty Wash Formation) have been correlated with the Diamond Peak, but may have been derived from the east (James, 1954).

It is reasonable that the upper limit of the Diamond Peak is younger toward the north than near Eureka, which is farther from the principal source of clastic sediments.

Lenticular bodies of conglomerate in the upper Chainman-lower Diamond Peak both in the Carlin area and near Eureka are debris flows and distributary channels on a submarine fan that extended southeast from the northern extension of the Antler belt (Rose, 1976). These lenticular bodies—cross sections of channel fillings as much as 500 m wide and 20 m thick—are well seen on the western slope of the Diamond Mountains near Eureka. Well-washed chert pebble conglomerates are in a series of overlapping channels near the base of the Dia-

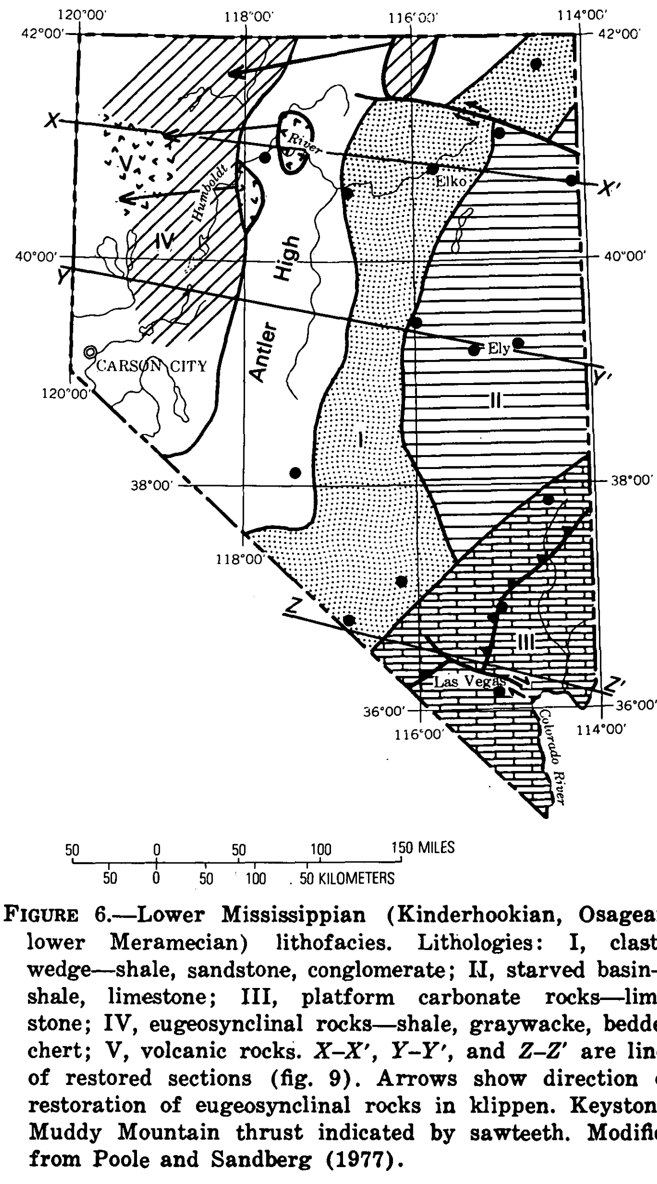


FIGURE 6.—Lower Mississippian (Kinderhookian, Osagean, lower Meramecian) lithofacies. Lithologies: I, clastic wedge—shale, sandstone, conglomerate; II, starved basin—shale, limestone; III, platform carbonate rocks—limestone; IV, eugeosynclinal rocks—shale, graywacke, bedded chert; V, volcanic rocks. *X–X'*, *Y–Y'*, and *Z–Z'* are lines of restored sections (fig. 9). Arrows show direction of restoration of eugeosynclinal rocks in klippen. Keystone-Muddy Mountain thrust indicated by sawteeth. Modified from Poole and Sandberg (1977).

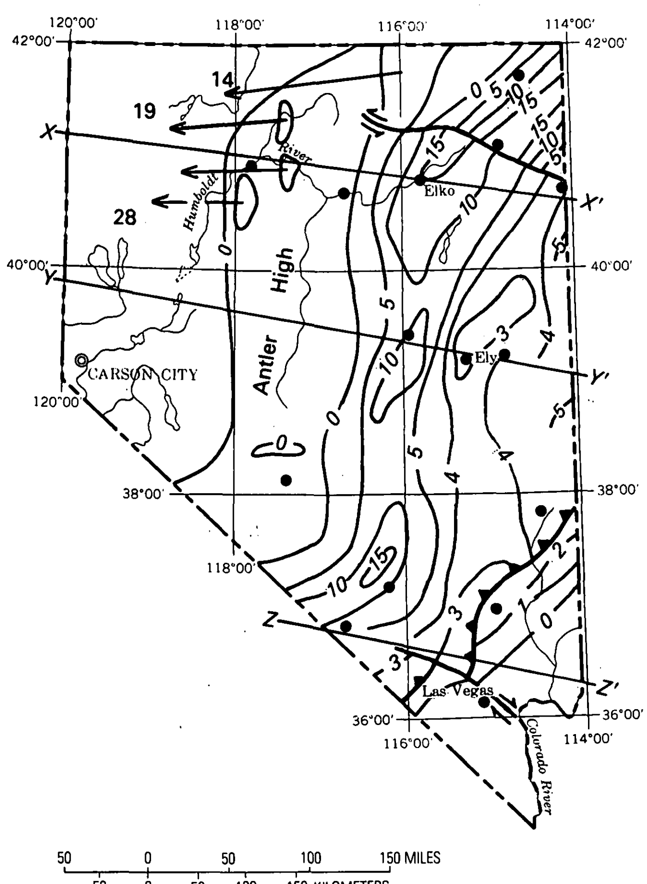


FIGURE 7.—Thickness of Upper Mississippian (upper Meramecian, Chesterian) rocks. Isopachs in hundreds of meters. Allochthonous rocks are restored to the original site of deposition. X–X', Y–Y', and Z–Z' are lines of restored sections (fig. 9). Modified from Poole and Sandberg (1977).

mond Peak Formation at Buck Mountain east of Eureka (Blomquist, 1971). Crossbeds that dip approximately 30° indicate southward transport. Regional studies of directional features in the Diamond Peak (fig. 8) indicate dominant southward transport, but some transport was from the east or the northeast, suggesting control of currents by a high area near Ely (Ely high).

Faunas in the Chainman-Diamond Peak transition zone in the Diamond Mountains north of Eureka are productoids, pelecypods, echinoids, crinoids, and trilobites. This assemblage is in muddy beds interlayered with pebble and cobble conglomerates. The depositional environment is interpreted as having been relatively shallow but below the zone of wave action, in contrast to the deep-water depositional site of the pebbly mudstones in the lower Chainman. This change indicates eastward filling of the foreland basin. Upper Mississippian deposits spread progressively eastward and cover the lower Mississippian starved basin, which was reduced to a small area on the Utah-Nevada border in Chester time (fig. 8).

The contact of the Diamond Peak and the Pennsylvanian Ely Limestone is in a zone of interbedded limestone, conglomerate, and finer clastic materials (Brew, 1971). The horizon chosen in a local area cannot be traced widely, and different limits have been used by different workers in the same area. The type Diamond Peak Formation has been variously interpreted as entirely Chesterian (Nolan, 1956; Brew, 1971) or as including basal Pennsylvanian strata (Dott, 1955, fig. 13; Easton, 1953; Langenheim and Larson, 1973). Arnold and Sadlick (1961) defined the Mississippian-Pennsylvanian

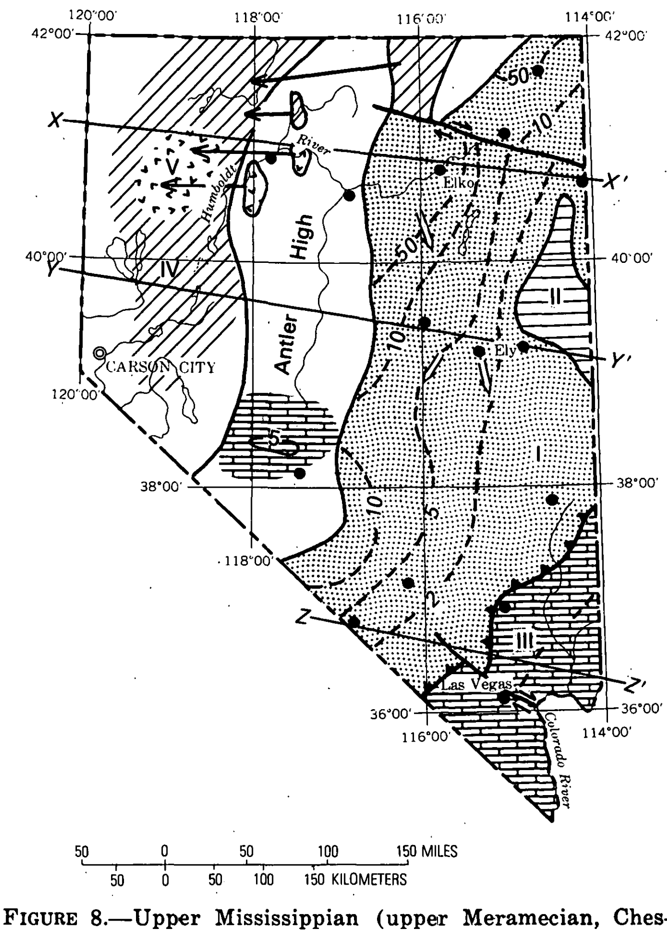


FIGURE 8.—Upper Mississippian (upper Meramecian, Chesterian) lithofacies. Allochthonous units are restored to site of deposition indicated by arrows. X–X', Y–Y', and Z–Z' are lines of restored sections (fig. 9). Lithologies: I, clastic wedge (isopleths indicate percent of conglomerate); II, starved basin, phosphatic shale, limestone; III, limestone; IV, eugeosynclinal sedimentary rocks; V, eugeosynclinal volcanic rocks. Double-shafted arrows indicate current directions from cross-lamination and bedding surfaces. Modified from Poole and Sandberg (1977).

transitional sequence of limestone, shale, and quartz arenite between the Chainman and the Ely in the Wendover area as the Jensen member of the Chainman. The Jensen was widely recognized by Bissell (1964, fig. 3).

Mississippian clastic-wedge sedimentary rocks in southern Nevada constitute the Eleana Formation, a 2,000-m thick unit of turbidites and lesser chert and shale, which extends from the vicinity of Goldfield and Beatty eastward toward Las Vegas. Turbidites suggest deposition in deep water (Poole and others, 1965; Poole and Sandberg, 1977). The age relations of the Eleana are not well known, but the formation evidently encompasses the entire Mississippian. The lithology of the beds indicates that deposition was well removed from a source of coarse clastic sediments, but some pebble beds have been observed in the vicinity of Beatty (figs. 7, 8, and 9, Z–Z′).

Mississippian formations in southeast Nevada comprise the Monte Cristo group and the lowermost Bird Spring group (see fig. 4). These units are a part of the carbonate bank that extended from central Utah southwestward through southern Nevada. A minor disconformity that separates the upper Monte Cristo (Yellowpine Limestone) from the uppermost Mississippian Battleship Wash Formation in the Arrow Canyon Range is within a single conodont zone.

MISSISSIPPIAN EUGEOSYNCLINE

Mississippian shale-volcanic-assemblage rocks in klippen of the Golconda thrust near Elko and Winnemucca are samples of the eugeosyncline that was to the west during the Carboniferous. North of Elko, the Schoonover Formation (Fagan, 1962) is 2,500 m of radiolarian chert, turbidite, and shale that has volcanic rock near the base. The Schoonover is now dated as Late Mississippian in age (Poole and Sandberg, 1977). Fagan (1962) believed that the formation is composed of deep-water sediments derived from a sourceland toward the west within the eugeosyncline.

The Gogh Canyon Formation in the Osgood Mountains northeast of Winnemucca is 1,500 m of pillow lava and volcanic breccia (Hotz and Willden, 1964). These early Osagean or late Kinderhookian rocks are believed to be shallow-water volcanic rocks and shallow siltstone (Hotz and Willden, 1964). The Inskip Formation in the East Range near Winnemucca resembles the Gogh Canyon.

Figure 9, a palinspastic interpretation of Mississippian rocks along a series of sections in southern, central, and northern Nevada, indicates the original position of the now-displaced facies belts and the role of the Antler orogenic belt.

PENNSYLVANIAN SYSTEM

PENNSYLVANIAN FORELAND BASIN

INTERVALS A, B, AND C

Mississippian clastic wedge sedimentary rocks in eastern Nevada are succeeded by the limestones of the Ely Formation in the northern part of the State and by the Bird Spring Group in the south. This carbonate deposition continued without significant interruption through Morrowan and Desmoinesian time (intervals A, B, C), the principal accumulation being in the southwest-plunging Ely-Bird Spring Basin in Clark, Lincoln, and southern White Pine Counties (figs. 1, 10). These rocks are thinner on the Ely Platform, which separates the southern basin from the supposed Butte Basin in the north in Elko County. These regional variations in the thickness of contemporaneous strata reflect continued warping of the Cordilleran miogeosyncline.

The Mississippian-Pennsylvanian boundary in the heart of the miogeosyncline has been well documented as occurring in the lower part of unit "C" of the Bird Spring Group or within the lowermost Ely Formation, or their temporal equivalents. On the western margin of the foreland basin or miogeosyncline, the systemic boundary is within the Diamond Peak Formation in some areas, as near Elko (fig. 9, Elko-Carlin section), where conditions controlling Diamond Peak deposition prevailed from the middle Mississippian into the Pennsylvanian. Middle Pennsylvanian strata near Elko contain lenses of chert pebble conglomerate like those of the Diamond Peak.

Morrowan through Atokan parts of the Bird Spring Group are characterized by cyclically alternating layers of argillaceous fine-grained limestone, nodular limestone, and thicker bedded, cherty, arenaceous medium-grained limestone in the southern part of the Ely-Bird Spring Basin (fig. 11). Desmoinesian parts of the Bird Spring Group here are somewhat more arenaceous. In the northern part of the basin and on the Ely Platform, the Morrowan through Atokan Ely Limestone consists of thick-bedded, more argillaceous and silty limestone and has been referred to as the Hogan Formation. The eastern part of the Butte Basin is occupied by rocks referred to as the Ely Limestone and Hogan Formation (Bissell, 1964); sandy and conglomeratic mid-

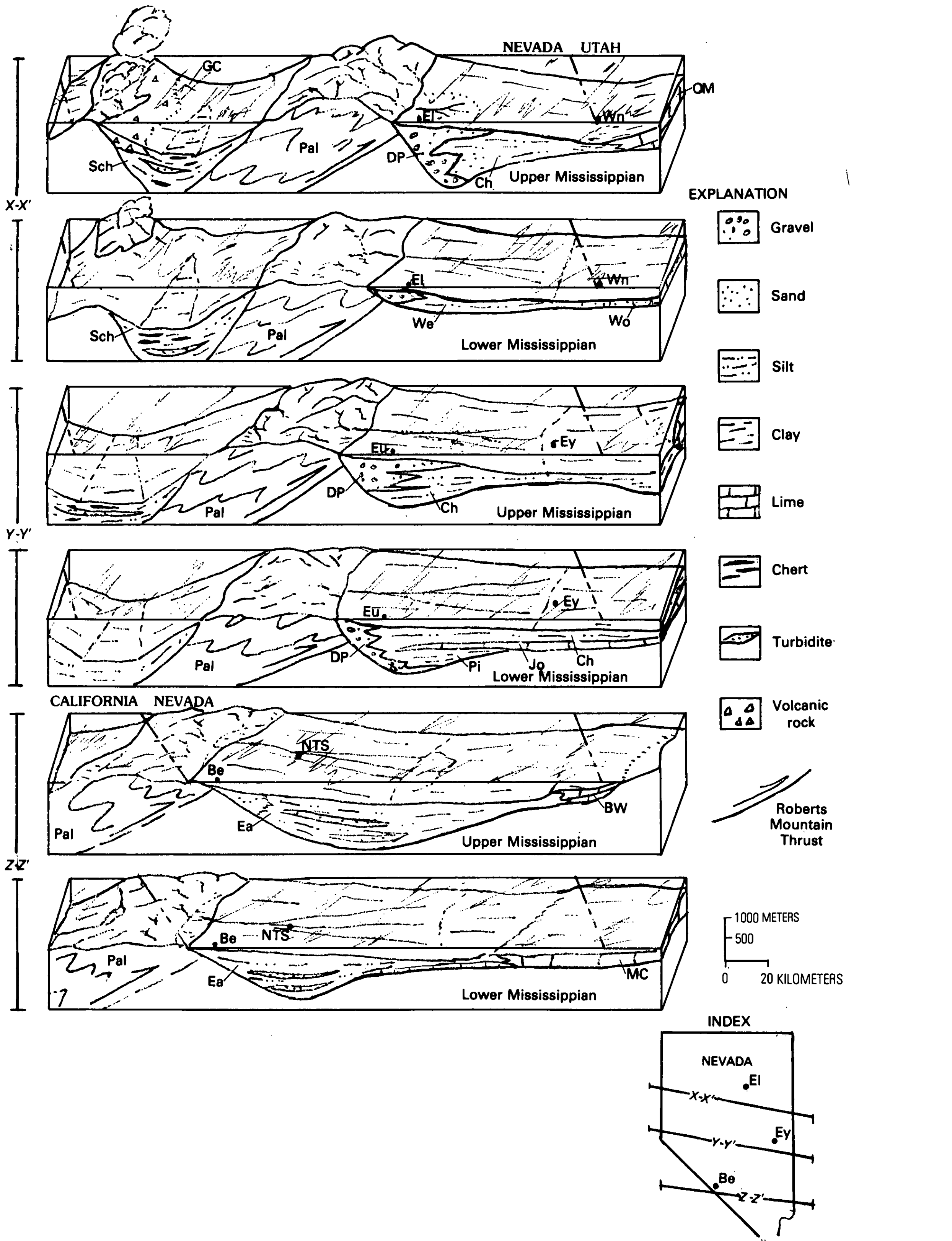


FIGURE 9.—Paleotectonic block diagrams of Mississippian rocks. Localities: Be, Beatty; El, Elko; Eu, Eureka; Ey, Ely; NTS, Nevada Test Site; Wn, Wendover Formation; BW, Battleship Wash Formation; Ch, Chainman Shale; DP, Diamond Peak Formation; Ea, Eleana Formation; GC, Gogh Canyon Formation; Jo, Joana Limestone; MC, Monte Cristo Formation; OM, Ochre Mountain Limestone; Pal, lower Paleozoic allochthon; Pi, Pilot Shale; Sch, Schoonover Formation; We, Webb Formation; Wo, Woodman Formation. (Diagram by E. R. Larson.)

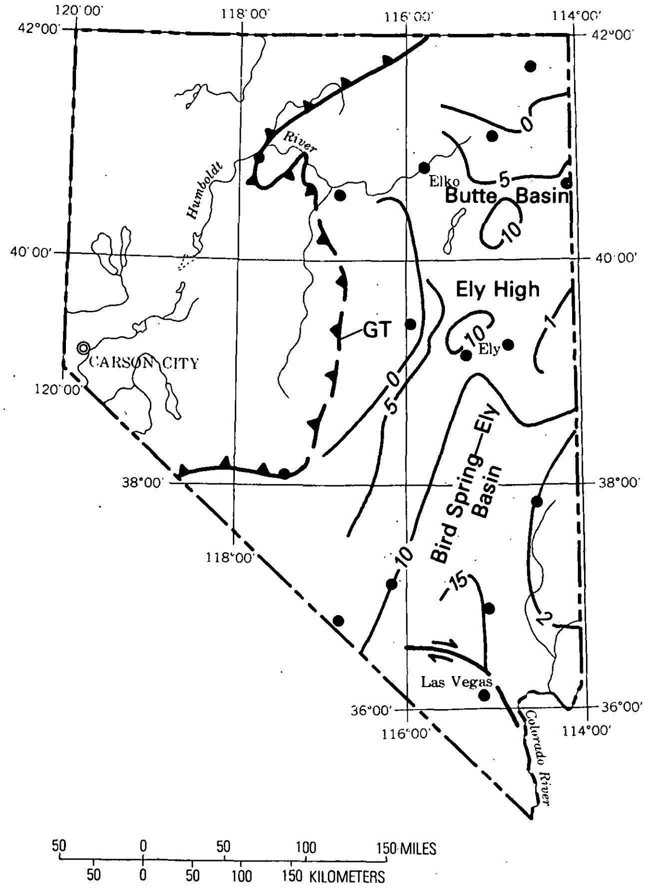


Figure 10.—Thickness of Pennsylvanian intervals A, B, C (Morrowan-Desmoinesian). Isopachs in hundreds of meters. GT, Golconda thrust. Modified from McKee and others (1975).

dle Atokan through Desmoinesian rocks to the west have been separately described as the Tomera Formation of the Ely Group (Dott, 1955). The underlying Morrowan and early Atokan Moleen Formation resembles the lower part of the Ely Limestone to the east.

Arenaceous and oolitic limestones, which are partially dolomitized, characterize the Callville Platform in southern Clark County and are assigned to the Callville Formation. Here, the Morrowan through Desmoinesian interval is broken by a hiatus representing Atokan time.

Pennsylvanian strata in north-central Nevada reflect the influence of the Antler orogenic belt. During the Mississippian and lower Pennsylvanian, this belt had acted as a barrier between sedimentary basins in eastern and western Nevada, but during the Atokan, the barrier was breached. The resulting islands were rugged enough to act as sources of coarse clastic material.

The Battle Formation, 730 m thick in the type section at Battle Mountain, has a 150-m basal conglomerate unit that reflects the composition of subjacent beds; a middle part consisting of *Fusulinella*- and *Chaetetes*-bearing limestone and conglomerate and ridge-forming sandstone; and an upper part of sandstone, siltstone, and conglomerate (Roberts, 1964, p. A29).

Lawson (1913) described these rocks as a fanglomerate, implying continental deposition. The formation as understood today represents marine deposition adjacent to a rugged source terrane. Recent studies by Drowley (1973) have shown that the beds included in the Battle Formation are isolated clastic accumulations deposited quite separately where gravel, boulders, and sand were being carried into a marine environment (fig. 12). To the west, the Battle Formation is overlain by the Highway Limestone of the same age, suggesting an interfingering relationship (fig. 12).

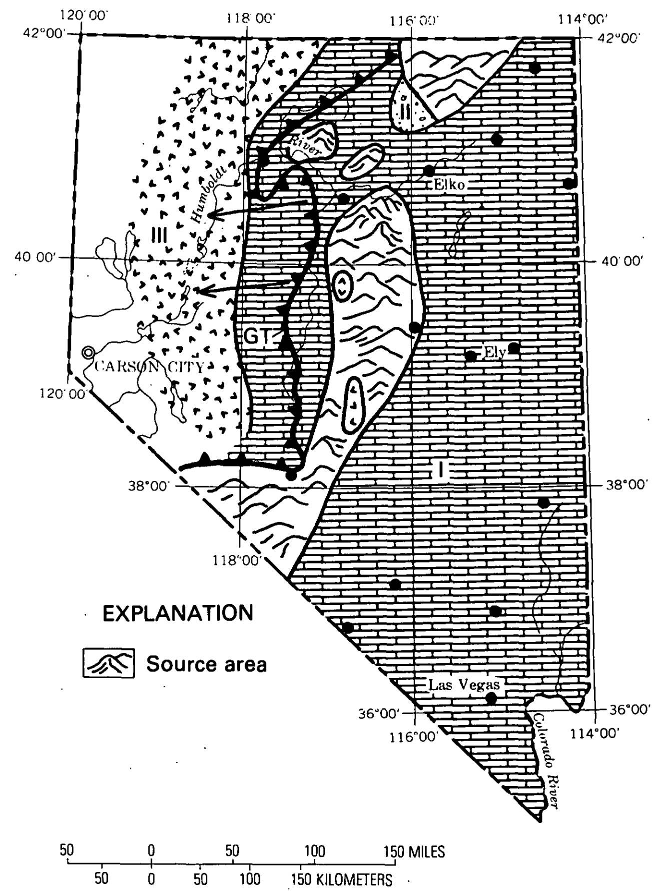


Figure 11.—Distribution of lithologies in Pennsylvanian intervals A, B, C (Morrowan-Desmoinesian). GT, Golconda thrust. Lithologies: I, limestone; II, conglomerate fan; III, eugeosynclinal rocks. Modified from McKee and others (1975).

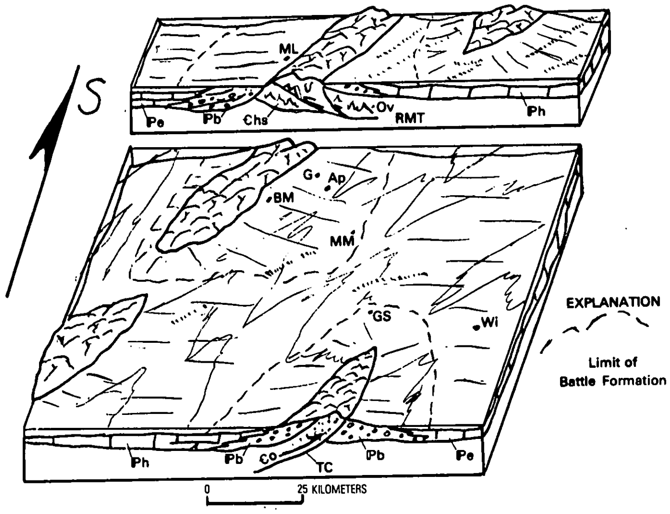


FIGURE 12.—Sites of deposition of Battle Formation. Localities: AP, Antler Peak; BM, Battle Mountain; G, Galena; GS, Golconda Summit; ML, Mount Lewis; Wi, Winnemucca. Formations: Ꞓhs, Cambrian Harmony and Scott Canyon formations; Ꞓo, Cambrian Osgood Mountain Quartzite; Ov, Ordovician Valmy Formation; ℙb, Pennsylvanian Battle Formation; ℙe, Pennsylvanian Etchart Limestone; ℙh, Pennsylvanian Highway Limestone; RMT, Roberts Mountain thrust; TC, Twin Canyon fault. (Diagram by E. R. Larson.)

To the east, pebbly limestone of the Moleen Formation (Dott, 1955) contains the *Chaetetes-Fusulinella* zone like the Battle and is its eastward equivalent.

INTERVALS D AND E

Desmoinesian rocks are separated from later Pennsylvanian rocks or from Permian rocks by a disconformity that is well documented throughout most of the miogeosyncline and shelf in Nevada. During the Missourian and Virgilian, the Butte Basin was divided by a northeast-trending positive area that shed sediments into the remaining part of the basin on either side. This positive area and the still-extant Ely Platform outlined a trough that extends from the Oquirrh Basin of Utah near Wendover southwestward toward central Nevada (fig. 13). The Strathern Formation, occupying the northwestern part of the Butte Basin, consists of chert-pebble conglomerate, silty limestone, and sandstone of Missourian through Wolfcampian age. It rests disconformably on Desmoinesian rocks of the Tomera Formation in the central part of the basin. Near Carlin, the Strathern forms a spectacular angular unconformity with the Diamond Peak Formation (Mississippian) and with pre-Strathern Pennsylvanian strata.

On the ridge between the basins, Wolfcampian rocks rest on formations as low in the section as Diamond Peak.

South of the Butte Basin on the Ely Platform, Missourian and Virgilian rocks are absent in White Pine and northern Lincoln Counties (fig. 13). Here the Wolfcampian Riepe Springs Limestone rests directly on Desmoinesian rocks. In the Bird Spring Basin, Missourian and Virgilian rocks of the Bird Spring Group, lithologically much like those below, are succeeded by Wolfcampian rocks variously assigned to either the Bird Spring Group or the Pakoon Formation. On the Callville Platform, Missourian rocks are absent in the Callville Formation, but the upper part of the formation is of Virgilian age. Wolfcampian rocks of the Pakoon Formation, a thick-bedded dolomite, succeed the Callville Formation disconformably.

Carbonate rocks also accumulated northwest of the Antler belt during Late Pennsylvanian and Early Permian time. Limestone-pebble and chert-

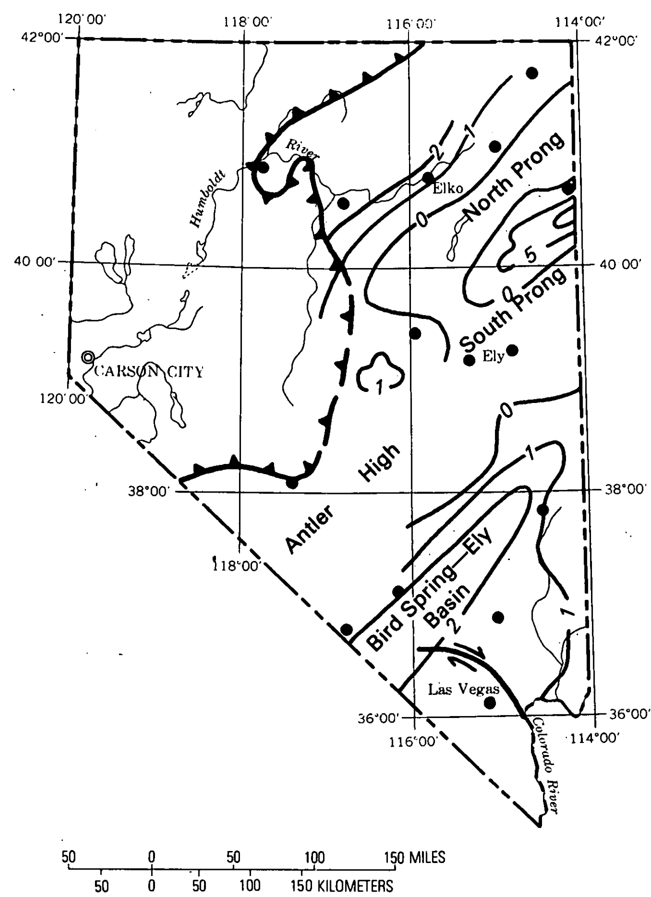


FIGURE 13.—Thickness of Pennsylvanian intervals D, E (Missourian-Virgilian). Isopachs in hundreds of meters. Modified from McKee and others (1975).

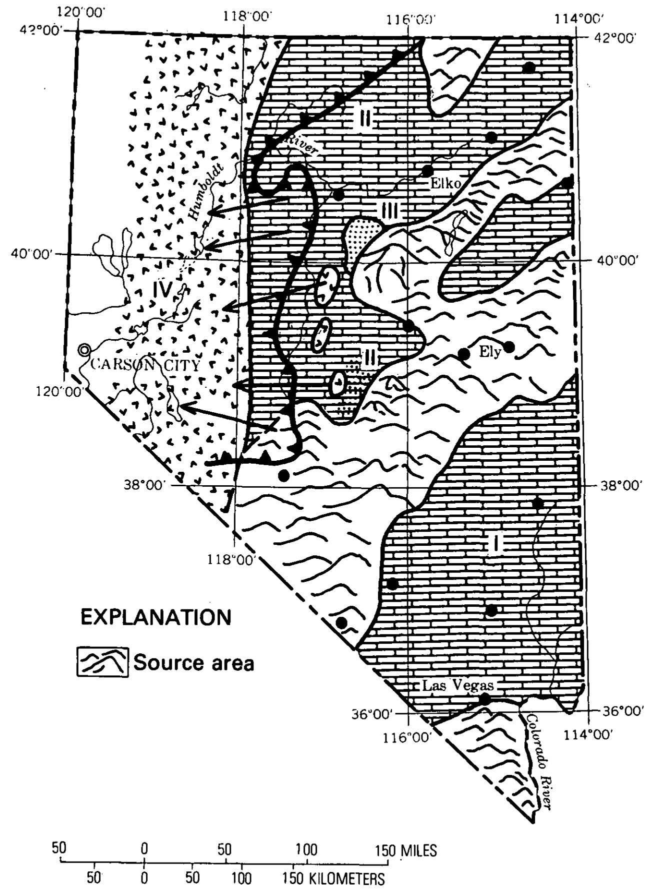


FIGURE 14.—Distribution of lithologies in Pennsylvanian intervals D, E (Missourian-Virgilian). Lithologies: I, limestone; II, conglomerate and limestone; III, sandstone; IV, eugeosynclinal rocks. Modified from McKee and others (1975).

pebble conglomerate and chert pebbly limestone (Wildcat Peak Formation) were deposited in the area of the Toquima Range west of Eureka (fig. 14). *Triticites* of possible Late Pennsylvanian age have been found in the lowermost beds of the formation (Kay and Crawford, 1964, p. 443; McKee, 1976, p. 23), but the formation has been dated as earliest Permian by Laule (1978) on the basis of a fusuline and coral fauna of Asiatic affinity.

PENNSYLVANIAN EUGEOSYNCLINE

Atokan through lower Permian sediments seen in klippen in the Battle Mountain-Winnemucca area (Roberts, 1964; Roberts and others, 1958; Ferguson and others, 1952; McMillan, 1974) and southward to Tonopah (Speed, 1971; Wilson, 1975) were deposited in a continuing eugeosynclinal system in western Nevada. These deposits of shale, chert, and volcanic rocks (Pumpernickel Formation), turbidites, and lesser limestone (Havallah Formation), originally thought to be sequential deposits (Ferguson and others, 1952), are now recognized as partly contemporaneous facies reflecting different source areas (figs. 4, 11, 13, 15). The poorly fossiliferous Pumpernickel Formation contains Atokan fusulines in limestone lenses, but the major lithology is bedded chert containing tubes and trails (lebensspoor) attributed to deep-water marine worms. Associated spilite is from submarine effusions or volcanic islands (fig. 15).

The Havallah Formation, which overlies the Pumpernickel in many areas, is Atokan to Leonardian. The graded arenite of the Havallah is believed to have been derived from the Antler belt and transported westward into an inner-arc basin (fig. 15). Figure 15 is a palinspastic synthesis of mid-Pennsylvanian relationships in northern Nevada.

The Carboniferous strata in Nevada reflect the tectonic development of the area that controlled the locus of source lands, sites of deposition, and the facies of the system. Continuing deformation is responsible for the present distribution of Carboniferous strata, and many of the problems of interpretation and reconstruction are still unresolved.

REPRESENTATIVE STRATIGRAPHIC SECTIONS

1. Monte Cristo-Bird Spring exposures on Nevada State Route 16, from crest of Spring Mountain Pass east to the valley floor. (Geologic map of Clark County (Longwell and others, 1965).)
2. A similar section along State Route 9, just northwest of the irrigated area in Muddy Valley.
3. Joana Limestone, White Pine Group, and Ely Limestone on U.S. Highway 50, beginning at Conners Pass and extending westward. (Connors Summit 15-minute quadrangle topographic map and White Pine County geologic map (Hose and Blake, 1976).)
4. A complete Joana-Chainman-Ely sequence on U.S. Highway 50 between Moorman Ranch and Antelope Summit. (Illipah 15-minute quadrangle topographic map and White Pine County geologic map (Hose and Blake, 1976).)
5. A well-exposed angular unconformity between the Diamond Peak Formation (Tonka of Dott, 1955) and the Strathern Formation at the west end of the tunnels on Interstate 80 east of Carlin. Post-Diamond Peak—pre-Strathern beds are exposed along old U.S. Highway 40 north of the Humboldt River.

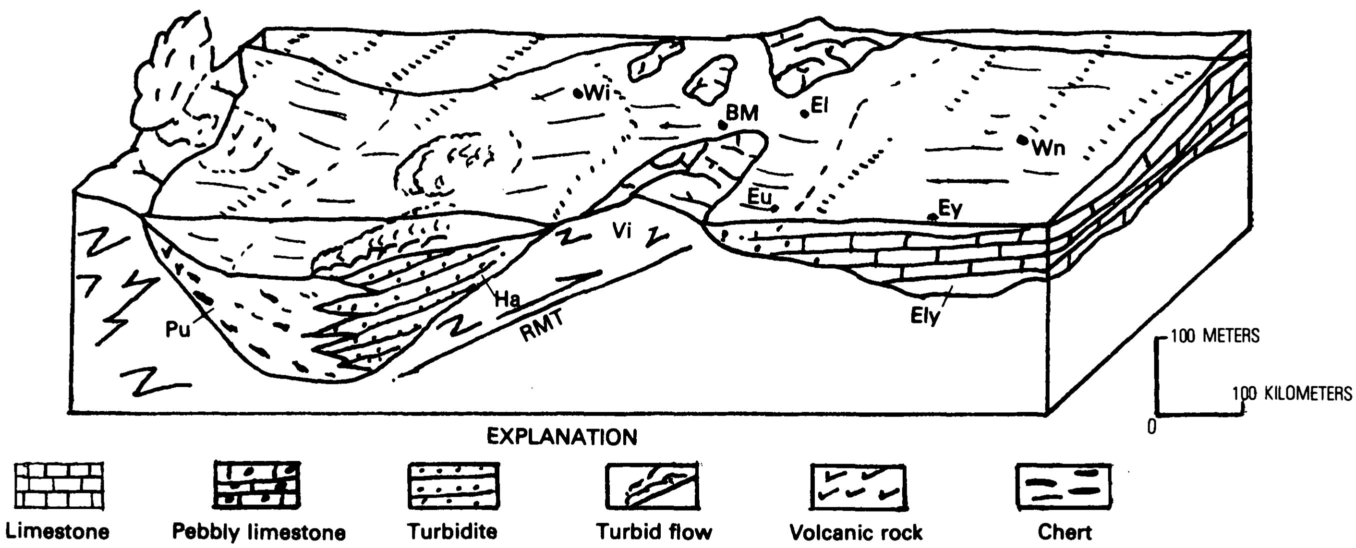


FIGURE 15.—Paleotectonic block diagram of Middle Pennsylvanian rocks. Localities: BM, Battle Mountain; El, Elko; Eu, Eureka; Ey, Ely; Wn, Wendover; Wi, Winnemucca. Formations: Ely, Ely Limestone; Ha, Havallah Formation; Pu, Pumpernickel Formation; Vi, Vinini Formation. RMT, Roberts Mountain thrust. (Diagram by E. R. Larson.)

6. Joana Limestone, Chainman Shale, and younger rocks on Interstate 80 east of Wells at Pequop Summit.
7. Pumpernickel and Havallah Formations at and near the Marigold mine. Take the Valmy turnoff approximately 15 miles (24 km) west of Battle Mountain (see Antler Peak geologic map (Roberts, 1964)). Pumpernickel chert and Havallah Formation together with turbidites in W. ½ Sec. 12, Havallah Formation Sec. 13. Badly silicified Battle Formation forms rusty ridges at the mine.
8. Antler Peak Limestone in good exposures in roadcuts on old U.S. Highway 40 west of Golconda Summit (SW, SW, Sec. 6, T. 35 N., R. 41 E.; Golconda Summit geologic map (Erickson and Marsh, 1974).)
9. Pumpernickel Formation exposed in borrow pits on the north side of Interstate 80, 1 mile (1.6 km) west of Golconda Summit.

PALEONTOLOGY

By JOSEPH LINTZ, JR.[3]

During the Carboniferous, sedimentation was nearly continuous in eastern Nevada, and all the widely recognized stages are represented (fig. 16). Poorly exposed eugeosynclinal assemblages in western Nevada are essentially nonfossiliferous, although scattered faunules have permitted a rough correlation of these beds with their eastern analogs.

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Most of the paleontologic discussion here is limited to the eastern part of the State; Carboniferous fossils have been relatively well studied there, and a considerable literature has accumulated. The fossils suggest a shelf area in southern Nevada and a shift to miogeosynclinal conditions farther north.

FAUNAL ZONES

The faunal zonation is shown in table 1. A few comments are in order. The Foraminifera are common throughout Nevada but are abundant on the craton in the southern part of the State. Douglass (1974) has given an excellent summary of fusulinid occurrences in Nevada. Local fusulinid faunules have been described by Cassity and Langenheim (1966), Rich (1961), and Slade (1961).

For the Mississippian stages, the ammonoids offer the best correlations for those parts of Nevada in deeper basins near the margins of the cratonic block (Gordon, 1970). Mississippian foraminifers, usually endothyrids (Brenckle, 1973), are also present and offer good control. The longer ranging corals, brachiopods, and bryozoans are frequently found, but their long ranges reduce their value as stratigraphic indicators.

For the Pennsylvanian stages, the fusulinids are nearly ubiquitous and are the most treasured stratigraphic indicators. Occasional marker horizons, such as the *Chaetetes-Rhipidomella* zone, are widely recognizable by the unusual abundance of the named taxa. Common faunal elements of less stratigraphic

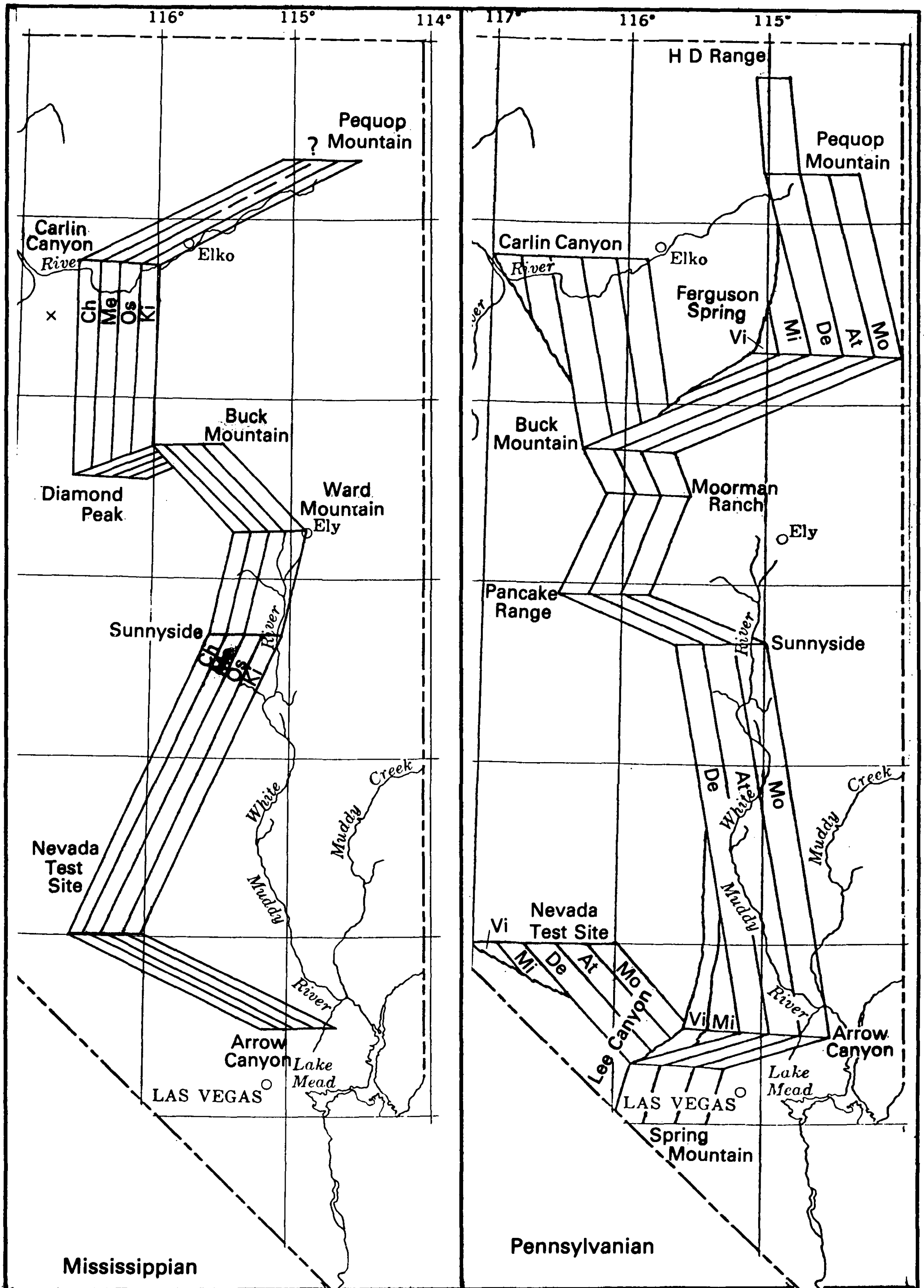

FIGURE 16.—Carboniferous stages in Nevada. Mississippian stages: Ki, Kinderhookian; Os, Osagean; Me, Meramecian; Ch, Chesterian. Pennsylvanian stages: Mo, Morrowan; At, Atokan; De, Desmoinesian; Mi, Missourian; Vi, Virgilian.

TABLE 1.—*Carboniferous index fossils of Nevada*

Period and Epoch	Foraminifers	Ammonoids	Conodonts	Other fossils
		Pennsylvanian		
Virgilian	*Triticites cullomensis* *T. kellyensis* *T. hobblensis* *T. plummeri* *T. ventricosus* *Dunbarellinella hughesensis*	----------	----------	----------
Missourian	*Triticites provoensis* *T. springvillensis* *Kansanella (Kansanella) grangerensis*	----------	----------	----------
Desmoinesian	*Fusulina megista* *Wedekindellina euthysepta*	----------	----------	*Hystriculina* spp. *Desmoinseia muricatina*
Atokan	*Fusulinella devexa* *F. acuminata* *F. fugax* *Profusulinella decora* *P. apodacensis* *P. copiosa*	*Pseudoparalegoceras kesslerense*	----------	*Multithecopora cannia* *Chaetetes favosus* *Komia* spp. *Chaetetes* sp. *Caninia torquia*
Morrowan	*Millerella marblensis* *Staffella expansa* *Paramillerella circuli*	*Gastrioceras branneri* *Diaboloceras* aff. *D. neumeieri* *Syngastrioceras* sp. *Stenopronorites* sp.	*Streptognathodus noduliferus* *Idiognathoides convexa* *I. parva*	*Caninostrotion* spp. *Hustedia miseri* *Rhipodomella nevadensis* *Flexaria* spp.
		Mississippian		
Chesterian	*Atesuella meandra* *Plectogyra* sp.	*Cravenoceras hesperium* *C. merriami* *Eumorphoceras bisulcatum* *Mooreoceras* sp.	*Gnathodus bilineatus.* *G. girtyi simplex*	*Faberophyllum* sp. *Ekvasophyllum* spp.
Meramecian	*Endothyra symmetrica* *Plectogyra* spp. *Granuliferella* spp. *Endothyra spiroides*	----------	*Taphrognathus varians* *T. varians/ Polygnathus mehli* *Eotaphrus burlingtonensis* *Polygnathus communis communis/ Eotaphrus* sp.	----------
Osagean	*Plectogyra tumula* *P. anteflexa* *Endothyra* spp.	----------	*Pseudopolygnathus triangulus nudus* *Gnathodus bilineatus/ G. cuneiformis* *G. semiglaber/ Pseudopolygnathus marginatus* *Polygnathus communis communis*	----------
Kinderhookian	*Granuliferella* spp.	*Protocanites lyoni*	*P. communis communis/ Pseudopolygnathus dentilineatus*	*Torynifer* aff. *T. cooperensis*

value in the Pennsylvanian are the associations of corals, brachiopods, and bryozoans.

Pennsylvanian mollusks, especially ammonoids, are less frequently found than are those of the Mississippian. Gastropods are present in moderate quantity, but pelecypods are less common. Gordon (1970) has reviewed Carboniferous ammonoid zonation in the Western United States, and Gordon and others (1957) have established goniatite zones in the Chainman Shale equivalents of western Utah.

Conodonts have been recognized in both argillaceous and carbonate rocks in eastern Nevada and in the generally nonfossiliferous allochthonous eugeosynclinal sequences in central Nevada. They have proved invaluable in regional correlation and classification.

Carboniferous conodont faunas in Nevada have been described by Dunn (1965), Pierce and Langenheim (1974), and Rice and Langenheim (1974). Webster (1969) showed the relation of conodonts to fusuline zones.

Brachiopods are typical of the Carboniferous. A representative collection will contain many of the major taxa, including spiriferids, productids, rhynchonellids, and strophomenids. Brachiopods are almost always the dominant forms of any collection from either Mississippian or Pennsylvanian sedimentary rocks.

Trilobites are poorly represented in the Carboniferous faunas of Nevada. After the abundance of phacopids in the Devonian, trilobites were greatly reduced in numbers; a gradual diminution took place throughout the Mississippian and Pennsylvanian and into the Permian. Carboniferous trilobites from Nevada are members of the less exotic, more conservative stocks, such as the proetids.

Echinodermata are abundant but almost always consist of the ubiquitous columnals. Calices are rarities, except in the cratonic area, where Webster and Lane (1970) have described several genera and species. *Pentremites*, so abundant in the Mississippian of the Mississippian River valley, is essentially unknown in Nevada.

SELECTIVE LIST OF COLLECTING LOCALITIES

The purpose of this list is to enumerate some localities where representative collections might be made by interested persons. In selecting the localities, accessibility has been considered. Thus, localities within the Nevada Test Site are omitted because access to that rather large area of south-central Nevada is restricted. The order of presentation is generally from south to north along the eastern part of Nevada.

Spring Mountains (SW. Clark County).—Spring Mountains, the type locality of the Bird Spring Formation, is northeast of the town of Goodsprings and is best reached over unpaved roads from the Blue Diamond-Arden Highway. A complete sequence of Pennsylvanian fusulinids is available, plus other common invertebrates that have yet to be catalogued in the literature. (Goodsprings 15-minute quadrangle topographic map.)

Arrow Canyon (E. Clark County).—Arrow Canyon has become well known through the field studies of R. L. Langenheim, G. D. Webster, N. G. Lane, and their students. Located on the Arrow Canyon 15-minute quadrangle topographic map, the area has easy access from Las Vegas and contains a superior sequence of fossils from Kinderhookian into the Wolfcampian. Ample papers have described the sites and the fauna in detail (Langenheim and Langenheim, 1965).

Shingle Pass (NW. Lincoln County).—Shingle Pass, a less accessible site in the southern Egan Range between State Route 38 and U.S. Highway 93, is near Sunnyside at Whipple Ranch. Good Pennsylvanian exposures are on the north side of the pass. This site has a good Pennsylvanian fauna (Kellogg, 1963) of fusulines and of generally undescribed larger invertebrates (Shingle Pass 7½-minute topographic map.)

Dutch John Mountain (N. Lincoln County in T. 8 N., R. 65 E., projected).—Dutch John Mountain is one of the best localities for Mississippian faunas, being accessible from U.S. Highway 93, approximately 12 miles (20 km) south of Geyser maintenance station. This site is on the geologic map of Lincoln County (Tschanz and Pampeyan, 1970).

Ward Mountain (central White Pine County).—On the outskirts of Ely, Ward Mountain has a complete Mississippian to mid-Pennsylvanian-Desmoinesian section. Lithologies are typical for eastern Nevada. The Mississippian black shale is poorly fossiliferous, but good collections can be obtained from the Joana Limestone (Mississippian) and Ely Limestone (Pennsylvanian). (Ely 15-minute quadrangle topographic map and White Pine County geologic map (Hose and Blake, 1976).)

Duckwater (NE. Nye County).—Duckwater is mentioned because of the *Cravenoceras* faunule (Chesterian) described by Youngquist (1949). Very accessible black-shale exposures are east of Duckwater. (Duckwater 15-minute quadrangle topographic map.)

Buck Mountain (W. White Pine County).—Buck Mountain has fossils ranging in age from Kinderhookian to Desmoinesian. It is reached by a gravel road leading north from U.S. Highway 50, 3 miles (4.8 km) east of Pancake Summit. (Buck Mountain and Pancake Summit 15-minute quadrangle topographic maps and geologic map of White Pine County (Hose and Blake, 1976).)

Carbon Ridge (W. White Pine County).—Carbon Ridge is one of the better exposures of the Diamond Peak conglomerate and its Chesterian faunule. The preservation is good, although fossils have a tendency to split between shell layers. Accessibility is excellent. Turn south off U.S. Highway 50 onto State Route 20, proceed about 1 mile (1.6 km), then turn right onto the gravel road leading up Secret Canyon. A little more than 2 miles (3.2 km) from the turnoff at the cattle guard are conspicuous hogbacks of chert pebble conglomerate. The fossils are in the softer sedimentary rocks between the conglomerates. (Pinto Summit 15-minute quadrangle topographic map and Nolan and others, 1974.)

Carlin Canyon (SW. Elko County).—Fossil localities in the Carlin Canyon are numerous, both Mississippian and Pennsylvanian Systems being well represented. The structures are somewhat complicated by extensive faulting, so that beds are seldom continuous over any distance. Conglomerates predominate, and one should seek the finer clastic rocks and limestone. The area is adjacent to Interstate 80 and is readily reached via old U.S. Highway 40. Carlin Canyon extended would include Grindstone Mountain. This place is one of the few in northern Nevada that has Missourian and Virgilian faunas. (Carlin and Dixie Flats 15-minute quadrangle topographic maps (Dott, 1955; Smith and Ketner, 1975).)

South Doby Summit (Elko County).—The well-known locality at South Doby Summit is 7 miles (11.2 km) north of Elko on State Route 51. On the right side of the road is a black-shale borrow pit containing a very good Chesterian faunule. The shale is soft and the fossils are silver hued, producing showy specimens. However, the specimens are as fragile as the shale, and care must be taken to prevent a high attrition rate. (Elko West 7.5-minute quadrangle topographic map and Smith and Ketner, 1975.)

Pequop Mountains (central Elko County).—Twenty-two miles (35 km) east of Wells, Interstate 80 crosses the Pequop Mountains. At the summit are excellent Mississippian and Pennsylvanian faunules. The literature contains no report of the presence of the middle Mississippian, but Kinderhookian and Chesterian fossils and a complete Pennsylvanian section have been reported. (Pequop Summit 7.5-minute quadrangle topographic map and Thorman, 1970.)

Ferguson Springs (E. Elko County).—Some 14 miles (22.4 km) south of Wendover on U.S. Highway 50 Alternate is a highway maintenance station. To the west, on the east flank of the Goshute Mountains, are Pennsylvanian and Permian limestones. The Pennsylvanian is complete here. Access is easy and the excellent fauna includes fusulinids, brachiopods, bryozoans, rugose and tabulate corals, and mollusks. (Ferguson Mountain 7.5-minute quadrangle topographic map.)

NOTABLE FOSSIL TAXA

Like every other region, Nevada has unusual finds from Carboniferous sedimentary rocks. Any listing of such a group is inescapably arbitrary and subjectively incomplete.

Dimegalasma eurekensis.—*Dimegalasma eurekensis,* from the Chesterian Diamond Peak Formation and equivalents, is possibly the largest spiriferid known. G. A. Cooper, upon receipt of the holotype at the U.S. National Museum, said that it was the second largest brachiopod he had seen (Lintz and Lohr, 1958).

Komia eganensis.—*Komia* was originally described by Korde (1951) as an alga. Wilson and others (1963), on the basis of their work describing *Komia eganensis,* using specimens from Nevada, concluded that the genus *Komia* should be assigned to the Stromatoporoidea. *Komia eganensis* is known from Arrow Canyon, Sunnyside, Ward Mountain, and Moorman Ranch, occurring in association with *Fusulinella acuminata* about 50 m above the *Profusulinella/Chaetetes* marker horizon (Wilson and others, 1963).

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—California, Oregon, and Washington

By RICHARD B. SAUL, OLIVER E. BOWEN, CALVIN H. STEVENS, GEORGE C. DUNNE, RICHARD G. RANDALL, RONALD W. KISTLER, WARREN J. NOKLEBERG, JAD A. D'ALLURA, ELDRIDGE M. MOORES, RODNEY WATKINS, EWART M. BALDWIN, ERNEST H. GILMOUR, *and* WILBERT R. DANNER

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Historical review and summary of areal, stratigraphic, structural, and economic geology of Mississippian and Pennsylvanian rocks in California, Oregon, and Washington

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ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—CALIFORNIA, OREGON, AND WASHINGTON

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ABSTRACT

Rocks of Carboniferous age in California, Oregon, and Washington are described in seven separate reports. These rocks are a part of the sedimentary record in the Cordilleran geosyncline and in what appears to have been an adjacent, tectonically active ocean basin to the west and northwest.

Rocks deposited near the axis of the geosyncline are described in two reports on exposures in southeastern California. One report covers a group of economically important, metamorphosed, sparsely fossiliferous carbonate-rich sequences in the southwestern Mojave Desert. The other contains data on a well-preserved record of largely carbonate deposition on an eastern cratonic shelf and in a more western foreland basin, in the area east and southeast of the Sierra Nevada. Here, facies changes and unconformities appear to reflect tectonic events in an orogenic highland to the north and perhaps west of that region.

In the eastern Sierra Nevada, metamorphosed Carboniferous sedimentary rocks are exposed discontinuously in roof pendants. Fossils are rare. Correlation is mainly based on similarity of lithology to the nearest fossiliferous rocks. Regional differences in strike between these rocks and rocks of similar character and age to the east suggest a large-scale tectonic dislocation or regional angular unconformity. In the north-central Sierra Nevada, the Carboniferous section is predominantly a chert-epiclastic sequence and is sparsely fossiliferous.

In the eastern Klamath Mountains of California, the Carboniferous is mainly argillite and volcaniclastic rocks containing minor amounts of chert, greenstone, limestone, and siltstone. The sparse faunas, which range in age from late Mississippian through Pennsylvanian to early Permian, are part of an Asiatic province and have few species in common with typical North American faunas.

In Oregon, sedimentary rocks of Carboniferous age are known near the head of Crooked River in the southern Blue Mountains. Two formations are named. The older, composed mainly of limestone, is middle and upper Meramecian and Chesterian. The younger, of Morrowan age, is composed of conglomerate, sandstone, and plant-bearing mudstone. In northeastern Oregon, in an argillite, a fusulinid fauna is assigned to the Desmoinesian. In Oregon, all Carboniferous strata are probably allochthonous, having been emplaced during plate movements.

Scattered Carboniferous rocks in western and eastern Washington range in age from Chesterian to Desmoinesian and possibly Missourian. Some of the Carboniferous rocks in western Washington contain Tethyan faunas and may have been tectonically transported north and east from the Permian equatorial area. Fossils in the Carboniferous rocks of northeastern Washington appear to have eastern North American or Rocky Mountain affinities.

INTRODUCTION

By RICHARD B. SAUL

THE CORDILLERAN GEOSYNCLINE

The Grand Canyon of the Colorado River is one of the most impressive natural features on the North American Continent. It is 1,500 to 1,800 m deep. In its walls the Paleozoic Era is represented by about 1,300 m of sedimentary rocks (fig. 1). About 400 km west of the Grand Canyon, in the Panamint Range of California, is a thicker, more

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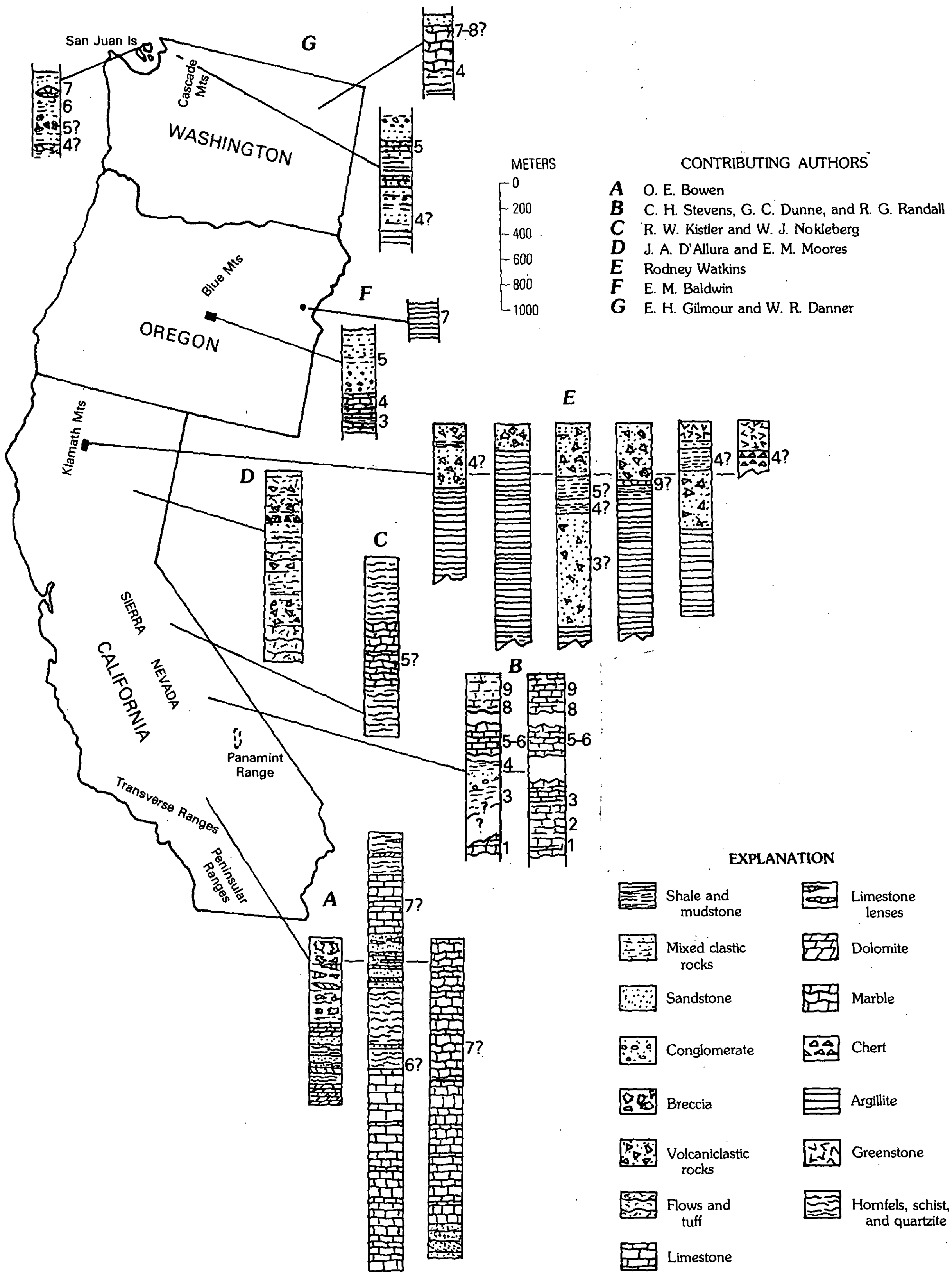
San Juan Is
G
Cascade Mts
WASHINGTON
Blue Mts
F
OREGON
Klamath Mts
D
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SIERRA
NEVADA
CALIFORNIA
Panamint Range
Transverse Ranges
Peninsular Ranges
A
B
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EXPLANATION
Shale and mudstone
Mixed clastic rocks
Sandstone
Conglomerate
Breccia
Volcaniclastic rocks
Flows and tuff
Limestone
Limestone lenses
Dolomite
Marble
Chert
Argillite
Greenstone
Hornfels, schist, and quartzite

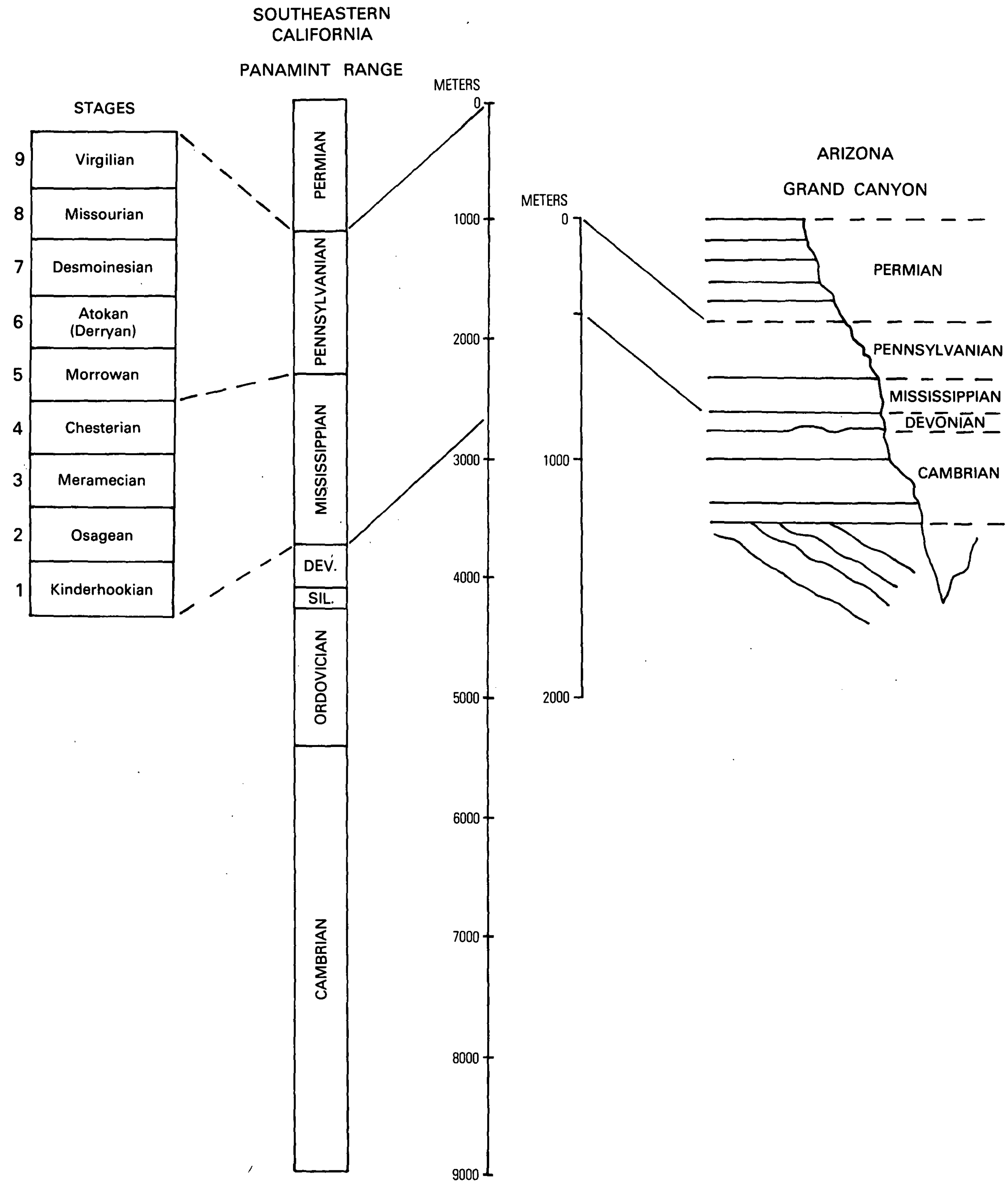

FIGURE 1.—Regional aspects of the Carboniferous rocks in and west of the axis of the Cordilleran geosyncline in California, Oregon, and Washington. Because of the uncertainties of many age assignments, the commonly altered or unfossiliferous character of the rocks, and the scale of this figure, the thicknesses are approximate and the lithology is generalized. The upper right-hand columns at F and G are localities, not described sections.

complete Paleozoic section but one that is faulted and folded. If undisturbed, the section in the Panamint Range would require a canyon about 9,000 m deep for full exposure. Indeed, in California, if the entire conformable sedimentary sequence, including the late Precambrian, were to be considered, the section would be more than 15,000 m thick. Unfortunately such an unbroken sequence of Paleozoic rocks is not exposed in California, Oregon, or Washington, although the section in southeastern California can easily be pieced together. There are, instead, widely scattered exposures of faulted, deformed, and in places, intruded and metamorphosed parts of the sequence. Therein has lain some of the difficulty facing Paleozoic stratigraphers in general and students of the Carboniferous in particular in the Cordilleran geosyncline.

EARLY WORK

Careful examination of the rocks of the region began in the late 1800's. By the turn of the century, a base of useful literature had begun to accumulate. Because of the wide variety in the age and type of rocks exposed in California, Oregon, and Washington, these areas have attracted a large and diverse cadre of earth scientists. The literature has accrued exponentially. Most of the early work was done through the aegis of the U.S. Geological Survey. Descriptions of the rocks of Carboniferous age are minor parts of the older publications. The early work usually centered on problems of ground-water supplies and the geology of mining districts. The abundant carbonate rocks were good hosts to ores of both base and noble metals. In 1912, James M. Hill published index maps of topographic and geologic surveys of the Western States that illustrate the influence of the mining districts on those endeavors. Not until recent years have workers chosen to study the rocks and rock-forming events of specific Periods in the Paleozoic.

As topographic maps became available for certain districts, detailed geologic studies of specific quadrangles were published. One such work, involving well-exposed rocks of Carboniferous age, is "Geology and ore deposits of the Goodsprings quadrangle, Nevada" by D. F. Hewett (1931). This effort to assist the development of a mining district is a classic of structural and stratigraphic exposition. Through such investigations, great progress was made in understanding the geologic history of the Cordilleran geosyncline.

In the eastern Transverse Ranges of California and in the regions to the east and south, early descriptions of Paleozoic rocks were generally part of academically oriented geologic mapping. However, such reports contributed to the search for industrial minerals. The work by Vaughan (1922) in which he described the Furnace Limestone is an example.

Knopf and Kirk (1918) were among the earliest to describe the rocks and mineral resources of the thick Paleozoic section along the east flank of the Sierra Nevada. Some of the best early work in the Sierra Nevada and Klamath Mountains of California was conducted by Diller and is presented in U.S. Geological Survey Bulletins and Folios, and in professional journals.

In 1902, Condon wrote a geologic history of Oregon in which Paleozoic rocks were described, but most of the Carboniferous rocks were located by E. L. Packard (1928, 1932) and his students.

The Paleozoic section in western British Columbia extends southward, to a limited extent, into the State of Washington. Most of the earliest work was done by Canadians such as Clapp (1909). In 1927, McLellan published a report on the San Juan Islands, in northwestern Washington, which describes structural, stratigraphic, and economic aspects of the geology. Here, as elsewhere in the Western States, significant geologic research emerged around the turn of the century.

DISTRIBUTION OF THE ROCKS

In the three States covered by this report, the least altered, most fossiliferous, and most complete sections of Carboniferous rocks are exposed in the Basin Ranges east of the Sierra Nevada in California. Near and south of the latitude of the Transverse Ranges of California (fig. 1), rocks of probable or possible Carboniferous age are generally altered and difficult to date, at least in part because of the emplacement of Mesozoic plutonic rocks. Similarly, rocks of Carboniferous age are sparsely distributed in roof pendants in the batholithic rocks of the Sierra Nevada. A section of metamorphosed sedimentary rocks of late Paleozoic age, some of which may prove to be Carboniferous, forms a belt in the western foothills of the Sierra Nevada. A mixed sequence of sparsely fossiliferous clastic, volcaniclastic, and carbonate rocks of Carboniferous age is exposed in the eastern Klamath Mountains in northwestern California.

Rocks ranging in age from Devonian through Permian have been mapped in a faulted and folded section in the Blue Mountains of central Oregon. There the Carboniferous comprises fossiliferous, mixed calcareous, clastic, and argillaceous rocks and minor amounts of limestone overlain by conglomerate, wacke, and plant-bearing mudstone. An argillite of possible Pennsylvanian age is to the northeast, near Baker.

In the State of Washington, Carboniferous rocks are found in two different sections, both of which are more extensively exposed to the north in British Columbia. One belt is in the Northern Cascade Range; the other is exposed in the San Juan Islands and in the foothills of the Northern Cascades and Middle Cascades on the mainland (Danner, 1977). The rocks in the Cascade Range consist of argillite and siltstone of Mississippian?-Pennsylvanian age overlain by fossiliferous Pennsylvanian limestone. The section in the San Juan Islands comprises volcanic flows and breccia, shale, siltstone, tuff, graywacke, lenses of fossiliferous limestone, and a few thin seams of low-grade coal.

ECONOMIC IMPORTANCE

The presence of coal strata in the Eastern States may have caused early expectations of coal in the Carboniferous of the West. Although thin local seams are present in the Pennsylvanian of the San Juan Islands, in the State of Washington (McLellan, 1927), no commercially exploitable deposits have been found. The role of Carboniferous carbonate rocks as host for metallic ores has been noted. The growth of population on the Pacific Coast has increased demand for carbonate rocks for industrial purposes. Much of the information on deposits of known or potential value in Carboniferous rocks is produced or collated by State agencies: in California, the California Division of Mines and Geology, Department of Conservation, Resources Agency; in Oregon, the State of Oregon, Department of Geology and Mineral Industries; and in Washington, Washington State, Department of Natural Resources, Division of Geology and Earth Resources.

PRESENT RESEARCH

An excellent summary of current research exists in a symposium volume titled "Paleozoic paleogeography of the western United States" (Pacific Coast Paleogeography Symposium, 1977). Because that symposium involved or influenced most of the contributors to this work, a rough summary is presented as follows: The axis of the Cordilleran geosyncline had a north-northeasterly trend. The axis lay athwart what is now southeastern California, resulting in the thick accumulation of Paleozoic marine rocks in that region. The thinner, Grand Canyon section accumulated on the cratonic shelf along the southeast margin of the geosyncline. To the northwest, in the region of central California, an orogenic highland belt from which sediments were derived trended northeastward. A trench lay along the southeast margin of the highland. The boundary between the trench and the highland is described as a system of northwest-dipping thrust faults. To the northwest, through northern California, Oregon, and Washington, carbonate rocks become subordinate to clastic, volcaniclastic, and volcanic rocks. This may have been a region of island arcs in or west of an ocean basin adjacent to and west of the Cordilleran geosyncline. Across the western basin, rocks and fossil faunas from a different province appear to have been rafted against the continental plate above an east-dipping subduction zone.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the California Division of Mines and Geology, Department of Conservation, Resources Agency; the State of Oregon, Department of Geology and Mineral Industries; and Washington State, Department of Natural Resources, Division of Geology and Earth Resources.

CARBONIFEROUS AND PROBABLE CARBONIFEROUS METASEDIMENTARY ROCKS IN THE SOUTHWESTERN MOJAVE DESERT, CALIFORNIA

By OLIVER E. BOWEN

ABSTRACT

This paper deals with partly metamorphosed late Paleozoic rocks in the southwestern Mojave Desert. In the light of the relatively abundant faunas in the Carboniferous rocks of eastern California, a review of the lithology and gross stratigraphy of these sparsely fossiliferous rocks seems in order.

The probable Carboniferous rocks in the southwestern Mojave Desert have the following things in common:

1. The lithofacies consist predominantly of quartz-mica schist, limestone and quartzite.
2. Fossils are widely scattered and the preservation is seldom better than fair.

3. The Carboniferous rocks occur as small or large pendants suspended in granitic rocks.
4. In most pendants the stratigraphic section is incomplete.
5. In general the rocks are well exposed.

The most detailed studies have been done on the Oro Grande series which crops out in many pendants found within or close to the quadrilateral determined by the cities of Victorville, Kramer Junction, Barstow, and Lucerne Valley. The maximum thickness of the Oro Grande series is 2,952 m, whereas the type section of Quartzite Mountain is only 723 m thick.

The Oro Grande series is believed to correlate with the Furnace Limestone and the Chicopee and Sarrogossa Quartzite formations of the San Bernardino Mountains. These formations aggregate 3,409 m; the Furnace Limestone makes up 2,447 m of the total.

Limestone is by far the most important industrial raw material quarried in the southwestern Mojave Desert, and the portland cement industry is the largest consumer. Quartzite is quarried extensively for use in portland cement, railroad ballast, road base, and allied construction materials. Other products such as magnesite, gold, silver, copper, and tungsten have been produced intermittently.

HISTORY

Before publication by the author of "Geology and mineral deposits of Barstow quadrangle" (Bowen, 1954) and "Mineral deposits in the Oro Grande series" (Bowen and Ver Plank, 1965), the only reasonably detailed description of the probable Carboniferous rocks in the southwestern Mojave Desert was by W. J. Miller in 1944. Miller briefly described lithologic units at the type section of the Oro Grande series but found no diagnostic fossils. Brief references to the lithology of the late Paleozoic rocks are to be found in Hershey (1902), Darton (1915), and Pack (1914). A good description of the Furnace Limestone and associated quartzite formations was published by F. E. Vaughan in 1922. Woodford and Harris supplemented Vaughan's work in 1928. Both of these papers discuss the probable Carboniferous sequence in the San Bernardino Mountains immediately adjoining the Barstow 30-minute quadrangle to the south. The same area was discussed in a short paper by Guillou in 1953 and in a more extensive one by Richmond in 1960. Richmond's paper contains the best description of the stratigraphy and fossil record of the late Paleozoic rocks of any dealing with the San Bernardino Mountains. Weise (1950) discussed probable Paleozoic rocks in the Neenach quadrangle.

Between 1960 and 1965, geologic maps of four 15-minute quadrangles were released by the U.S. Geological Survey covering the entire older Barstow 30-minute quadrangle (Dibblee 1960a, b, c, d). As these map sheets were preliminary, very little discussion of the map units was possible (fig. 2).

GEOLOGIC SETTING

All the areas being discussed are underlain by granitic rocks; the metasedimentary rocks in general are found in pendants surrounded by granitic rocks. At no place is the base of the probable Carboniferous section exposed, and the basal contact for any one sequence is along an instrusive rock, or in a few places, a fault.

In two places, at least, the probable Carboniferous rocks are overlain by metasedimentary rocks containing diagnostic Permian fossils. Near Sidewinder Mountain, west of the Reserve quarry of Southwestern Portland Cement Company, the Permian Fairview Valley Formation rests nonconformably on the Oro Grande series. In the Goler Wash vicinity of the El Paso Mountains, fusulinid-bearing Permian rocks are conformable on a sequence of rocks that may be partly Carboniferous and partly Permian.

The lithology of the probable Carboniferous rocks (fig. 3) reflects few diastrophic events. Limestone tends to be pure, clay shale has little or no coarse sand or pebbles, and the quartzite, except in the El Paso Range, is made up of clean sand. These rocks most probably result from intense chemical weathering in a warm, humid climate.

Sea-floor dolomitization was predominant, although some dolomite may have formed by introduction of magnesium mobilized at the time of influx of the granitic rocks.

Coarse conglomerate and sedimentary breccia, probably reflecting mountain-building activity, appear in the Permian and probably in the Pennsylvanian as well, as would be expected in concluding epochs of the Paleozoic. This characteristic is in keeping with events reflected in the Permian-Carboniferous sequence in eastern California.

LITHOLOGY AND LITHOLOGIC SEQUENCE

The only detailed, measured sections of probable Carboniferous age are in either the Barstow 30-minute quadrangle or the adjacent Shadow Mountains 15-minute quadrangle (Bowen, 1954). In the Shadow Mountains, the Oro Grande series reaches a thickness of 3,414 m and is primarily a crystalline limestone-quartz-mica schist sequence. Thin quartzite and hornfels layers are present within the schist, but they do not form mappable units except at a large scale. The much thinner (1,285 m) type sec-

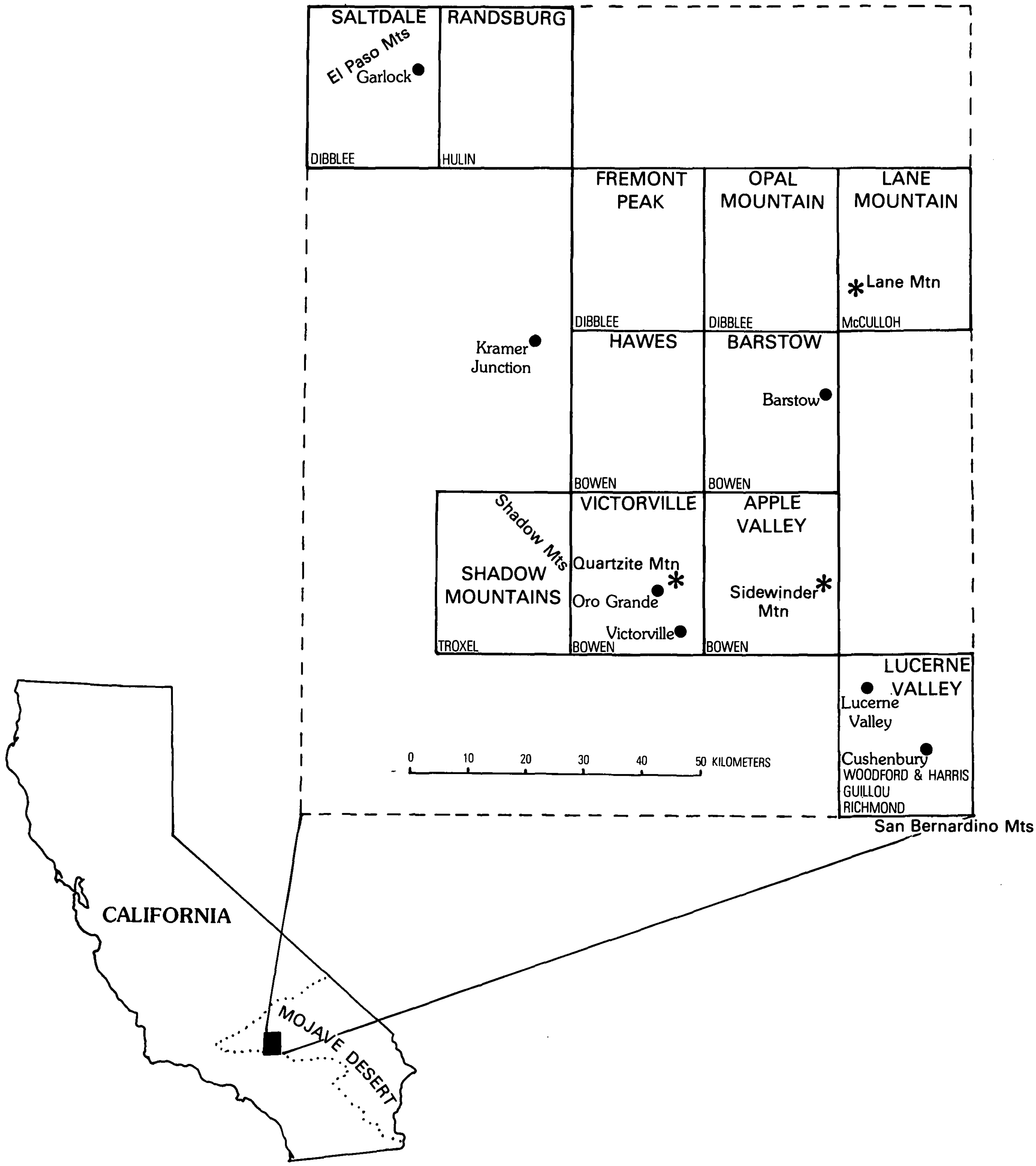

FIGURE 2.—Location of quadrangles in which Carboniferous rocks are exposed. Names of authors responsible for mapping appear in southwest corner of quadrangle.

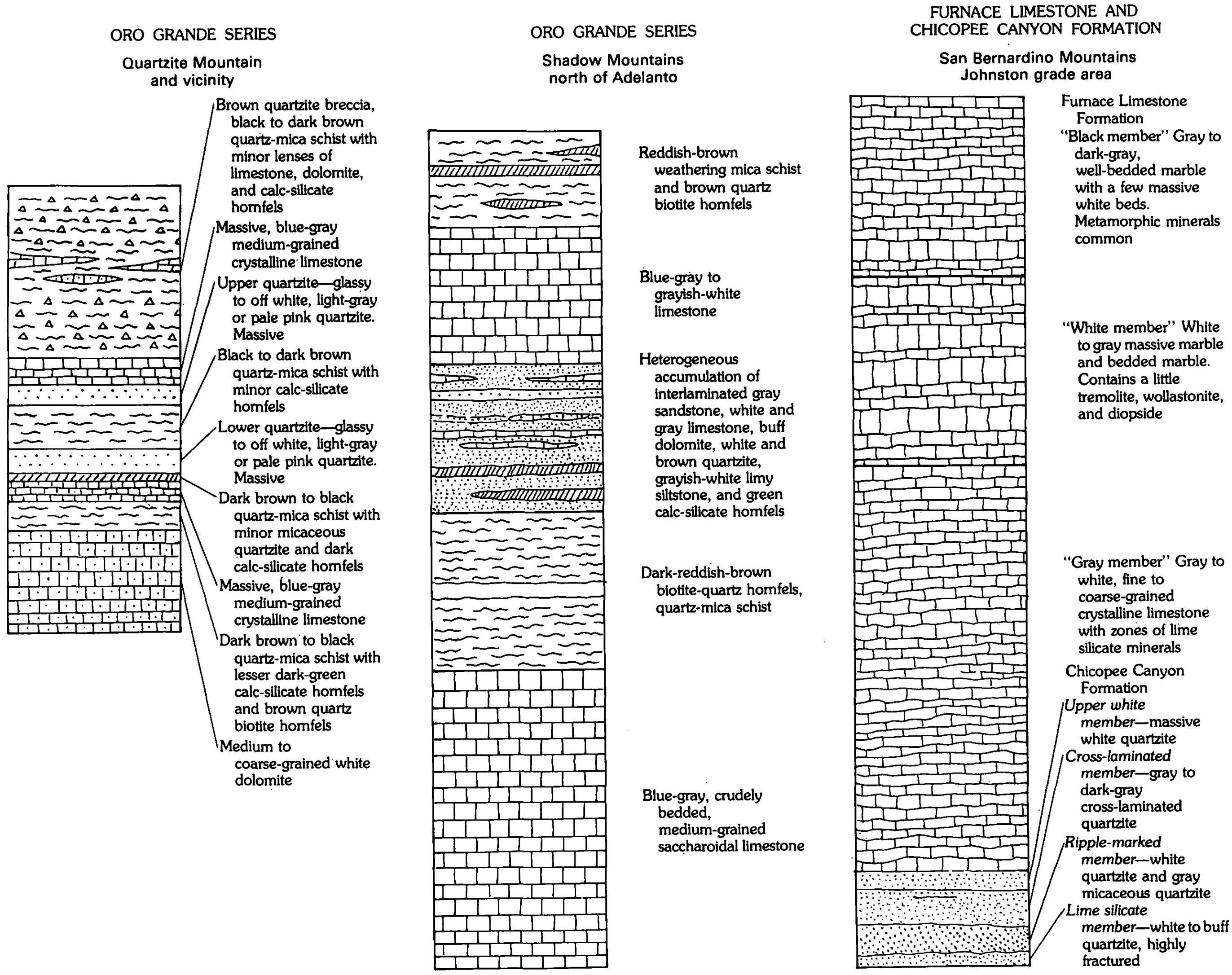

FIGURE 3.—Columnar sections of Carboniferous(?) rocks in the southwestern Mojave Desert.

tion on Quartzite Mountain contains two quartzite units 65 and 90 m thick, respectively.

Typical of the Shadow Mountains section of the Oro Grande series is a heterogeneous sandstone (unit 3) containing both quartz and calcite grains (intermingled and in discrete beds), white and gray crystalline limestone, buff dolomite, white and brown quartzite, grayish-white siltstone, and greenish calc-silicate hornfels. This sandstone is distinguishable, as it is interbedded with crystalline limestone and quartzite—both much more intensely metamorphosed.

The five-unit Shadow Mountains section of the Oro Grande series is, top to bottom, as follows:

	Meters
5. Reddish-brown-weathering mica-quartz schist and brown quartz-biotite hornfels	400
4. Blue-gray to grayish-white crystalline limestone	581
3. Light-colored mixed unit of sandstone, limestone, dolomite, quartzite, and schist	600
2. Reddish-brown to black mica schist and dark-brown biotite-quartz hornfels	672
1. Blue-gray, massive, medium-grained, crudely bedded crystalline limestone	1161
Total	3414

In and adjacent to the type section of the Oro Grande series on Quartzite Mountain in the Apple Valley 15-minute quadrangle, or in the Barstow

30-minute quadrangle, 9 units of the Oro Grande series are mappable at scales of 2,000 feet (610.5 m) per inch or larger. Some of these have subunits. From top to bottom, the sequence is as follows:

	Meters
9. Interbedded dark quartz-mica schist; coarse-grained, blue-gray crystalline limestone; quartzite breccia; and blue-gray and white crystalline limestone	400
8. Blue-gray, medium-grained crystalline limestone	138
7. Massive pinkish-white quartzite with rare traces of bedding	65
6. Black to dark-brown mica-quartz schist	82
5. Glassy to off-white or light-gray to pale-pink quartzite	90
4. Dark-brown to black quartz-mica schist with thin beds of quartzite	28
3. Blue-gray, massive, medium-grained crystalline limestone	90
2. Dark-brown to black quartz-mica schist with lesser dark-green calc-silicate hornfels	116
1. Coarse-grained, massive, white crystalline dolomite with thin lenses and patches of limestone	436
Total	1445

Numerous other pendants of the Oro Grande series rocks are scattered through the southwestern Mojave Desert, but the sections are less complete. The Oro Grande series correlates most closely with the Furnace Limestone and the Chicopee Formation of the San Bernardino Mountains. Its base is not exposed, but it is overlain unconformably by the Permian Fairview Valley Formation.

The complete section of the Oro Grande series undoubtedly is materially thicker than the 3,414-m Shadow Mountains section. McCulloh (oral commun., 1953) has described a nonfossiliferous sequence in the Lane Mountain area north of Barstow that is lithologically similar to the Oro Grande series but that contains more quartz-mica schist; it is 8,720 m thick. Vaughan (1922), Woodford and Harris (1928), Guillou (1953), and Richmond (1960) have estimated the Furnace Limestone to be 1,333 to 3,633 m thick at various places in the San Bernardino Mountains; all these sections are incomplete. If we take the quartzite of the Chicopee Formation and the Furnace Limestone to be more or less equivalent to the Oro Grande series, the total Carboniferous section in the Victorville-Cushenberry Canyon area may well be 4,100 m thick. McCulloh's section in the Lane Mountain area may indicate that the thickness of the Carboniferous strata in the southwestern Mojave Desert may be more than 8,700 m.

The Garlock series, an apparently continuous section of Paleozoic rocks, has been described by Hulin (1925) and Dibblee (1952) in the El Paso Mountains north of Garlock in the Randsburg and Saltdale 15-minute quadrangles. Although this sequence is probably isoclinally folded, it totals 12,716 m. Presence of one horizon of Permian fusulinids close to the apparent stratigraphic midpoint in the Garlock series suggests that it is at least half Permian or younger. Older rocks are apparently conformable with the Permian sequence. The Garlock assemblage is unlike the probable Carboniferous and Permian rocks south of the Garlock fault in the southwestern Mojave Desert, and it differs materially from the Permo-Carboniferous rocks to the north.

AGE OF THE ROCKS

The presence of two poorly preserved brachiopods (probably *Chonetes* sp., C. W. Merriam, oral commun., 1956) about 868 m above the lowermost exposed beds in the Oro Grande series section in the Shadow Mountains suggests a probable Pennsylvanian age for much of the sequence there. Presence of round crinoid columnals and debris in patches of obvious calcarenite at several localities in the Oro Grande series in the Quartzite Mountain area suggests a Paleozoic rather than a Mesozoic age. Strata from the type locality on Quartzite Mountain I consider to be approximately equivalent to the two lower members of the Shadow Mountains section. These are also equivalent to the Furnace Limestone and the Chicopee Formation (quartzite) in the San Bernardino Mountains. The lithology and sparse fossils indicate correlation among these formations.

The Oro Grande series is unconformably overlain by the Permian Fairview Valley Formation west of the Reserve quarry of Southwestern Portland Cement Company so that it is unlikely that any of the beds of the Oro Grande series in either the Quartzite Mountain or Shadow Mountains areas contain rocks of Permian age.

The fossil evidence thus far recovered from the Furnace Limestone of the San Bernardino Mountains indicates strata of both Mississippian and Pennsylvanian age. The Oro Grande series is also probably in part Pennsylvanian and in part Mississippian.

ROCKS AND MINERALS OF ECONOMIC IMPORTANCE

Limestone.—By far the most important industrial raw material taken from Carboniferous sequences in the southwestern Mojave Desert is limestone, and by far the largest consumer is the portland cement industry. The limestone deposits are medium to coarsely crystalline and are blue-gray to white. Purities as much as 98 percent $CaCO_3$ are common. Portland cement plants consume at least 4 million tons of limestone annually from probable Carboniferous deposits in the southwestern Mojave Desert. Amcord Inc., Riverside Division, with a plant at Oro Grande; the Kaiser Cement and Gypsum Corporation, with a plant at Cushenbury; and the Southwestern Portland Cement Company at Victorville are the principal consumers of limestone from the Oro Grande series and the Furnace Limestone.

Limestone from these rocks also plays an important role in the manufacture of beet sugar, glass, lime, and steel. It is used as white filler, roofing granules, road base, railroad ballast, and for aggregate in asphalt concrete and portland cement concrete. Pfizer, Inc., at Cushenbury, is the principal producer of these products, although several others are intermittently active in marketing competitive products.

Marble.—Marble dimension stone and ornamental stone have been quarried intermittently from the Oro Grande series and Furnace Limestone. The sawed polished dimension stone was used principally for floor and wall facings in large buildings. The old California State Mining Bureau headquarters at the Ferry Building, San Francisco, Calif., had a large slab of white, green, and black dolomite from the Three Color Marble quarry on the northeast side of Sidewinder Mountain. Building facings were also cut in the Gem quarry, about 1 mile north of Sidewinder Mountain and from the vicinity of Quartzite Mountain near Victorville. The ease of getting low-cost marble from Italy has largely eliminated quarrying of California's very beautiful marble.

Quartzite.—The very pure quartzite from the Oro Grande series, which has a silica content generally more than 98 percent has been used extensively for manufacture of silica refractories and portland cement. Because of its hardness and durability, this quartzite also has been used extensively for railroad ballast, road base, and aggregate for asphalt concrete and portland cement concrete. Amcord, Inc. and California Portland Cement Company have been the largest consumers.

Metallic minerals.—Gold, silver, copper, and tungsten have been produced in a small way from vein and replacement deposits in the Oro Grande series. The Carbonate gold mine near Oro Grande is noteworthy because of pockets of coarse gold found in the vicinity of the main shaft at a depth of 55.4 m. This mine was discovered about 1880, but most of the gold was produced between 1900 and 1942. Many other gold prospects are found in the Oro Grande series in the Victorville-Oro Grande and Cushenbury Canyon vicinities. All the gold deposits contain silver.

The only copper mine of much consequence has been the Amazon just east of Quartzite Mountain. Discovered about 1880 and first worked by Mormons, the early production was probably small. Between 1900 and 1942, a few hundreds of tons of chalcopyrite ore was shipped to smelters. The principal ore mineral was chalcopyrite, and it was commonly associated with magnetite. The wall rocks are limestone and quartzite cut by shear zones trending N. 10° W. to N. 10° E. Ore is found in the shear zones and in adjacent wall rocks as replacements.

Magnesite.—A small tonnage of high-purity magnesite was mined during 1940 on the Red Ball claims half a mile northeast of Sidewinder Mountain, for use in oxychloride cement. This cement was in demand for airfields in the Pacific theater during World War II. The magnesite occurs as hydrothermal veins, associated with black hornblende lamprophyre, cutting dolomite of the Oro Grande series. Most of the dikes and veins strike N. 35° E. and dip southeast at angles of 10° to 30°. The thickest veins are not much more than 1 m wide and can be as thin as about 25 cm.

CARBONIFEROUS STRATIGRAPHY OF PART OF EASTERN CALIFORNIA

By CALVIN H. STEVENS, GEORGE C. DUNNE, and RICHARD G. RANDALL

ABSTRACT

Carboniferous rocks in eastern California were deposited along the western margin of the North American craton. During Mississippian time, carbonate sediment was deposited on a westward-sloping shelf in the southeastern part of the area; mostly fine siliceous clastic sediment accumulated in a moderately deep water foreland basin to the northwest. During Pennsylvanian time, limestone was deposited through-

out the region; that to the southeast was deposited on a relatively shallow water, carbonate shelf, whereas that to the northwest was deposited as a turbidite sequence that accumulated in relatively deep water. Widespread unconformities apparently formed during the late Mississippian (late Meramecian), and during the late Middle Pennsylvanian (Desmoinesian). The unconformities and changes in lithologic and thickness patterns probably reflect tectonic events associated with the Antler orogenic belt, postulated to have extended north and perhaps west of the area of this study.

INTRODUCTION

This paper summarizes the Carboniferous stratigraphy in an area of about 20,000 km^2 in eastern California bounded by three major geologic features: the Garlock fault on the south, the Death Valley-Furnace Creek fault zone on the east, and the Sierra Nevada batholith on the west (fig. 4). South of the Garlock fault, along which as much as 60 km of left-lateral displacement may have taken place (Davis and Burchfiel, 1973), exposures of Carboniferous rocks are sparse and poorly dated. East of the Death Valley-Furnace Creek fault zone, Carboniferous rocks may have been displaced 100 km in a right-lateral sense (Poole and Sandberg, 1977). In the Sierra Nevada to the west, few rocks of Carboniferous age have been identified. Thus, the present study area is a natural geologic unit for analysis of the Carboniferous rocks exposed here.

We thank Forrest Poole for many helpful suggestions concerning both the study and the manuscript, and David Andersen and Rachel Gulliver who helped greatly in improving the manuscript.

GENERAL GEOLOGIC RELATIONSHIPS

Geomorphically, the area of study forms part of the westernmost Basin and Range province. Carboniferous rocks crop out widely in many of the ranges where exposures generally are excellent; intervening alluviated valleys, however, preclude long distance continuous tracing of units. Because numerous thrust faults of Mesozoic age are exposed or are inferred to underlie much of the study area (Stewart and others, 1966; Burchfiel and others, 1970; Johnson, 1971; Stevens and Olson, 1972), uncertainty exists as to the original locations of various terranes. Because the amount of slip on these faults is not known, no palinspastic restorations have been made in the compilation of our facies and isopachous maps; however, restorations are desirable, and they may become feasible in the future when the major structural features of the region are better understood.

Three localities of known or inferred Carboniferous rocks near the margins of the study area deserve mention; they are not otherwise included in the compilation because they cannot be correlated readily with other units in the area. These are: the Garlock Formation of Dibblee (1967) exposed just north of the Garlock fault in the El Paso Mountains; a slightly metamorphosed section in the Benton Range (fig. 4, loc. 26) near the northern boundary of the area; and metasedimentary rocks in the Mount Morrison pendant (loc. 25) in the eastern Sierra Nevada.

The Garlock Formation evidently includes rocks of several ages, some older and some younger than Carboniferous (F. G. Poole, oral commun., 1977). A flysch sequence there is considered by Poole (1974) to be Mississippian and possibly Pennsylvanian in age. No fossils have been obtained from the rocks in the Benton Range (fig. 4, loc. 26), but an argillite and carbonate sequence in the east-central part of the range appears very similar to the Rest Spring Shale and Keeler Canyon Formation in the Inyo Mountains to the south. The Mount Baldwin Marble in the Mount Morrison pendant in the Sierra Nevada (loc. 25) has yielded probable Early Pennsylvanian fossils (Rinehart and Ross, 1964), so it evidently is coeval with the lower part of the Keeler Canyon Formation. The underlying clastic unit may correlate with some of the siliceous clastic rocks in the Perdido Formation-Rest Spring Shale sequence in the Inyo Mountains.

PREVIOUS WORK

The first significant work on Carboniferous stratigraphy in eastern California was that of McAllister (1952), who named and described four Carboniferous units (Tin Mountain Limestone, Perdido Formation, Rest Spring Shale, and Tihvipah Limestone) exposed near Quartz Spring (loc. 20) in the northern Panamint Range. The stratigraphic scheme devised by him has been followed by most subsequent workers in the area, although the Mississippian nomenclature has since been slightly augmented (Hall and MacKevett, 1958); in addition, the Keeler Canyon Formation, named by Merriam and Hall (1957) for a largely Pennsylvanian limestone unit, evidently includes McAllister's Tihvipah Limestone —a name now largely abandoned—as well as younger limestone. Johnson (1957) proposed the name Anvil Spring Formation for a sequence of Carboniferous and Permian rocks in the southern Panamint Range (loc. 1), but inasmuch as that section is composed of several readily recognized, pre-

Area and number	Reference	Tin Mt. Limestone	Perdido Formation	Argus Limestone	Rest Spring Shale	Lower Keeler Canyon Fm.
Butte Valley (1)	(unpub. data)	?	present	235	0	present
Tucki Mountain (2)	Hunt and Mabey (1966)	300	145	75	230	760?
Tucki Mountain (3)	Randall (1975)	460	300	140+	225	650+
Bendire Canyon (4)	Moore (oral commun.)	?	?	?	?	120–210
Argus Range (5)	Holden (1976)	120	120	200	0	present
Argus Range (6)	Hall (1971)	115	115	200	0	30+
SE Darwin Hills (7)	Moffitt (oral commun.)	?	405+	95	13	0–10
SE Darwin Hills (8)	(unpub. data)	?	present	123	present	present
Panamint Butte (9)	Hall (1971)	135	120	200	0	445+
Santa Rosa Hills (10)	(unpub. data)	130	455	25	30	present
Santa Rosa Hills (11)	Hall and MacKevitt (1962)	135	(data not applicable)		0–15	<120
Marble Canyon (12)	Pelton (1966)	135	352?	113?	0	present
Goldbelt Springs (13)	Stadler (1968)	175	95	75	0	present
SE Inyo Mountains (14)	Elayer (1974)	105	120	0	215+	present
SE Inyo Mountains (15)	Merriam (1963)	25–105	0–25	0	>305	45–60
SE Inyo Mountains (16)	Stuart (1976)	25–105	15	0	present	60
SE Inyo Mountains (17)	Kelley (1973)	?	60	0	335	present
Ubehebe Mine (18)	McAllister (1956)	145	185	0	60–305	<45
Quartz Spring (19)	Langenheim and Tischler (1960)	110	210	0	present	present
Quartz Spring (20)	McAllister (1952)	145	185	0	>120	>60
Dry Mountain Quad. (21)	Burchfiel (1969)	0–130	185–215	0	230–275	present
Mazourka Canyon (22)	Stevens and Ridley (1974)	0	45–185	0	535	present
Mazourka Canyon (23)	Ross (1965)	0	90–185	0	<760	present
Tinemaha Reservoir (24)	(unpub. data)	?	present	0	present	595+
Mount Morrison (25)	Rinehart and Ross (1964)	?	?	?	present?	150?
Benton Range (26)	(unpub. data)	?	?	?	present?	present?
Talc City Hills (27)	Gulliver (1976)	present	present	present	present	present

Figure 4.—Location and number of measured sections and thickness of rock units in meters.

viously named Carboniferous and Permian units, this formational name herein is abandoned. The nomenclatural scheme used here is shown on a diagrammatic cross section (fig. 5).

Gross lateral changes in Mississippian rocks have been investigated by Pelton (1966) and Randall (1975), but detailed regional studies of these units have yet to be made. Pennsylvanian rocks are much more poorly known. Scattered Pennsylvanian sections have been studied, but this system has not been subjected to regional analyses. Therefore, many details of the stratigraphic relationships still are in doubt, and the nature of lithologic and thickness differences from place to place is poorly known.

CARBONIFEROUS PALEOTECTONIC SETTING

The paleotectonic setting of most of eastern California during the Carboniferous seems reasonably clear, although the significance of some sections is not well understood and not all the late Paleozoic tectonic belts recognized farther northeast in Nevada can be traced readily into California (fig. 6). Mississippian paleogeographic belts in northeastern Nevada were, from east to west, a carbonate shelf, a foreland basin, and the Antler orogenic highland (Poole, 1974; Rose, 1976; Poole and Sandberg, 1977). In eastern California, the carbonate shelf and foreland basin have been recognized. Part of the Antler orogenic highland has been identified tentatively in the Sierra Nevada (Speed and Kistler, 1977) northwest of the study area. Unlike Mississippian strata, the Pennsylvanian section in eastern California differs considerably from that in northeastern Nevada. A carbonate shelf, however, is recognized on the east in both areas.

STRATIGRAPHIC SETTING

Carboniferous rocks are underlain throughout most of the region by Devonian rocks assigned to the Lost Burro Formation. The contact seems conformable but abrupt, and was interpreted by Langenheim and Tischler (1960) and Poole (1974) to represent an interruption in sedimentation. The only known exception to this stratigraphic interpretation is in the Independence quadrangle in the extreme western Inyo Mountains (loc. 23). There, the Lost

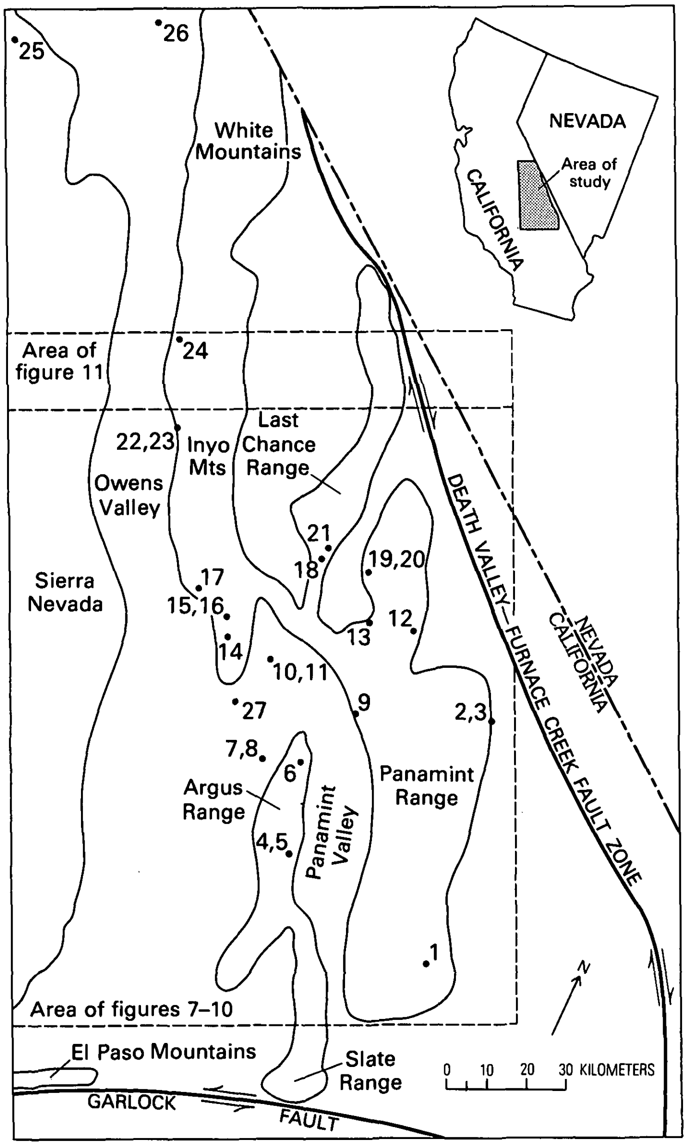

FIGURE 4.—Continued.

Burro Formation is not present, and the position of the Devonian-Mississippian contact still has not been established. The systemic boundary may be at the base of the Perdido Formation, as restricted by Stevens and Ridley (1974), or it may be at the base of the Squares Tunnel beds of Stevens and Ridley. The contact below the Squares Tunnel beds clearly is disconformable; the contact at the base of the Perdido Formation also may be disconformable.

Carboniferous rocks of the region apparently are overlain by rocks of Permian age everywhere except in part of the southern Inyo Mountains (loc. 15), where locally the Pennsylvanian part of the Keeler Canyon Formation is overlain unconformably by Triassic marine strata. The boundary between Permian and Carboniferous rocks is generally within the Keeler Canyon Formation, and in most areas it is not marked by a lithologic change.

As far as is known, no major tectonic event affected this area during the Carboniferous, but minor events are indicated by changing lithofacies and isopachous map patterns.

GENERAL STRATIGRAPHIC RELATIONSHIPS

Within the area of study, the Carboniferous section can be divided as shown in figure 5. Two widespread disconformities, one formed during the Late Mississippian (late Meramecian) and the other formed during the Middle Pennsylvanian (Desmoinesian), are recognized. The older disconformity generally is marked by a sharp lithologic change from limestone, locally dated as middle Meramecian, to fine siliceous clastic rocks locally considered late Chesterian in age (Poole and Sandberg, 1977). Poole and Sandberg suggested that a middle Meramecian uplift affected much of western United States.

The second widespread disconformity, separating lower Middle from Upper Pennsylvanian rocks, is recognized over much of the area. Most lower Middle Pennsylvanian (Derryan) rocks are fine grained and characterized by the presence of round chert nodules commonly the size of golfballs. In the central and western parts of the area, the Upper Pennsylvanian sequence above the unconformity is represented by interbedded fine- and coarse-grained carbonate units, which represent a turbidite sequence. Beds immediately above the unconformity apparently range in age from Missourian in the southern Inyo Mountains (loc. 15) to Wolfcampian (Early Permian) in the Santa Rosa Hills (loc. 10). The break between the two Pennsylvanian sequences often appears abrupt.

CARBONIFEROUS UNITS

TIN MOUNTAIN LIMESTONE

Wherever the Tin Mountain Limestone has been recognized, it overlies the Devonian Lost Burro Formation with apparent conformity, although Poole (1974) and Poole and Sandberg (1977) considered the contact disconformable. At the type area near Quartz Spring (loc. 20), the contact is marked by a change from very light-gray quartzite of the Quartz Spring Sandstone Member of the Lost Burro Formation to the clayey lower Tin Mountain Limestone (Langenheim and Tischler, 1960). There, the lower part of the Tin Mountain Limestone con-

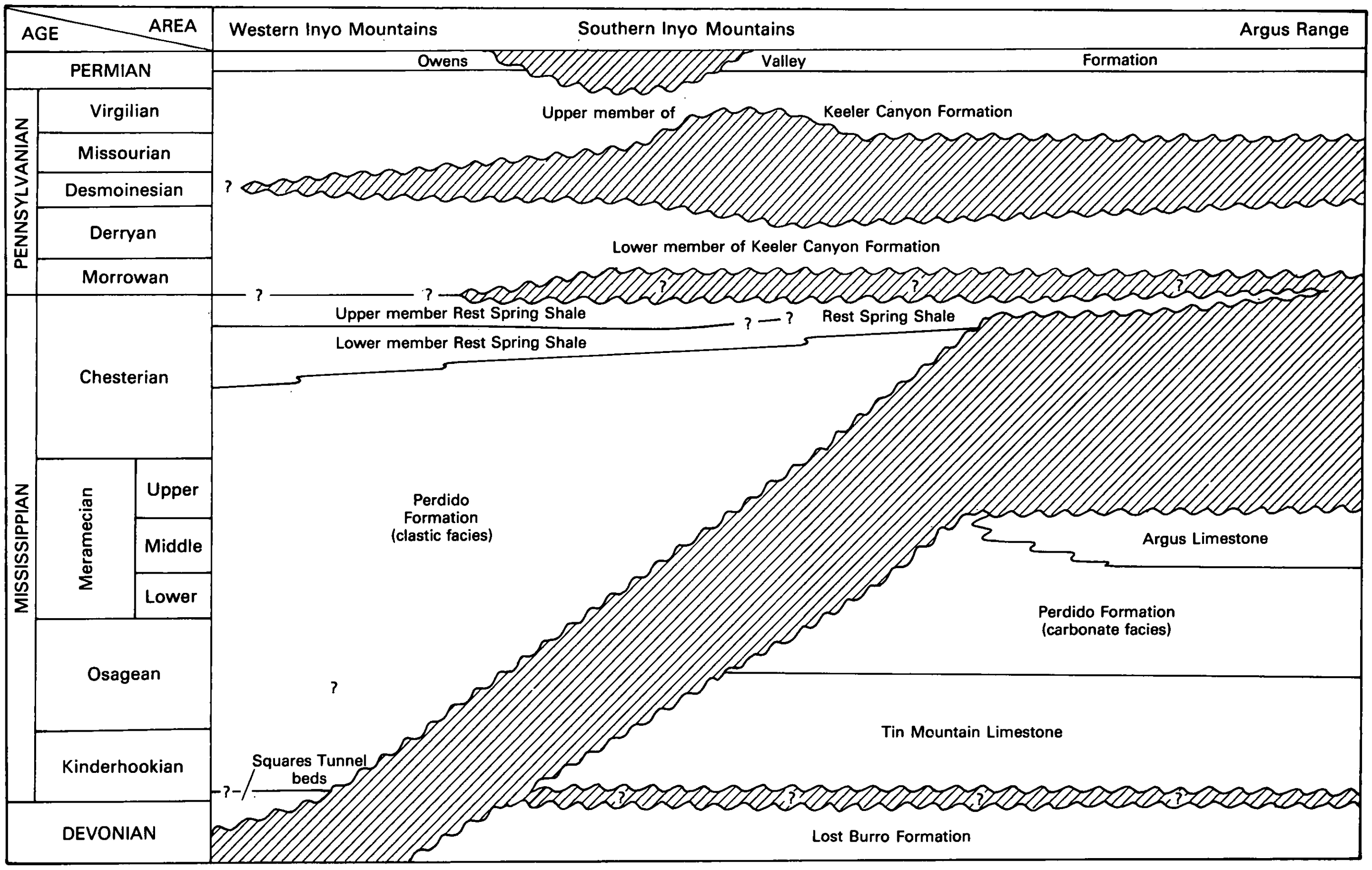

FIGURE 5.—Diagrammatic cross section of eastern California showing stratigraphic nomenclature and correlations.

sists of dark-gray limestone beds 5 to 15 cm thick separated by much thinner beds of light-bluish-gray to pale-red shale (McAllister, 1952). The upper part consists of medium-gray limestone in beds a few centimeters to 50 cm thick. Some beds are composed largely of crinoid stems. The persistence of dark limestone, pale-red shale partings, very dark-gray chert only in nodules, and an abundance of *Syringopora* is considered diagnostic of the Tin Mountain Limestone (McAllister, 1952).

Throughout most of the region considered, the Tin Mountain Limestone is 105 to 145 m thick (fig. 7). Thinning, however, takes place northwestward in both the Inyo and Last Chance Ranges. This thinning has been attributed to post-Tin Mountain and pre-Perdido erosion because locally uppermost Tin Mountain Limestone beds are missing (Burchfiel, 1969). At Tucki Mountain (loc. 3), where all Carboniferous units apparently are anomalous with regard to thickness and/or lithology, the Tin Mountain Limestone is much thicker than elsewhere (300 m estimated by Hunt and Mabey, 1966; 460 m computed by Randall, 1975). Stevens and others (1974) have suggested that the rocks composing Tucki Mountain and adjacent parts of the southern Panamint Range have been displaced several tens of kilometers, but resolution of this question requires further work.

The Tin Mountain Limestone is the most fossiliferous Carboniferous formation in the region. Langenheim and Tischler (1960) listed 18 species of brachiopods, 13 species of corals, 6 species of gastropods, and smaller numbers of several other groups of fossils. In the Panamint Butte quadrangle (locs. 6, 9), Hall (1971) listed 18 species of brachiopods, 15 species of gastropods, 14 species of corals, and smaller numbers of 5 other groups. These diverse fossils demonstrate that the Tin Mountain Limestone is Early Mississippian in age (Langenheim and Tischler, 1960; Hall, 1971).

Limited evidence suggests that the Tin Mountain Limestone was deposited in fairly shallow water. Cerioid corals, which are especially abundant on the east side of the central Argus Range (loc. 5),

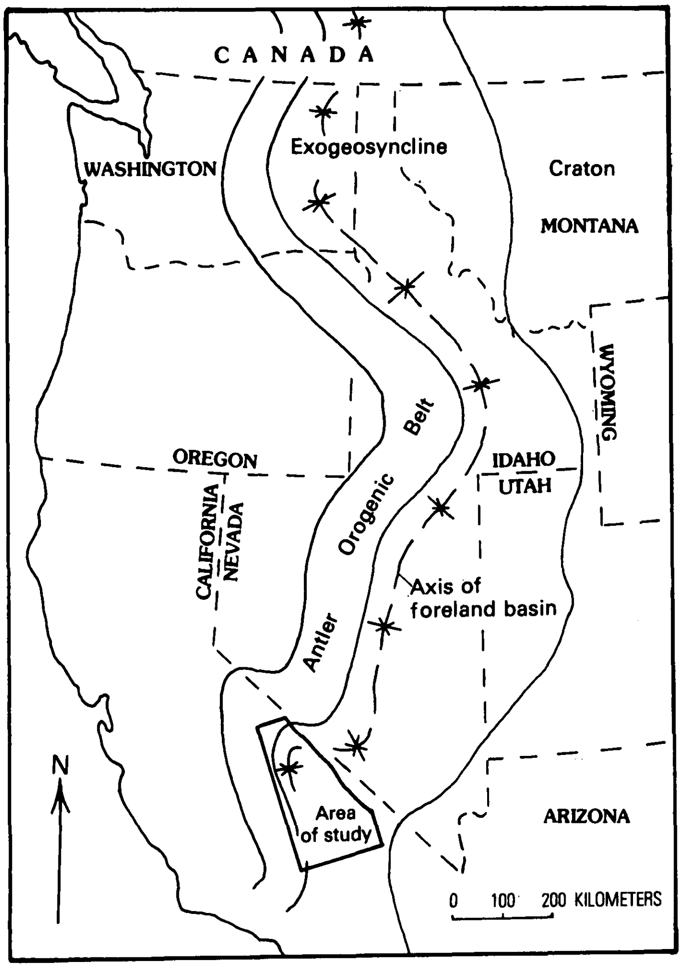

FIGURE 6.—Paleotectonic map of eastern California during the Carboniferous Period.

may be the best indicators of environment. In Upper Mississippian rocks in the Santa Rosa Hills and Lower Permian rocks in northeastern Nevada, corals with this morphology evidently lived in relatively shallow water near the shelf margin (Stevens, 1977). Evidence for strong currents or extremely shallow water is meager except in the Quartz Spring area, where Tischler and Langenheim (1960) noted bioclastic limestone beds more than 1 m thick that have well-developed crossbedding and lenses of coarse fragmental rock. Deposition of the Tin Mountain Limestone is interpreted here to have been on a broad shelf that had good circulation at water depths perhaps less than 20 to 30 m.

PERDIDO FORMATION

The Perdido Formation overlies the Tin Mountain Limestone with apparent conformity except in the south-central to western Inyo Mountains (locs. 15, 22) and in the Last Chance Range (loc. 21), where the Tin Mountain Limestone is thin or absent.

Two distinct facies of the Perdido Formation are recognized—a southeastern carbonate and a northwestern clastic facies (fig. 8). At the type locality in the Quartz Spring area, the clastic facies overlies the carbonate facies. Elsewhere, apparently only one facies is present in any given area, except locally, where the two facies may have been juxtaposed by thrust faulting, as in the Talc City Hills (Gulliver, 1976).

Carbonate facies.—At the type locality (loc. 20), the lower cherty limestone facies of the Perdido Formation rests on the Tin Mountain Limestone. The contact is a well-defined bedding plane which separates thick-bedded, coarse-grained pelmatozoan limestone below from relatively thin-bedded, fine, cherty, argillaceous limestone above (Langenheim and Tischler, 1960). The lower part of the Perdido Formation consists primarily of nodular layers of limestone 10 to 15 cm thick, which are mostly fine grained and black. Minor coarse-grained and pelmatozoan limestone is present. Interbedded black chert weathers reddish orange or very dark gray and is found in almost all limestone units. Most chert beds are less than 15 cm thick, and many are nodular and discontinuous layers. Bedding-plane partings of yellowish-gray to red shale are widespread and make up slightly more than 25 percent of the basal cherty sequence (Langenheim and Tischler, 1960). A middle, chert-free unit in the lower Perdido Formation is characterized by beds 5 to 60 cm thick and by the presence of more shale beds than below. Limestone beds in the upper unit of the lower Perdido Formation are thicker than those at the base, ranging from 15 cm to more than 1 m in thickness, but the black limestone and cherty shale is much like that of the lower unit.

Clastic Facies.—The upper part of the Perdido Formation at the type locality (loc. 20) is mainly red weathered and calcareous siltstone (Langenheim and Tischler, 1960). A few layers of cherty and shaly limestone as much as 50 cm thick are present, and the uppermost 12 m consists of limestone conglomerate, limestone, and calcareous siltstone or shale. A limestone conglomerate at the base of the 12-m-thick upper sequence contains boulders of Perdido limestone, chert, and siltstone as much as 10 cm in diameter in a richly fossiliferous limestone matrix (Langenheim and Tischler, 1960). A somewhat similar limestone conglomerate, interpreted by us as a submarine debris flow, is found in the southeasternmost Inyo Mountains.

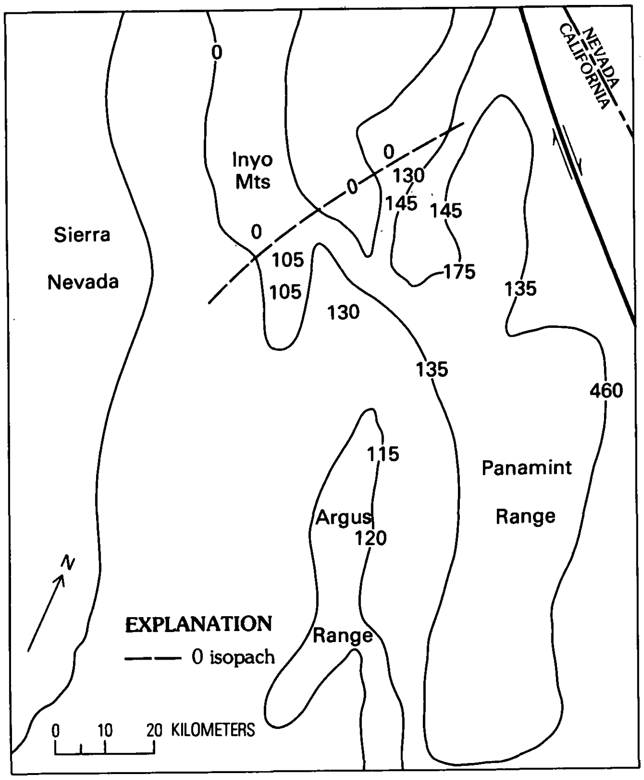

FIGURE 7.—Thickness, in meters, of the Tin Mountain Limestone.

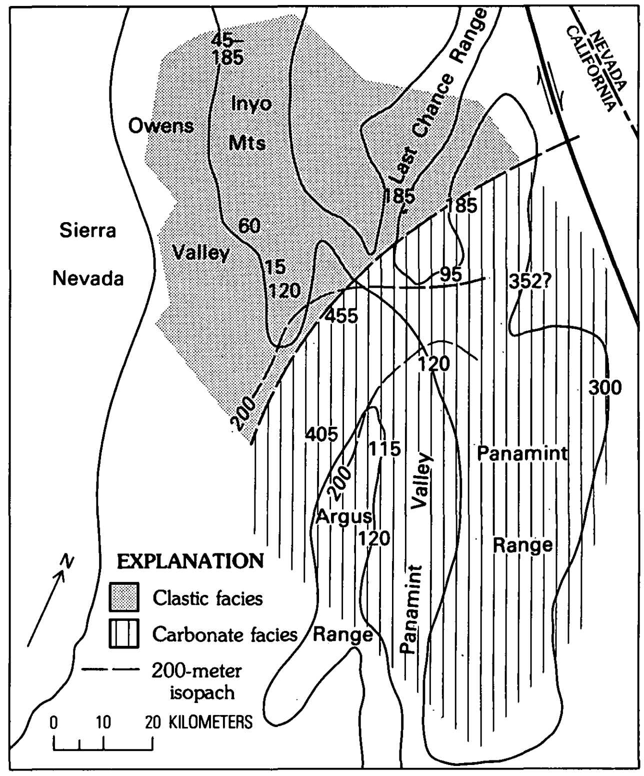

FIGURE 8.—Thickness, in meters, and distribution of facies of the Perdido Formation.

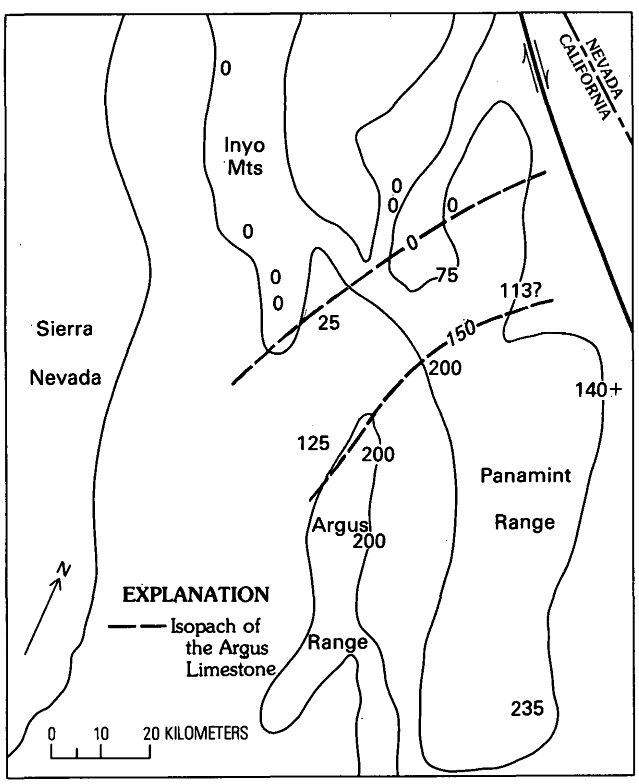

FIGURE 9.—Thickness, in meters, of the Argus Limestone.

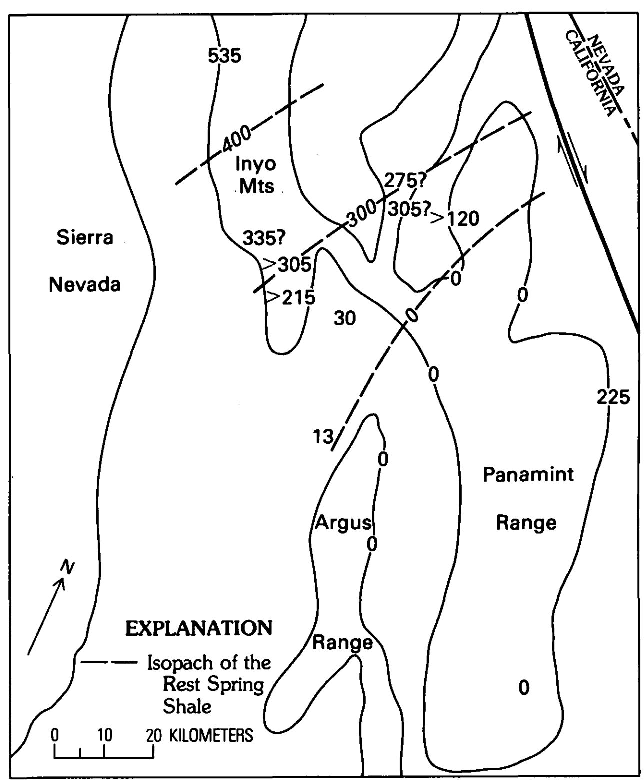

FIGURE 10.—Thickness, in meters, of the Rest Spring Shale.

Perdido Formation facies interpretation.—Thickness of the Perdido Formation is quite variable (fig. 8). The maximum thickness of 455 m is near the transition between the southeastern carbonate and northwestern clastic facies. Northwestward, into the clastic facies, thicknesses decrease and then generally increase. Southeastward into the carbonate facies, thicknesses decrease abruptly except for the anomalously thick section at Tucki Mountain (loc. 2).

Stratigraphic relations between the carbonate and clastic facies of the Perdido Formation are perplexing. As yet, we have no faunal evidence that the carbonate rocks are replaced laterally (to the northwest) by clastic rocks of the same age, although this seems to be the most likely explanation. In both the Quartz Spring area (loc. 20) and southeasternmost Inyo Mountains (east of loc. 14), carbonate debris flows derived from a carbonate shelf are found in the clastic facies, indicating that during deposition of the clastic facies a carbonate shelf existed within or adjacent to the study area.

Except for locally abundant pelmatozoan columnals, fossils generally are uncommon in the Perdido Formation. Langenheim and Tischler (1960) reported a plant, corals, brachiopods, pelecypods, gastropods, and cephalopods, which they interpreted as Late Mississippian in age. Conodonts have been recovered from several parts of the carbonate facies of the Perdido Formation (strata formerly mapped as Lee Flat Limestone) in the Santa Rosa and Darwin Hills (G. C. Dunne, California State University and R. H. Miller, San Diego State University, unpub. data, 1977), which indicate that it is Meramecian in age. Ammonoids from the clastic facies of the Perdido Formation in the Quartz Spring area (loc. 20) have been dated as Chesterian. This unit was placed with the Rest Spring Shale by Poole and Sandberg (1977), who interpreted this entire clastic sequence to be considerably younger than the Perdido carbonate facies.

The environment of deposition of the clastic facies of the Perdido Formation is perhaps better understood than that of the carbonate facies. In the west-central Inyo Mountains (loc. 22), the base of the Perdido Formation is marked by a debris flow containing slide blocks as much as 30 cm in diameter. These deposits are concentrated in submarine channels that previously had been partially filled with bedded radiolarian chert (Stevens and Ridley, 1974). These rocks are overlain by fine-grained clastic rocks and carbonate grain flows and turbidites. To the southeast, the clastic facies of the Perdido Formation is represented by a thin sequence of fine-grained, light-gray quartzite and dark siltstone. In the Inyo Mountains, shale and siltstone beds contain numerous trace fossils similar to those of the *Nereites* community illustrated by Poole (1974), which suggest deep-water deposition. Carbonate debris flows, interbedded with fine-grained clastic rocks, also suggest that the clastic facies was deposited in water considerably deeper than that inferred for the carbonate shelf.

The depositional environment of the carbonate facies is more difficult to interpret. The very fine grained limestone contains few shallow-water indicators except in the Quartz Spring area, where crossbedded, coarse-grained pelmatozoan limestone was reported by Langenheim and Tischler (1960). We believe that rocks of this facies were deposited on a moderately deep water shelf. As this interpretation is based partly on the relationship of the carbonate facies with the overlying Argus limestone, a more complete discussion is deferred.

ARGUS LIMESTONE

Argus limestone is an informal formational name used here for strata in the Argus and Panamint Ranges previously referred to as the Lee Flat Limestone. This unit rests conformably on, and is approximately coextensive with, the carbonate facies of the Perdido Formation. The contact between the two units is placed at a point where upward in the section, the limestone becomes distinctly coarser grained, more pelmatozoan-rich, lighter in color and more massive, and where chert becomes much less abundant. In contrast to the carbonate facies of the Perdido Formation, which thins southeastward, the Argus limestone (fig. 9) thickens southeastward, except at Tucki Mountain (loc. 3).

In most exposures, the Argus limestone contains no fossils other than pelmatozoan columnals, which commonly are so abundant that they form a major part of the formation. In the Santa Rosa Hills (loc. 10), however, a variety of fossils has been recovered, including 3 species of colonial rugose corals, 4 species of conodonts, and 12 species of Foraminifera. Several species of conodonts have also been recovered from the Argus limestone in the southeast Darwin Hills (loc. 8). These fossils indicate that the unit in these areas is mostly, if not entirely, middle Meramecian in age (Dunne and others, 1977). Corals from the Argus limestone in the Panamint Range (Hunt and Mabey, 1966) suggest a similar age for the unit in that region also.

The Argus limestone is believed to have been deposited in shallow water. The unit is characterized by abundant pelmatozoans in well-sorted and well-washed limestone that suggest reworking by moderately strong marine currents or waves. In addition, the abundance of cerioid corals at several horizons in the Santa Rosa Hills (loc. 10) suggests relatively shallow water, perhaps a shelf edge above deeper water to the west, as has been proposed for similar occurrences of Permian cerioid corals (Stevens, 1977).

Fossils show that the Argus limestone and the carbonate facies of the Perdido Formation are partly coeval; lithologic characteristics and geographic position suggest that the calcareous facies of the Perdido Formation is an offshore, deeper water facies of the Argus limestone. Farther northwest, the clastic facies of the Perdido Formation may have been accumulating simultaneously in deeper water. We speculate that the shallow-water Argus limestone prograded westward across the calcareous facies of the Perdido Formation, ultimately reaching the Santa Rosa Hills (loc. 10), but not the Quartz Spring area (loc. 20) or southern Inyo Mountains (loc. 14).

REST SPRING SHALE

According to our studies, the Rest Spring Shale is conformable on the clastic facies of the Perdido Formation in the northwestern part of the area. Thin clastic beds correlated with the Rest Spring Shale rest, perhaps disconformably, on the Argus limestone in the Darwin quadrangle (locs. 7, 10). At the type locality near Quartz Spring (loc. 20), the Rest Spring Shale consists of a lower argillaceous shale, which is olive gray to dark gray when fresh (McAllister, 1952); it grades upward into an upper unit of siltstone to fine-grained sandstone. Possible intraformational conglomerate containing sandstone clasts as much as 5 cm across was reported by McAllister (1952); he also described some beds showing channeling, crossbedding, and ripple marks.

Accurate thicknesses of the incompetent Rest Spring Shale are difficult to measure because this formation commonly is the locus of folding and faulting. At the mouth of Mazourka Canyon (loc. 23), however, a complete section is 535 m thick (Stevens and Ridley, 1974). The formation thins southeastward (fig. 10), while showing an increase in quartz-sand content. In the Argus Range near the apparent limit of deposition, a clastic unit only about 2 m thick may represent the Rest Spring Shale.

Fossils are uncommon in the Rest Spring Shale. Ammonoids occur at several localities, and there are a few occurrences of brachiopods, plants, bryozoans, and microfossils (McAllister, 1952). Merriam (1963) reported similar fossils in the southern Inyo Mountains. Ammonoids in the Inyo Mountains (Merriam, 1963; Ross, 1965) and the Quartz Spring area (McAllister, 1952) are considered Late Mississippian (Chesterian) in age. Chesterian fossils also occur in the underlying clastic facies of the Perdido Formation in the Quartz Spring area (McAllister, 1952), and Derryan (Middle Pennsylvanian) fossils are found near the base of the overlying Keeler Canyon Formation (Meriam, 1963); hence, the formation is Chesterian to perhaps as young as Derryan in age.

Environmental indicators in the Rest Spring Shale are sparse. Trace fossils in many of the western outcrops are similar to those reported by Poole (1974) as representing the *Nereites* community and interpreted to represent relatively deep water. The uniformly fine-grained nature, dark color, and paucity of fossils in most of the area also suggest a basinal environment far from shore. Channeling and crossbedding in the upper part of the sequence in the Quartz Spring area could be features associated with a submarine channel system in deep water, although alternatively they may reflect the onset of shallow-water conditions.

We propose that the basin in which the Rest Spring Shale accumulated was northwest of the edge of the carbonate shelf in the Santa Rosa Hills (loc. 10) and that it deepened northwestward beyond the Inyo Mountains. Thus, this area is inferred to be on the southeastern side of the foreland basin (fig. 6) that lay east of the Antler orogenic highlands (Poole, 1974; Poole and Sandberg, 1977; Speed and Kistler, 1977).

In much of the region, the Rest Spring Shale is divisible into two units here called the lower and upper members of the Rest Spring Shale. The two members generally are found together, and the overall distribution suggests that they are essentially coextensive.

Lower member.—The term Chainman Shale was used in the Inyo Mountains by several workers (Langenheim, 1963; Merriam, 1963; Stevens and Ridley, 1974) for this relatively homogeneous, very dark-gray shale unit. Trace fossils locally are abundant on bedding planes, but other fossils are rare. The section at the mouth of Mazourka Canyon (loc. 22) is the only locality known to us where this unit apparently is structurally undisturbed; here, this member is about 100 m thick (Stevens and Ridley, 1974). This unit, which may constitute the entire Rest Spring Shale at the type section near Quartz Spring, crops out widely in the Inyo and Last Chance Ranges.

Upper Member.—The upper member of the Rest Spring Shale, referred to as the Hamilton Canyon Formation by Langenheim (1962) and Stevens and Ridley (1974), overlies the lower member with apparent conformity. It is composed of dark-brown, blocky-weathering andalusite hornfels, generally containing obscure bedding. The outcrop character, bedding, composition, and color are different from that of the lower member, so the two units normally can be distinguished easily. This unit has been recognized from Tinemaha Reservoir (loc. 24) on the northwest to the Darwin Hills (northwest of loc. 8) on the southeast. It is present throughout the Inyo Mountains and probably is represented in the Last Chance Range in an allochthonous block described by Burchfiel (1969) as andalusite hornfels. Near the mouth of Mazourka Canyon (loc. 22), in an apparently undisturbed section, this unit is about 430 m thick (Stevens and Ridley, 1974).

KEELER CANYON FORMATION

The Keeler Canyon Formation overlies the Argus limestone or Rest Spring Shale everywhere in the area of study. Rich (1977) indicated an absence of earliest Pennsylvanian rocks in this area and Gordon and Poole (1968) reported their absence to the east. If an unconformity is present at the base of the Keeler Canyon Formation, however, it is not conspicuous. Two distinct, predominantly limestone units that differ in lithology and age were combined by Merriam and Hall (1957) to form the Keeler Canyon Formation. Here, the two units are referred to as the lower and upper members of the Keeler Canyon Formation and are described separately.

Lower member.—The lower member of the Keeler Canyon Formation at most locations consists of fine-grained, silty, dark-gray limestone with rounded, very dark-gray chert nodules and some calcareous shale. In the Santa Rosa Hills, Hall and MacKevett (1962) also reported thin limestone conglomerate beds. Fossils are not diverse, but fusulinids and pelmatozoan stems often are abundant, and horn corals and brachiopods occur sporadically. *Fusulinella,* which indicate a Derryan (early Middle Pennsylvanian) age, are reported from the southern Inyo Mountains and Santa Rosa Hills (Merriam, 1963), the Argus Range (Hall, 1971), Tucki Mountain (R. G. Randall, San Jose State University, unpub. data, 1975), and Tinemaha Reservoir (C. H. Stevens, San Jose State University, unpub. data, 1976).

The lower member of the Keeler Canyon Formation crops out throughout the region. Where the Keeler Canyon Formation overlies the Rest Spring Shale, the base is placed at the bottom of the lowest limestone in the predominantly limestone sequence, even though some shale similar to the Rest Spring is found higher in the section along Owens Valley. Where the Rest Spring Shale has not been mapped separately, the dark-gray silty limestone of the Keeler Canyon Formation is easily distinguished from the light-gray Argus limestone below.

The lower member ranges in thickness from a few meters, in parts of the Darwin Hills (loc. 8) and Santa Rosa Hills (loc. 10), to 100 to 120 m in most other areas (fig. 11). Much greater thicknesses are present at Tucki Mountain (loc. 3), where Derryan fusulinids have been collected near the top of a 650-m-thick section, and near Tinemaha Reservoir (loc. 24), where similar fusulinids are found near the top of a 600-m-thick section. In the Santa Rosa Hills, this member is overlain by conglomerate of probable earliest Permian age. Elsewhere, an unconformity above the lower member of the Keeler Canyon Formation apparently separates lower Middle from Upper Pennsylvanian strata; Desmoinesian strata may be missing in most of the area. The overall distribution of thicknesses of the lower Keeler Canyon Formation suggests that uplift and erosion, especially in the Santa Rosa and Darwin Hills area (locs. 10, 7) gave rise to some of these differences. The significance of the very thick sequences on the east at Tucki Mountain (loc. 3) and on the northwest at Tinemaha Reservoir (loc. 24), however, is not understood.

The environment of deposition also is difficult to evaluate. The fine-grained nature of the sedimentary rock, the paucity of fossils, and the stratigraphic position above the Rest Spring Shale or its thin equivalent, however, suggest a relatively deep-water shelf or basinal environment.

Upper member.—The upper member generally is represented in the northwestern part of the area by a calcareous turbidite sequence that presumably was deposited in deep water. Measurements of many flute casts and other directional features in the southern Inyo Mountains (loc. 15) show transport in a S. 55° W. direction, whereas approximately equivalent strata in the Ubehebe mine area (loc. 18) indicate transport southward (Tim Parker, oral

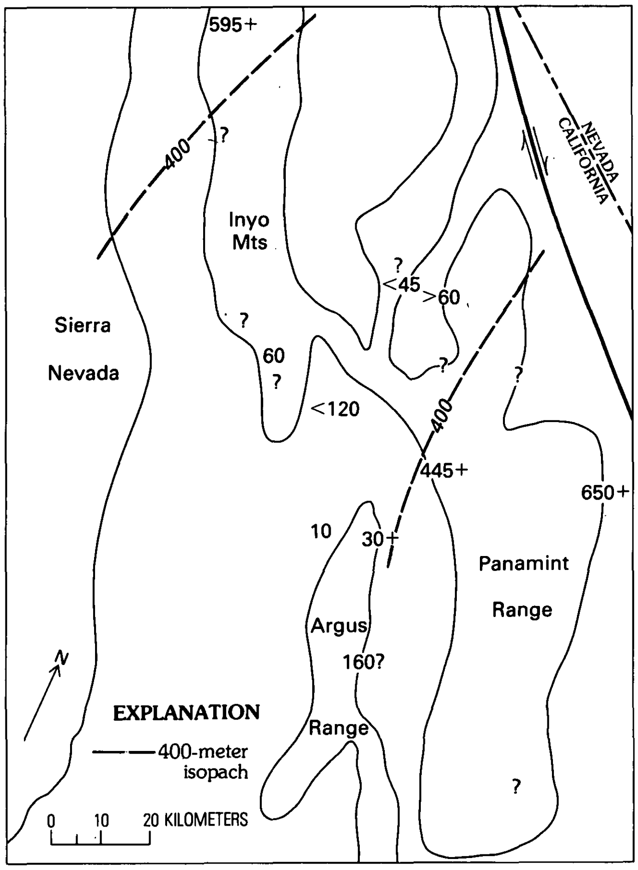

FIGURE 11.—Thickness, in meters, of the lower member of the Keeler Canyon Formation.

commun., 1976). Farther southeast in the Santa Rosa and Darwin Hills (locs. 10, 7), equivalent strata are very thin and conglomeratic. These areas may have furnished much of the debris for the turbidite sequence farther northwest. Still farther southeast, in the southern Panamint Range (loc. 1), coeval rocks are fine grained and contain brachiopods and bryozoans. These beds are interpreted as representing shallow-shelf deposition.

Thickness of the upper member ranges from a few meters to perhaps 1,000 m. Fusulinids are very abundant in the region between Panamint Valley and central Inyo Mountains (loc. 17), and these fossils show that the upper member of the Keeler Canyon Formation is Late Pennsylvanian and earliest Permian in age.

Generally, the upper Keeler Canyon Formation rests unconformably on the lower member, and in one locality in the southeastern Darwin Hills (loc. 8), it rests directly on the Rest Spring Shale.

INTERPRETATIONS AND SUMMARY

Carboniferous rocks in the region between the Sierra Nevada batholith and the Death Valley-Furnace Creek fault zone were deposited west of the western margin of the North American craton (Poole, 1974; Poole and Sandberg, 1977; Rich, 1977). All Carboniferous units of the southeastern part of the area of study were deposited in shallow-marine water, whereas all units in the northwestern part may have been deposited in deep water. In the intervening area (between the south-central Inyo Mountains and the Santa Rosa Hills), depositional environments may have fluctuated from relatively shallow to deep water as influenced by varying combinations of transgression, regression, deformation, and shelf progradation.

Carboniferous deposition began with formation of the dominantly shallow-water Tin Mountain Limestone on a shelf that extended as far northwest as the south-central Inyo Mountains. Tin Mountain deposition may have been terminated by rapid subsidence of the shelf because of eastward growth of the Antler foreland basin. Concurrently, submarine erosion of the Tin Mountain Limestone, localized near the original depositional margin of this unit in the southern Inyo Mountains and Last Chance Range, may have taken place at a relatively deep, hingeline position.

Deposition of the carbonate facies of the Perdido Formation began in moderately deep water on a depressed shelf in the southeastern part of the area; the clastic facies accumulated in deeper water to the west. The carbonate facies of the Perdido Formation may have accumulated relatively slowly, but the rate of deposition was greater than that of subsidence, so that later the shallow-water clastic Argus limestone began to prograde westward across the Perdido carbonate shelf. Progradation of the Argus limestone, however, was terminated before these clastic limestones reached the edge of the shelf.

To the northwest, over all of the area where the clastic facies of the Perdido Formation had been deposited previously, fine siliceous sediment of the Rest Spring Shale accumulated in relatively deep water. Subsidence of the shelf resulted in deposition of a thinner and possibly younger wedge of Rest Spring Shale on the margin of the carbonate shelf in the Santa Rosa Hills and Darwin Hills, a pattern similar to the eastward-spreading flysch sediment reported elsewhere along the eastern margin of the Antler flysch trough (Poole, 1974; Poole and Sandberg, 1977).

Events causing an increase in carbonate production and a decrease in the availability of fine, siliceous, clastic sediment during Early Pennsylvanian time resulted in the change in deposition to limestone in the lower Keeler Canyon Formation in a rather deep-water, carbonate-shelf or basinal environment. During Desmoinesian (Middle Pennsylvanian) time, nondeposition and/or erosion was widespread, but in the Late Pennsylvanian, deposition began again. In the eastern Darwin Hills and Santa Rosa Hills, little, if any, sediment accumulated initially, but to the west, thick sequences of limestone turbidites formed. Deposition in this turbidite basin continued into the Permian without a break. To the southeast, in the southern Panamint Range, deposition of relatively shallow-water shelf carbonates continued.

CARBONIFEROUS ROCKS OF THE EASTERN SIERRA NEVADA

By Ronald W. Kistler and Warren J. Nokleberg

ABSTRACT

Metamorphosed Carboniferous sedimentary rocks are exposed discontinuously in roof pendants in the eastern Sierra Nevada for about 110 km in a northwest-trending belt that averages about 3 km in width. The original Carboniferous strata were predominantly siliceous and calcareous mudstone and minor limestone, quartz-sandstone, and chert. Identifiable fossils are rare in these rocks, and age assignments in most roof pendants are made on the basis of lithologic correlations with the nearest fossiliferous rocks. The Carboniferous rocks are strongly deformed, and N. 20°–30° W. -trending folds are the dominant structures in all pendants along the belt. The Carboniferous rocks are west of and apparently unconformably overlay adjacent metamorphosed sedimentary rocks of Ordovician and Silurian(?) age. Farther west, the Carboniferous rocks are unconformably overlain by metamorphosed Mesozoic volcanic flows and sedimentary rocks. The regional strike of Carboniferous rocks in the eastern Sierra Nevada transects, at a large angle, regional trends of similar lithologies of the same age that crop out to the east. The transection of these regional trends is probably due either to severe, large-scale tectonic dislocations, or a regional angular unconformity.

INTRODUCTION

Metamorphosed Carboniferous sedimentary rocks are exposed discontinuously for about 110 km in a northwest-trending belt of roof pendants in the east-central part of the Sierra Nevada batholith (fig. 12). This paper summarizes lithologies, fossils, and structures in this belt, and discusses relations of Carboniferous rocks in this and adjacent belts.

PINE CREEK PENDANT

The Pine Creek pendant (fig. 12, loc. 1), mapped and described by Bateman (1965), contains three seemingly conformable units that were correlated with fossiliferous Pennsylvanian and Permian(?) strata in the Mount Morrison pendant exposed about 15 km to the northwest. The basal unit, which consists of about 350 m of pelitic hornfels, micaceous quartzite, and vitreous quartzite, is overlain by about 250 m of marble. The marble unit is overlain by an upper unit consisting of about 1,000 m of micaceous quartzite. The basal unit is intruded by granitic rock, and the upper unit is faulted against metavolcanic rock of Triassic or Jurassic age. The Carboniferous strata are tightly folded, fold axes plunging subhorizontally in vertical axial surfaces that strike about N. 20° W.

MOUNT MORRISON PENDANT

The Mount Morrison pendant (fig. 12, loc. 2), mapped and described by Rinehart and Ross (1964), contains more abundant and better preserved fossils than any other pendant in the eastern Sierra Nevada (Rinehart and others, 1959). The basal Carboniferous unit, the Bright Dot Formation (about 700 m thick) of Pennsylvanian(?) age, is composed of a lower unit of pyritic muscovite, siliceous hornfels, and metachert and an upper unit of calc-silicate hornfels alternating with siliceous hornfels. The Bright Dot Formation shows a gradational to sharp contact with the overlying Mount Baldwin Marble of Pennsylvanian age. This unit is composed of a 165-m-thick, medium-gray or bluish-gray fine-grained marble. Nodular chert is locally abundant and commonly contains impressions of brachiopod and crinoid molds. The Mount Baldwin Marble is conformably overlain by the 165- to 300-m-thick Mildred Lake Hornfels of Pennsylvanian and (or) Permian age. The Mildred Lake Hornfels is predominantly light- to dark-gray, massive siliceous hornfels that weathers red-brown but locally includes abundant greenish-gray calc-silicate hornfels layers. The Mildred Lake Hornfels is conformably overlain by the 300-m-thick Lake Dorothy Hornfels also of Pennsylvanian and (or) Permian(?) age. The Lake Dorothy Hornfels is composed of thin-bedded, microgranular plagioclase-pyroxene-quartz

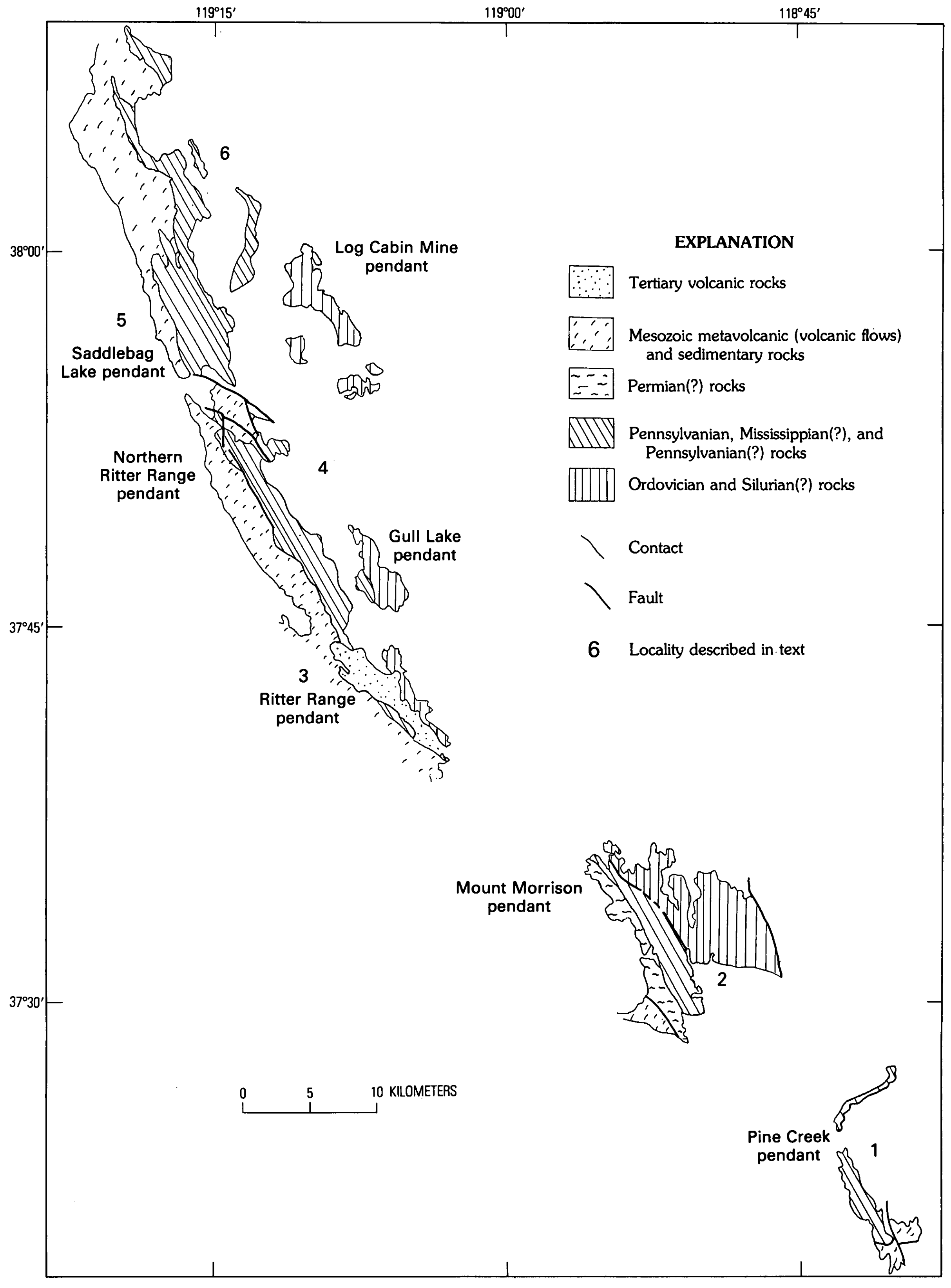

FIGURE 12.—Generalized geologic map showing distribution of Carboniferous rocks and adjacent stratified rocks in roof pendants in the eastern Sierra Nevada, California.

hornfels and alternating grayish-black and yellowish-gray layers that give the rock a striped appearance. The Lake Dorothy Hornfels is conformably overlain by the 1,000-m-thick Bloody Mountain Formation of Permian (?) age. The Bloody Mountain Formation is composed of medium- to dark-gray massive calc-silicate hornfels and siliceous hornfels that weather dark red-brown. Layers of marble are common in some intervals of this formation, and conglomerate lenses are sparse.

The basal Carboniferous unit, the Bright Dot Formation, is faulted against Ordovician and Silurian (?) metasedimentary rock on the east side. The upper unit, the Permian (?) Bloody Mountain Formation, is faulted against Triassic or Jurassic metavolcanic rock on the west side. Rinehart and Ross (1964) mapped tight northwest-trending folds in the Carboniferous rocks, and Russell and Nokleberg (1977), in their structural analysis, described two deformations in the Carboniferous rocks. The first and dominant deformation resulted in upright folds whose axes plunge subhorizontally in vertical axial surfaces striking about N. 20° W. The second deformation resulted in minor upright folds whose axes plunge moderately northwest or southeast in axial surfaces striking about N. 60° W. Rinehart and Ross (1964) and Russell and Nokleberg (1977) noted that the northwest-striking structures in the Carboniferous rocks diverge from the predominantly north-striking structures in the Ordovician and Silurian (?) rocks to the east. Both pairs of authors proposed that the Laurel-Convict fault, which separates the lower and upper Paleozoic rocks, occurred along an angular unconformity.

SOUTHERN RITTER RANGE PENDANT

The south end of the Ritter Range roof pendant (fig. 12, loc. 3), mapped and described by Huber and Rinehart (1965), contains Carboniferous rocks along its east margin. Hornfels and marble in this area were correlated with Pennsylvanian and Permian (?) rocks in the Mount Morrison pendant about 20 km to the southeast. Lithologies in the southern Ritter Range pendant are predominantly moderately well-bedded, dark-gray siliceous hornfels and lesser amounts of interbedded pelitic hornfels, slate, quartzite, chert, and small outcrops of crinoidal marble, similar to the Mount Baldwin Marble. Because of extreme deformation and poor exposure, the stratigraphic sequence is poorly understood. Beds strike about N. 30° W. in the Pennsylvanian and Permian (?) rocks, but strike north in isolated outcrops of Ordovician (?) and Silurian (?) metasedimentary rocks east of the roof pendant. Tertiary volcanic rocks and alluvium conceal the contact between lower and upper Paleozoic rocks. The Carboniferous rocks may be unconformably overlain by Triassic (?) metavolcanic rocks to the west. Fragments and cobbles of the underlying metasedimentary rocks are found in the basal conglomerate of the metavolcanic section (Huber and Rinehart, 1965).

NORTHERN RITTER RANGE AND GULL LAKE PENDANTS

The northern extension of the Ritter Range pendant (fig. 12, loc. 4), mapped and described by Kistler (1966a, b), contains a sequence of bedded quartzofeldspathic hornfels, calc-silicate hornfels, carbonaceous-cherty marble, and chert. These rocks are correlated with the Pennsylvanian and Permian (?) strata of the Mount Morrison pendant. In the central part of the northern Ritter Range pendant, four varieties of crinoids, two brachiopod genera, and one pelecypod genus that were recently found in an encrinal chert, establish a late Paleozoic, possibly Mississippian age (Mackenzie Gordon, Jr., written commun., 1977). In the Gull Lake pendant, Kistler (1966a, b) correlated a carbonaceous marble, a calc-silicate hornfels, quartzite, and quartzofeldspathic hornfels unit, and a marble and calc-silicate hornfels unit with the Pennsylvanian and Permian (?) strata exposed in the Mount Morrison pendant. In contrast, Rinehart and Ross (1964) had correlated these strata with Ordovician and Silurian (?) strata in the Mount Morrison pendant. The original correlation of Rinehart and Ross (1964) is followed here. The contact between the lower and upper Paleozoic rocks is exposed only in the Gull Lake pendant, where the upper Paleozoic (Carboniferous) rocks are now interpreted as resting depositionally on Ordovician and Silurian (?) strata. The Carboniferous rocks are in turn unconformably overlain by Triassic metavolcanic rocks. The basal unit of the metavolcanic rocks is, in some places, a conglomerate containing cobbles similar to lithologies in the underlying Carboniferous sequence. In the eastern Sierra Nevada, the contact between the Mesozoic metavolcanic rocks and the Paleozoic metasedimentary rocks is an angular unconformity that is sheared and locally faulted. Morgan and Rankin (1972) suggested that the contact between these two groups of rocks is everywhere a fault, whereas Brook, Nokleberg, and Kistler (1974) interpreted the contact as an unconformity that is locally faulted.

Carboniferous rocks within the northern extension of the Ritter Range pendant are tightly folded, and the geometry of bedding is dominated by a statistical fold axis plunging gently northwest in a vertical axial surface that strikes N. 30° W. These folds diverge from the folded north-striking strata in the Ordovician and Silurian(?) rocks below.

SADDLEBAG LAKE PENDANT

The Saddlebag Lake pendant (fig. 12, loc. 5), mapped and described by Brook (1977), contains Carboniferous strata consisting of pelitic, quartzofeldspathic, siliceous, and minor calc-silicate hornfels. At the north end of Saddlebag Lake, crinoid fragments and corals recently found in an encrinal limestone were identified by J. W. Durham as possibly Mississippian in age (C. A. Brook, oral commun., 1977). The Carboniferous rocks are tightly folded, the dominant, older folds plunging gently northwest in a vertical axial surface that strikes N. 20° W. The Carboniferous rocks are intruded by Upper Triassic granitic rocks to the east and are unconformably overlain by Permian and Triassic metavolcanic rocks to the west. Fragments and cobbles of the underlying metasedimentary rocks are found in the basal conglomerate of the metavolcanic rock section.

The north end of the Saddlebag Lake pendant (fig. 12, loc. 6), mapped and described by Chesterman (1975) and Chesterman and Gray (1975), was separated into three units: (1) quartzofeldspathic hornfels, calc-silicate hornfels, biotite schist, and minor marble; (2) quartzite, quartzofeldspathic hornfels, metaconglomerate, and minor calc-silicate hornfels; and (3) mineralized encrinal limestone containing crinoid stems of probable Carboniferous age. These three lithologic units unconformably overlie Ordovician and Silurian(?) metasedimentary rocks and are unconformably overlain by Triassic metavolcanic rocks. The contacts with older and younger wallrock units, however, are obliterated by granitic intrusions.

SUMMARY

In the eastern Sierra Nevada, Carboniferous rocks of Mississippian(?) and Pennsylvanian age crop out in a northwest-trending belt approximately 110 km long, but averaging only 2 to 3 km in width. Everywhere to the east, the adjacent stratified rocks are older and include either Ordovician, or Ordovician(?) and Silurian(?) metasedimentary rocks. Everywhere to the west, the adjacent stratified rocks are Triassic and (or) Jurassic metavolcanic rocks. In the Mount Morrison pendant, the Permian(?) metasedimentary rocks are found between Carboniferous rocks and Mesozoic metavolcanic rocks to the west. Permian metasedimentary rocks may be found in all the pendants. Fossils useful in dating the rocks are absent.

Lithologies in the Carboniferous belt are predominantly quartzofeldspathic, siliceous, and calc-silicate hornfels. Marble, quartzite, conglomerate and chert are sparse. Weathering of disseminated pyrite, common in quartzofeldspathic hornfels, imparts a red-brown color to these otherwise dark-gray rocks. Volcanic rocks are not recognized in this Carboniferous sequence. A minor facies change may be manifested by more abundant quartz sandstone in the southern part of the belt. Sparse diagnostic fossils indicate that the northern part of the belt may include Mississippian rocks not recognized to the south.

Carboniferous strata in the eastern Sierra Nevada are tightly folded, and detailed structural studies in several of the pendants indicate at least two episodes of deformation (Brook, 1977; Kistler, 1966a; Russell and Nokleberg, 1977). The more prominent and older deformation is defined by trends of lithologic units, structural geometry of bedding, major folds, faults, and related minor structures. The most pronounced structures of the older deformation are tightly appressed folds having subvertical axial surfaces that strike N. 20°–30° W. and fold axes that plunge subhorizontally or gently northwest. This deformation, which also folded the unconformably overlying Triassic metavolcanic rocks, took place during the Middle Triassic. This timing is defined in the Pine Creek pendant where N. 20°–30° W.-trending major folds were intruded and warped by forceful emplacement of Upper Triassic plutons (Bateman, 1965; Evernden and Kistler, 1970). Folds of similar style and orientation in the Saddlebag Lake and northern Ritter Range pendants are intruded by Upper Triassic plutons in the Mono Craters quadrangle (Kistler, 1966a; Evernden and Kistler, 1970).

The abrupt eastern boundary of outcrops of Carboniferous rocks in some places is parallel to the dominant N. 20° W. trend of folds and faults in the belt. In the Mount Morrison pendant, the eastern limit of Carboniferous rocks is a major fault. Rinehart and Ross (1964) suggested that this contact of Carboniferous rocks with adjacent Ordovician and Silurian(?) rocks is a faulted unconformity because north-striking bedding and folds in the lower Paleozoic rocks sharply diverge from northwest-striking bedding and folds in the upper Paleozoic rocks. This structural divergence between lower and upper Paleozoic rocks is found along the entire belt.

REGIONAL CONSIDERATIONS

The regional strike of Carboniferous rocks in the eastern Sierra Nevada transects, at a large angle, regional trends of similar rocks of the same age (Poole, 1974; Stevens, 1977; Speed, 1977), the western limit of ensialic continental crust (Kistler and Peterman, 1978), the trace of the Golconda allochthon (Speed, 1977), and the Permian volcanic-arc terrane of Speed (1977) (fig. 13). The transection of these regional trends is probably due to severe, large-scale tectonic dislocations. Considerable tectonic transport is indicated by observations made in the southeast in the Inyo Mountains, where the pre-Middle Jurassic allochthon of the Inyo thrust includes Pennsylvanian and Permian strata (Stevens and Olson, 1972). Triassic emplacement of the Inyo allochthon and deformation of the Carboniferous rocks in the eastern Sierra Nevada are possibly contemporaneous. If the folding and thrusting were contemporaneous, the northwest-trending contact of the Carboniferous rocks with the lower Paleozoic rocks in the eastern Sierra Nevada may be tectonic. If so, the Carboniferous strata were transported from the west to their present position. Similarities of Carboniferous fossils in the eastern Sierra Nevada with Carboniferous fossils of the platform facies in the Great Basin to the east, however, suggest that any thrust displacement need not have been great.

An alternative explanation for the northwest-trending contact between the Carboniferous and Ordovician and Silurian(?) metasedimentary rocks in the eastern Sierra Nevada is a locally faulted regional angular unconformity. A similar regional angular unconformity was described by Speed (1977) for a continental borderland terrane (fig. 13) in northwestern Nevada, west of the Mississippian Antler foreland basin of Poole (1974) and Poole and Sandberg (1977). In this terrane, Pennsylvanian and Permian terrigenous clastic and carbonate rocks were deposited unconformably on deformed lower Paleozoic rocks of the ancestral Antler orogenic belt. The Carboniferous belt in the eastern Sierra Nevada has a similar unconformable relation to the underlying lower Paleozoic rocks. If the Carboniferous strata in the eastern Sierra Nevada are correlative with the continental borderland terrane

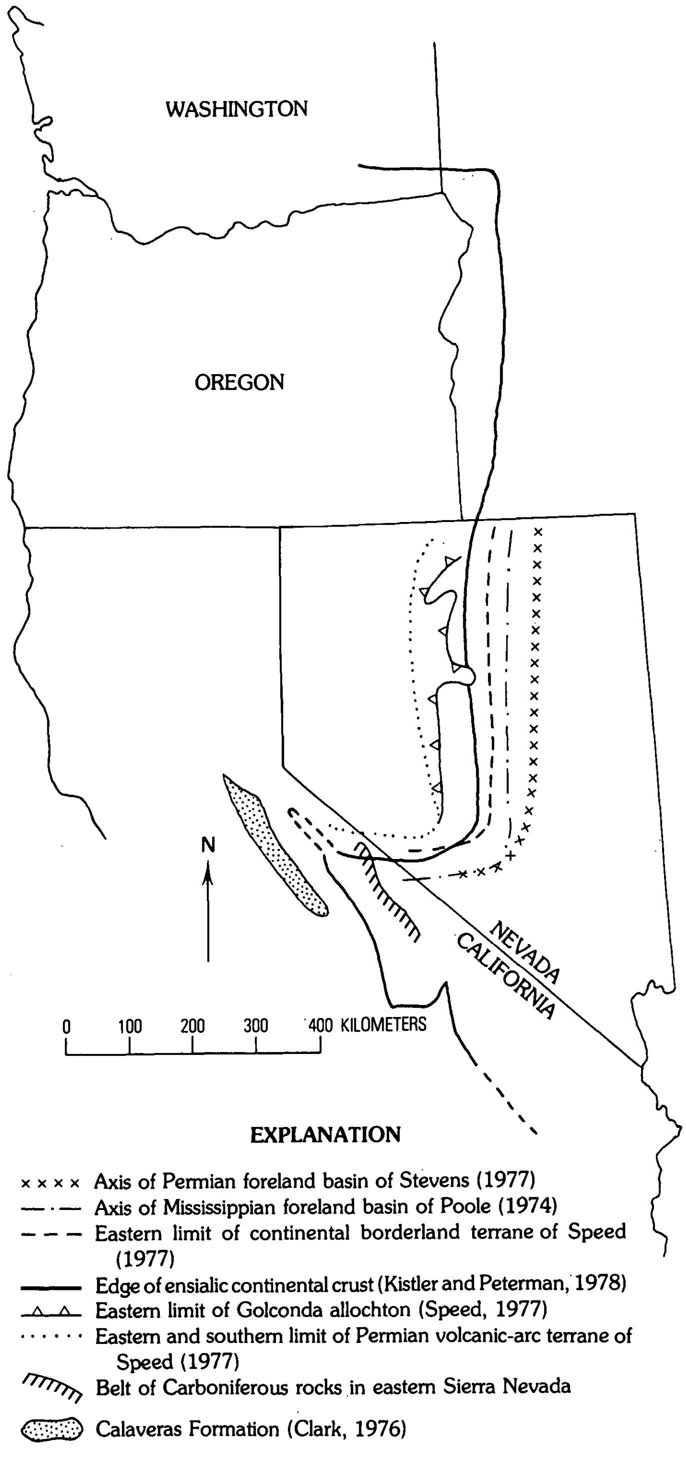

FIGURE 13.—Diagram showing relation of belt of Carboniferous rocks in the eastern Sierra Nevada to adjacent belts of similar age.

of Speed (1977), they need not have been thrust from the west to their present position.

The Calaveras Formation of Carboniferous age is found to the west in a northwest-trending belt in the foothills of the central and northern Sierra Nevada (Clark, 1976) (fig. 13). The main mass of the Sierra Nevada batholith separates the two belts. The lithologies and deformational histories of the two belts differ greatly. Further studies are required to explain these strong differences in parallel-trending belts of Carboniferous rocks.

CARBONIFEROUS ROCKS OF THE NORTHERN SIERRA NEVADA, CALIFORNIA

By JAD A. D'ALLURA and ELDRIDGE M. MOORES

ABSTRACT

Paleozoic rocks of the northern Sierra Nevada are present in two belts, an eastern and a western, separated by a peridotite belt, the last possibly partly of late Paleozoic age. In the eastern belt, Carboniferous rocks are represented mainly by a chert-epiclastic sequence intercalated between thick upper Devonian(?)-Mississippian and Permo-Triassic andesitic volcanic sequences. This stratigraphic sequence may reflect island-arc volcanism in Devonian time, formation and subsidence of a remnant arc in Carboniferous time, and renewal of arc volcanism in Permian time.

Known Carboniferous rocks in the western belt are limited to isolated exposures of carbonate rocks, which generally represent exotic blocks in a chaotic deposit.

INTRODUCTION

In the northern Sierra Nevada, Paleozoic rocks are present in two main terranes, the eastern and the western belts (see fig. 14). The eastern belt (figs. 15 and 16) is best known north of lat 38° 30′ N., where it consists generally of a stratigraphic sequence, from bottom to top as follows (fig. 15): a pre-Devonian sequence consisting of quartzose sandstone, lithic sandstone, phyllite, chert, limestone, and minor volcanic mafic and ultramafic rocks generally assigned to the Shoo Fly Formation and its equivalents, overlain by Devonian-Carboniferous, Permo-Triassic, and Jurassic volcanic-sedimentary sequences. The eastern belt is bordered on the west by the Feather River peridotite belt, a 15 × 150-km belt of metamorphosed ultramafic and associated mafic rock. The only published radiometric age date on igneous rocks in this belt is a single Pb-U date of 300 m.y. on an albitite (Weisenburg and Ave Lallemant, 1977). West of this peridotite belt and its possible southeast extension (see fig. 14), rocks of Carboniferous age are present in a generally chaotic unit, referred to as the Calaveras complex (Schweickert and others, 1977). These terranes as herein defined differ significantly from the original usage of Diller (1908) and Turner (1897) and follow more closely that of Heitanen (1974, 1976) and Moores (in press).

EASTERN BELT

In the northeastern Sierra Nevada, pre-Late Jurassic rocks consist of a 120-km-long exposure of thick volcanic-sedimentary sequence (figs. 14, 15, and 16), which was isoclinally folded and thrust eastward during the Late Jurassic "Nevadan" deformation. All rocks generally show greenschist-facies metamorphic assemblages, as well as relict mineralogy of the original rocks. Primary textures are largely preserved and facilitate paleoenvironmental reconstruction. North of the Lakes Basin area (fig. 14), the rocks crop out in two structural blocks separated by the Grizzly Mountain fault. Near and south of the Lakes Basin, the rocks form a single belt that dips steeply to the east.

Carboniferous rocks in this part of the Sierra Nevada belong primarily to the Peale Formation (Diller, 1908; McMath, 1966), though part of the over- and underlying rocks may also be of Carboniferous age (see fig. 15). Peale rocks grade into the over- and underlying volcanic sequences (see fig. 15). They include deposits of both submarine and possibly subaerial volcanic rocks and nonvolcanic chert, slate, lithic wacke, breccia, and conglomerate. The thickness of the unit is variable but ranges from a minimum of less than 480 m to a maximum of more than 1,100 m (Diller, 1908; Turner, 1894, 1896, 1897; Lingren, 1900; Durrell and Proctor, 1948; McMath, 1966; Durrell and D'Allura, 1977; and D'Allura, Moores, and Robinson, 1977).

PEALE FORMATION

The Peale Formation has been divided into two members: the lower one is composed primarily of feldspar-phyric trachyte or (quartz) latite flows, tuff, and breccia; the upper unit is composed of manganiferous chert and epiclastic sedimentary rocks as well as local pyroclastic rocks (McMath, 1966; D'Allura, 1977). Neither member is equally thick or represented in all areas.

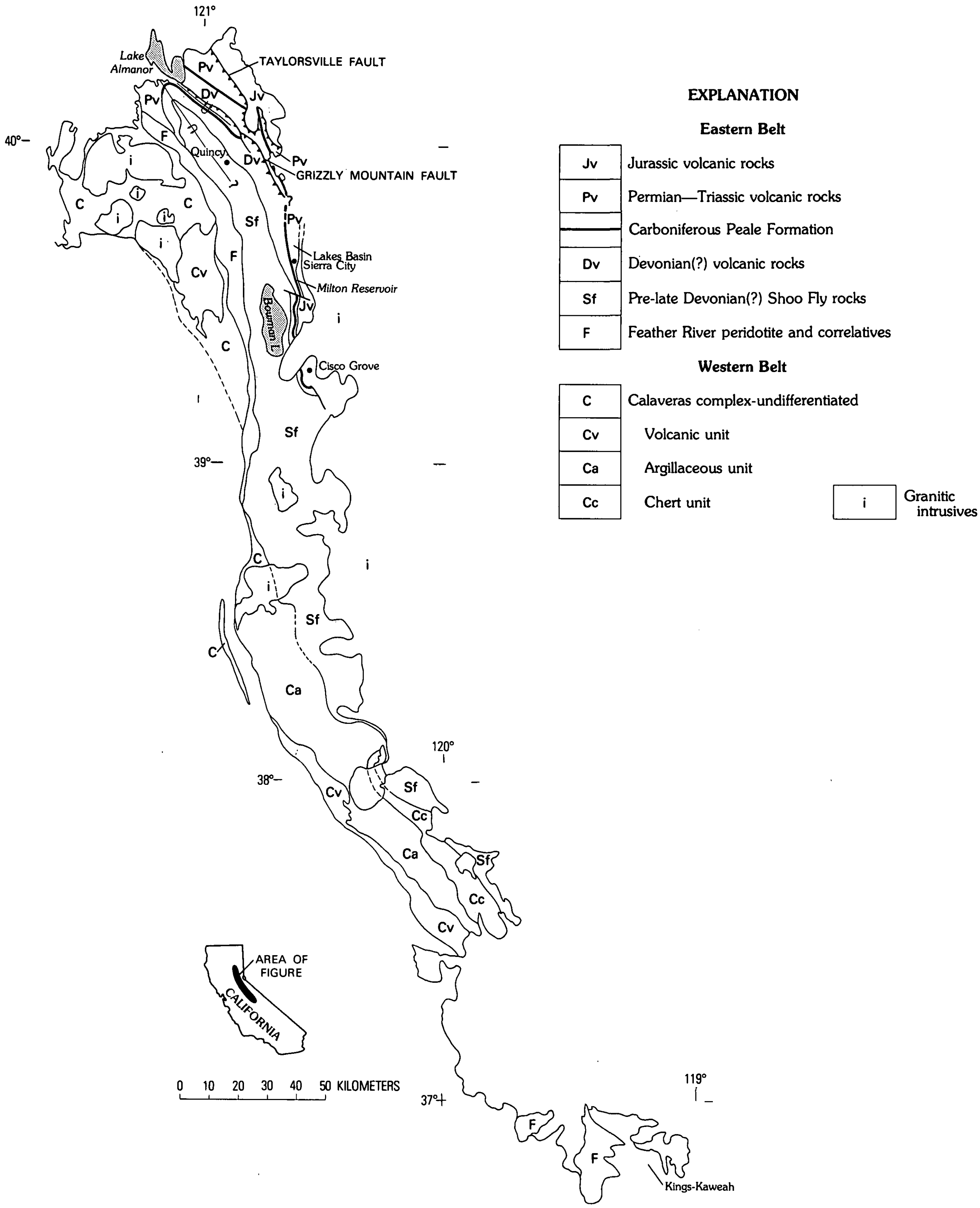

FIGURE 14.—Generalized map of prebatholithic units, northern Sierra Nevada, California.

FIGURE 15.—Simplified description of Paleozoic rocks of the northeastern Sierra Nevada, California.

Formation	Description	
Robinson	Gray or black, green andesitic wacke, conglomerate, or breccia; some calcareous horizons. Permian fossils found in Taylorsville area.	Arlington-green volcaniclastic sandstone, shale, mafic volcanics, silicic tuffs.
Reeve	Gray to dark green andesitic flows, pillows, and volcaniclastic rocks; characterized by large platey plagioclase phenocrysts. Middle Permian fossils found in Taylorsville and Quincy areas as well as northeast of the Lakes Basin area.	
Goodhue	Dark green to nearly black andesitic basalt pillows, breccia and tuff; augite, plagioclase, and pseudomorphed olivine phenocrysts. Late Pennsylvanian to Early Permian fossil found in boulder, Lakes Basin area.	
Peale		
upper member	Upper calcareous horizon, conglomerate or breccia, lithic wacke and purple siltstone overlying red, gray, white, or black chert; some volcanic admixture; composition variable. Early Mississippian fossil found at base of unit in Taylorsville area; Early Permian fauna in probable upper member rocks, Bowman Lake area.	
lower member	Dark gray to black or green trachyte to (quartz) latite flows, pillows, and volcaniclastic rocks; locally contains pink alkali feldspar phenocrysts.	
Taylor	Green andesite pillows, flows and volcaniclastic rocks; augite, diopsidic augite and plagioclase phenocrysts.	
Elwell	Black chert containing phosphate nodules; silicic and andesitic volcaniclastic rocks. Late Devonian fossils, Lakes Basin area.	
Sierra Buttes	Light green dacite and rhyolite volcanic rocks.	
Shoo Fly	Quartz wacke, pelitic rocks, and chert. Locally contains ultramafic rocks and serpentinite.	

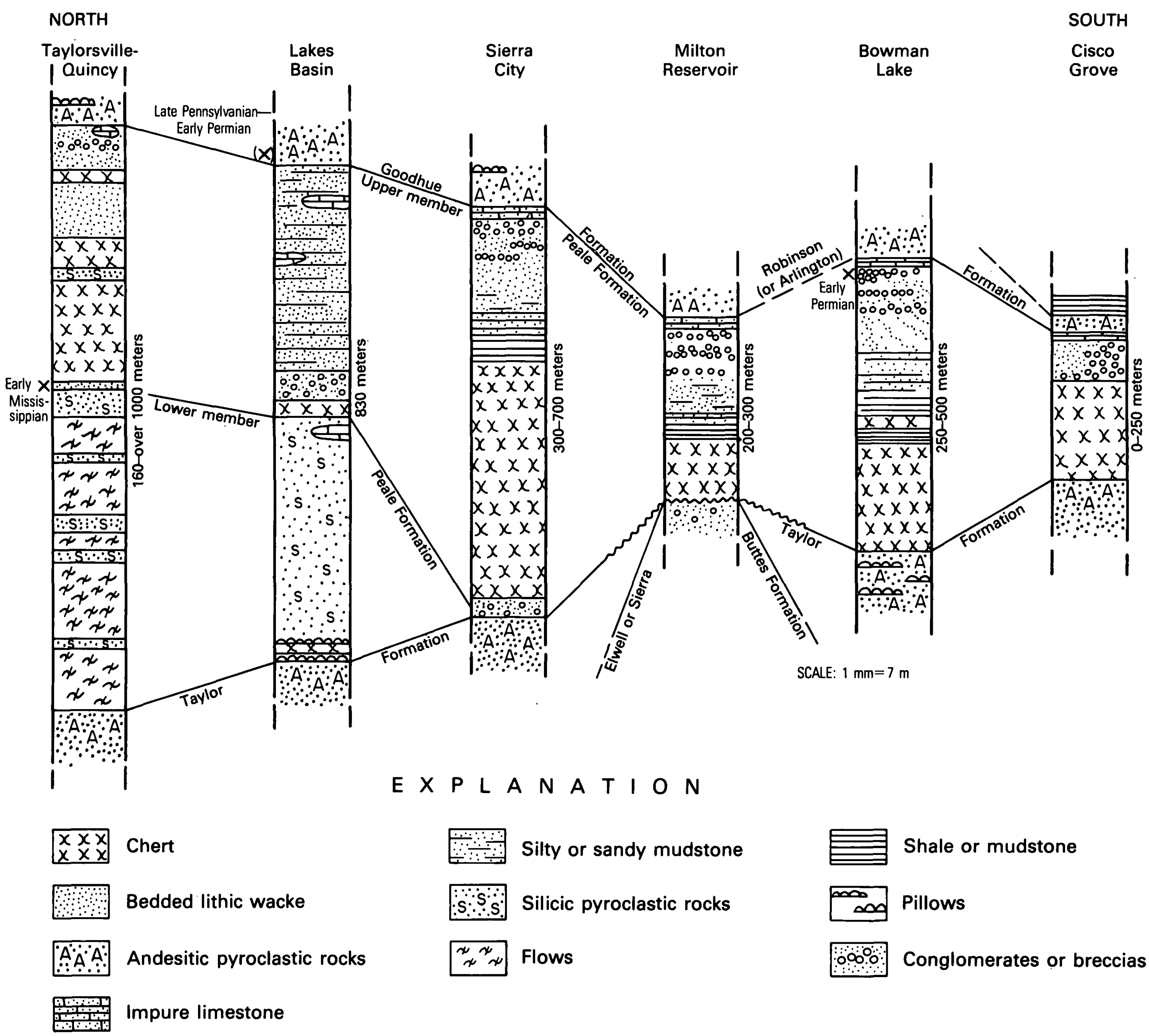

FIGURE 16.—Generalized correlation section of Carboniferous rocks from Taylorsville to Cisco Grove area, northeastern Sierra Nevada, California. Rock types are highly variable in most areas and are only schematically shown. Thickness of the calcareous unit is highly exaggerated.

LOWER MEMBER

Silicic volcanic rocks of the lower member of the Peale Formation intrude and overlie Taylor andesite. The thickness of this unit ranges from zero to more than 1,000 m. Purple-black and dull-gray to pale-green trachyte or (quartz) latite dikes, flows, and pyroclastic rocks are most common, though the character of the rocks changes from area to area. Most rocks are characterized by small phenocrysts of equant to lath-shaped white to pink alkali feldspars in an aphanitic groundmass.

Exposures of the lower member are restricted from the Lakes Basin north; the unit is best exposed in three areas: in the Lakes Basin, east and northeast of Quincy, and south of Lake Almanor. On both sides of the Grizzly Mountain fault, north of lat. 40° N., the lower member consists of flows and comagmatic breccias. East of the fault, pink feldspar-bearing, light- to moderately green tuff and breccia dominate to the north, whereas purple-black, fine-grained shallow intrusive and extrusive rocks containing minor amounts of pyroclastic rock are most common to the south. In the Lakes Basin, lower Peale rocks are composed of reddish-purple to black pillow lava overlain by graded silicic tuff and volcanic breccia (Durrell and D'Allura, 1977). Except for some thin beds of ash and volcanic breccia east of Sierra City, lower member rocks do not crop out south of the Lakes Basin.

Three probable vent or near-vent facies have been recognized—one 11 km north-northeast of Quincy, another 16 km east of Quincy, and a pillowed locality in the Lakes Basin area. The former two areas are characterized by fine-grained shallow sills and dikes, flows, and comagmatic pyroclastic rocks.

In thin section, most feldspar is pink, perthitic, or anti-perthitic, although small white albitized plagioclase laths also are common in some areas. The pink or red color results from minute hematite inclusions. The groundmass may show aphanitic or trachytic textures but is invariably largely recrystallized to mosaics of quartz and albite with muscovite, magnetite, and hematite. Calcite and ferrostilpnomelane are common accessory minerals. Micropoikilitic quartz and albite which has probable perlitic or orb texture is preserved by dusty halos of ferrostilpnomelane; these features suggest that at least a few of the rocks were originally glassy. The dark-gray to black or green color in the rocks results from abundant magnetite or ferrostilpnomelane or green biotite, respectively. Chemically the rocks are quartz normative, although modal quartz is rare. Quartz veins or accumulations are common locally and suggest silicification, the finer grained pyroclastic rocks having been most affected.

UPPER MEMBER

The upper member of the Peale Formation consists of a distinctive assemblage from bottom to top of chert, clastic rocks, and calcareous deposits. Each of these rock types provides the basis for a separate unit, as described below. The proportion of rock types, as well as the clast type in the clastic unit, varies from place to place.

Chert unit.—In the lowest of the three units of the Peale upper member, chert is the most common and laterally continuous rock type, ranging in thickness from 20 to 350 m. The chert may be red, purple, green, gray, black, or white. In places, for example, east of the Grizzly Mountain fault east of Quincy, this unit constitutes a minor part ($\frac{1}{16}$ to $\frac{1}{8}$) of the member, whereas elsewhere, for example, south of Bowman Lake, it dominates the section.

The chert beds generally range in thickness from 10 to 30 cm and are commonly interbedded with thin shale or silty shale. In places, the chert is associated with syngenetic manganese deposits. In some areas, for example, south of Bowman Lake, silty mudstone beds are abundant towards the top of the member. Penecontemporaneous deformation structures are relatively common in both the chert and interbedded fine-grained clastic rocks.

Poorly preserved relict Radiolaria exist in thin section as clear mosaics of quartz. The rocks carry variable amounts of biotite or ferrostilpnomelane.

Clastic unit.—Clastic rocks as much as 350 m thick overlie the chert unit in all areas. They consist of slate, phyllite, lithic and volcanic wacke, and conglomerate and breccia.

Though variable in extent and thickness, the clastic rocks increase in importance toward the south and are especially important near Bowman Lake (where fossils have been discovered in rocks tentatively assigned to this unit). North of the Lakes Basin, olive-drab to black or maroon slate and subordinate volcanic or lithic wacke and tuff make up this unit, which is slightly thinner than the underlying chert unit. In the Lakes Basin area, the unit is represented by 350 m of dark, partly calcareous chloritic slate overlying a thin chert unit. South of the Lakes Basin, silty mudstone, purple muddy siltstone, and lithic wacke with lenses of pebble conglomerate or breccia dominate the Peale exposures.

Most clasts are intraformational, composed of white or gray chert and purple to dark-green siltstone or mudstone. However a small percentage (generally 5 percent or less) of coarser grained nonintraformational clasts are present in all areas.

Though seldom abundant, clasts derived from underlying volcanic units are locally present. Purple volcanic clasts typical of the lower member of the Peale are found in and north of the Lakes Basin area. Clasts of typical bipyramidal quartz-bearing Sierra Buttes dacite and augite-rich green Taylor andesite are sparsely present everywhere and are abundant south of Taylorsville and southeast of Bowman Lake. Clasts of quartz wacke and veined gray to black chert, probably derived from the Shoo Fly Formation, are ubiquitous but are most common near and south of Bowman Lake.

Other clasts not derived from any exposed or recognized units are scattered through the section. These clasts include, in order of decreasing abundance, volcanic rock fragments, fine- to medium-grained diorite, gabbro, and pink granite. Volcanic rock fragments in this group are dominantly silicic in composition, although andesitic or basaltic clasts are common. Amygdaloidal structure in andesites and rare perlitic structure in silicic rocks are preserved.

The volcanic and plutonic clasts dominate at the fossil locality northeast of Bowman Lake. There, amygdaloidal andesitic volcanic rocks, porphyritic green and purple silicic volcanic rocks, purple siltstone, abundant green or red chert typical of the underlying Peale chert, medium-grained plutonic rocks, and minor but significant carbonate organic debris constitute the bulk of the clasts. Although, except in the uppermost beds, clastic hornblende is a rare or subordinate phase in the matrix of most of the conglomerate or breccia beds, it is common in the area south of Milton Reservoir. This hornblende-bearing breccia or conglomerate probably interdigitates with hornblende tuff of the overlying Goodhue Formation. These rocks are probably part of the Peale, not the Goodhue, though their stratigraphic position is by no means certain.

Calcareous unit.—In most areas south of the 40th parallel, a calcareous unit forms the top of the upper member of the Peale Formation. Nowhere more than 35 m thick, this unit consists of gray silty limestone (marble and calc-silicate rock near plutons) containing scattered clasts of angular to subrounded rock fragments, the latter including chert. East of Quincy, the unit is represented by a discontinuously exposed skarn in a roof pendant of a quartz diorite pluton, or by a sparsely calcareous zone east of the Grizzly Mountain fault. East-southeast of Sierra City, the unit is 4–10 m wide and consists of calcareous rocks interbedded with basaltic wacke and chert. At the fossil locality near Bowman Lake, the unit is represented by fossiliferous limestone lenses intercalated with the upper conglomeratic beds of the clastic unit. The unit is not present 2.5 km southeast of Bowman Lake, but it reappears near Cisco Grove.

TAYLOR AND GOODHUE-ARLINGTON FORMATIONS

As parts of the underlying Taylor Formation or the overlying Goodhue-Arlington Formations may have been deposited in Carboniferous time, a brief discussion of them is in order.

The Taylor is composed of pyroxene-bearing basaltic andesite and andesite tuff, volcanic wacke, and breccia. Its composition varies from base to top and from place to place. However, it is generally more mafic in its lower part and more silicic toward the top. A greater volume of silicic andesite in the Lakes Basin area and a more mafic composition than usual in the Bowman Lake area are exceptions to the above trend. Local vents marked by accumulations of pillows and coarse breccia are present southwest of Taylorsville, in the Lakes Basin area, and east and south-southeast of Bowman Lake. The volcanic deposits decrease in grain size west of the Grizzly Mountain fault from Taylorsville north, and south of Cisco Grove.

The basal part of the Goodhue Formation in the Lakes Basin area is composed of green hornblende andesite tuff and fine volcanic breccia. The thickness of this basal unit varies. The unit is present in significant amounts mainly east of Quincy, in the Lakes Basin area, and locally as fine- to medium-grained tuff near Milton Reservoir.

The rocks above the basal hornblende-bearing unit are dark-green olivine pyroxene basalt and andesitic basalt tuff, volcanic wacke, breccia, and flows. The olivine, which commonly shows a good crystal habit, has been totally altered to various metamorphic mineral assemblages including serpentine, epidote, hornblende, ferrostilpnomelane, biotite, and chlorite.

The Goodhue formation generally is present east of the Grizzly Mountain fault from Taylorsville south, and from Quincy south to Milton Reservoir. West of the fault and north of Quincy, the Peale is overlain by rocks assigned to the Arlington Formation.

Arlington rocks consist chiefly of graded volcanic wacke deposits, which in some places show abundant bottom markings. They generally become finer grained and bedded from around Mt. Hough northwestward towards Lake Almanor.

In the exposures south of Milton Reservoir, no rocks are assigned to the Goodhue Formation. However, in this area, thin beds of green fine-grained tuff crop out below easily recognized younger Reeve rocks (see fig. 15) and correlate with andesitic tuff and volcanic wacke that overlie gradationally the upper calcareous unit of the Peale in the Bowman Lake and Cisco Grove areas. Possibly much of this andesitic material may be upper Paleozoic and may be correlative with the Robinson (or Arlington) Formation.

CONTACTS

The contacts between the Peale and the overlying and underlying units appear generally conformable and gradational. The basal contact with the Taylor is intrusive in places but is mostly depositional, marked by a change in volcanic-rock composition.

From the relations described above, clearly the upper contact of the Peale with the overlying volcanic deposits is gradational and complex. Where the contact with the Goodhue is exposed, chert or clastic (or calcareous) rocks are interbedded with crystal-vitric or hornblende basaltic or andesitic tuff, the latter of which grades in turn into dark-green olivine basalt typical of the Goodhue Formation. Where overlain by Arlington rocks, the contact is again gradational, and placing the contact between clastic rocks of the Peale and those of the overlying Arlington is difficult.

Southeast of Bowman Lake, the upper calcareous conglomerate-breccias or breccia-conglomerates are interbedded with green silicified tuff and crystal-lithic tuff of probable Arlington (or Robinson) equivalent volcaniclastic rocks. However, to the south, the Peale rocks are eroded. Here the Arlington (?) volcaniclastic rocks lie directly over the fine-grained purple lithic wacke and silty mudstone with a moderate angular unconformity. The basal volcaniclastic rocks contain rounded pebbles of chert, quartzite, and some aphanitic volcanic rock fragments derived from preexisting rocks.

BIOSTRATIGRAPHY

Fossils in the volcanic-sedimentary complex of the northeastern Sierra Nevada are scarce, and precise age assignments are not possible. However, rock units apparently transgress the boundaries of the Carboniferous System. The volcanic rocks of the Peale lower member and of the Taylor Formation are undated and may be partly or wholly Late Devonian or Early Mississippian in age. The rocks of the Peale upper member contain undated radiolaria; however, these rocks may range in age from Early Mississippian to possibly Early Permian, on the basis of two fossil localities, separated by 70 km.

Near Taylorsville, McMath (1966) discovered an Early Mississippian brachiopod fauna close to the base of the upper member of the Peale Formation. An undescribed trilobite from the same locality also suggests a similar age (McMath, written commun., 1973). East of Bowman Lake, fusulinids, including a single specimen of *Triticites*, several specimens of *Schwagerina* generally similar to *S. diversiformis*, and a specimen that is probably *Thompsonella* (?) (C. A. Ross, written commun., 1975) were found in the matrix of a chert and volcanic breccia-conglomerate near the top of the upper member (D'Allura and others, 1977). The fauna, which suggests an Early Permian age, may have been reworked from older sediments, but the possibility seems remote as the fossils are discrete entities, have not been silicified, and are associated with individual bryozoan, crinoid, and coral (?) fragments.

The rocks of the basal Goodhue Formation have been dated tentatively as Late Pennsylvanian to Early Permian on the basis of the transported fossil *Helicoprion sierrensis* (Durrell and D'Allura, 1977). Hence Goodhue and Peale rocks may be partly time equivalent.

ENVIRONMENTAL INTERPRETATION

The volcanic-sedimentary rock sequence of which the Taylor, Peale and overlying rocks are part, probably reflects the existence of two volcanic island-arc sequences, one of probable Devonian-Mississippian age (Taylor, lower Peale) and one of probable Permo-Triassic age (Goodhue, Reeve, Arlington) (D'Allura and others, 1977). Hence, earliest Carboniferous time probably was marked in this region by arc-type volcanism.

The gradation from the volcanic lower Peale to the upper Peale, which contains radiolarian chert and some manganese deposits, suggests that in Carboniferous time, arc-type volcanism gradually yielded to a period of pelagic sedimentation marked by only sporadic volcanism (represented by scattered volcanic beds in the chert unit).

The succeeding gradation in the upper Peale from chert to clastic rocks with intercalations of coarse, mostly locally derived, debris suggests increasing crustal instability or a shifting locus of crustal instability. The lenticular nature of individual units, the changes over short distances of thickness and lithology of the clastic unit, and the presence of turbidite or grain flow deposits, suggest local relief possibly within an area of fault-bounded highs and basins. That this relief was not constructional in origin is suggested by the relatively minor volcanic activity. The source of plutonic and some volcanic-rock clasts remains unknown, although it conceivably may have been partly the Feather River peridotite belt to the west.

The gradation in turn from epiclastic deposits to tuffaceous rocks of the overlying Goodhue-Arlington Formations, suggests the gradual onset of volcanism of the Permo-Triassic arc complex after these structural disturbances.

In an earlier article (D'Allura and others, 1977), we proposed a plate-tectonic model whereby the cessation of arc volcanism, followed by pelagic sedimentation, then renewed arc volcanism, represented the formation and subsidence of a remnant arc (Karig, 1974). The range in fossil dates for this cessation in volcanism includes the time of the Antler orogenic episode in central Nevada (Silberling and Roberts, 1962).

WESTERN BELT

Recent work on the western or Calaveras terrane indicates that much of it is made up of a partly chaotic sequence of rocks. Schweickert and others (1977) have separated the Calaveras into a series of units, from west to east, as follows: a volcanic unit composed of mafic pillow lava, pillow breccia, tuff breccia, and bedded tuff, which interfingers locally with so-called "slaty mudstone and diamictite"; an argillite unit, dominantly clastic, composed of meter- to kilometer-sized fragments of quartzite, chert, calcareous rocks, and argillaceous sedimentary rocks; a chert unit, characterized by thick sequences of well-bedded, rhythmic chert containing black argillaceous partings which interfingers with "diamictite"; a quartzite unit, composed of well-bedded quartzite and quartz-rich sandstone, interfingering with chaotic units.

Regional correlations and reconnaissance suggest, however, that the eastern quartzite unit in the southern and central Sierra Nevada may be a southern extension of the Shoo Fly rocks of the eastern belt (Moores, in press); the contact between the western and eastern belts possibly is a suture represented in part by a mylonite zone (Schweickert, 1977) and in part by the Feather River ultramafic belt. This suggested correlation is illustrated on figure 14. One complication with this suggested picture is the possible correlation of the Feather River peridotite with the Carboniferous Kings-Kaweah ultramafic complex to the south (Saleeby, 1977) which is *west* of the main Calaveras terrane, rather than *east* of it, as is the Feather River peridotite.

Most fossiliferous limestone inclusions and masses reflect shallow-water deposition. They yield Carboniferous or Permian fossils, but no systematic distribution of ages within these blocks has been recognized. These limestone blocks and the chaotic units that include them represent either submarine landslide deposits derived from carbonate-bearing volcanic islands or (Schweickert and others, 1977) the juxtaposition of diverse rocks in an accretionary prism.

ACKNOWLEDGMENTS

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CARBONIFEROUS ROCKS OF THE EASTERN KLAMATH MOUNTAINS, CALIFORNIA

By Rodney Watkins

ABSTRACT

The Carboniferous System is represented in the eastern Klamath Mountains of California by the Bragdon Formation and the conformably superjacent Baird Formation. Both formations are sparsely fossiliferous. The Bragdon Formation conformably overlies the Middle Devonian Kennett Formation and, where the Kennett is missing, the pre-Kennett Balaklala Rhyolite. The Bragdon Formation consists primarily of argillite, locally prominent volcaniclastic facies, and minor shelly siltstone. The fossil faunas are Late Mississippian in age. The Baird Formation is conformably overlain by the Lower Permian McCloud Limestone. The Baird Formation consists mainly of volcaniclastic rocks and minor amounts of argillite, shelly siltstone, chert, greenstone, and limestone facies. The fossil faunas range in age from Late Mississippian through Pennsylvanian to Early Permian.

The Bragdon and Baird Formations record a shoaling throughout Carboniferous time. Upward increase in wacke beds in the argillite indicates change from distal to proximal turbidite deposition. The distribution of volcaniclastic facies suggests episodic deposition from local shifting source areas, such as volcanic islands, during a transition from basin and slope to shelf conditions.

The fauna of the two formations—mainly brachiopods, crinoids, bryozoans, mollusks, and corals—is part of an Asiatic province and has few species in common with typical Carboniferous faunas of North America.

INTRODUCTION

The Carboniferous System is represented in the eastern Klamath Mountains of California by the Bragdon Formation and the overlying Baird Formation. These units were named and mapped by Fairbanks (1894), Hershey (1902), and Diller (1905, 1906). Subsequent stratigraphic studies have been made by Kinkel, and others, (1956), Albers and Robertson (1961), Lydon and Klein (1969), and Watkins (1973).

BRAGDON FORMATION

The Bragdon Formation covers extensive parts of Shasta and Trinity Counties, but its base is ex-

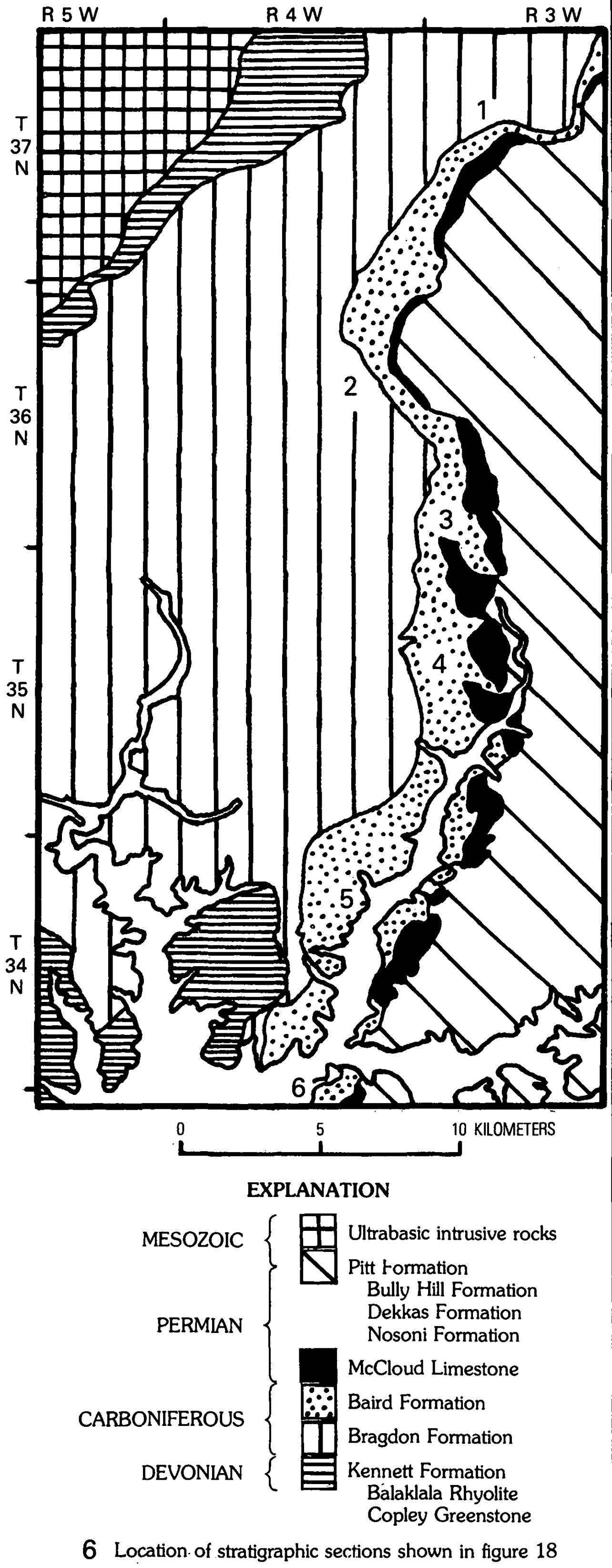

FIGURE 17.—Geologic map around the northern part of Shasta Lake, eastern Klamath Mountains, California.

posed in only a few areas near the Sacramento Canyon (fig. 17). Here it conformably overlies the Kennett Formation, which contains conodonts of Middle Devonian age (Savage, 1976). In places, the Kennett is missing, and the Bragdon conformably overlies the Balaklala Rhyolite of pre-Kennett age. The Bragdon Formation is barren of fossils for most of its geographic and stratigraphic extent. Locally, however, brachiopods of Late Mississippian age occur in the uppermost part of the Bragdon on the east slopes of Hanland Peak (Watkins, 1973, 1974). The upper contact of the Bragdon is mapped at the highest stratigraphic appearance of distinctive, pebble-bearing wacke beds which conformably underlie the Baird Formation. This contact probably is not everywhere the same age.

The Bragdon Formation consists mainly of rocks designated as the argillite facies in figure 18. Several small sections totaling 87 m in thickness were measured in the upper drainage of Middle Salt Creek. Sediment types in these sections are: 33.1 percent argillite that has flat parallel lamination; 8.4 percent silty and tuffaceous argillite that has flat parallel lamination; 0.8 percent silty and tuffaceous argillite containing plant debris; 3.1 percent lithic wacke that has flat parallel lamination and/or low-angle planar cross lamination; 18.2 percent lithic wacke, massive, or in beds graded from coarse sand at base to fine sand at top; 36.4 percent lithic wacke including clasts from pebble to boulder size and a variety of sedimentary structures. Wackes form tabular beds that have sharp flat contacts and basal, flute, and load casts. The thickness and stratigraphic frequency of wacke beds increase up section in the Bragdon Formation, culminating in conglomeratic units 16 to 20 m thick. Bioturbation is notably absent in the argillite facies, except for minor intervals in the upper few hundred meters of the Bragdon. Bioturbated fossiliferous siltstones associated with this facies near Kabyai Creek were described by Watkins (1973). The Bragdon Formation also contains volcaniclastic facies around the summit of Hanland Peak. These sedimentary rocks are identical in character with the volcaniclastic facies of the Baird Formation, which is described below.

BAIRD FORMATION

The Baird Formation (fig. 17) consists of a variety of sedimentary and volcanic rock types. Several hundred meters of the argillite facies is present on Hirz Mountain, distinguished from the argillite fa-

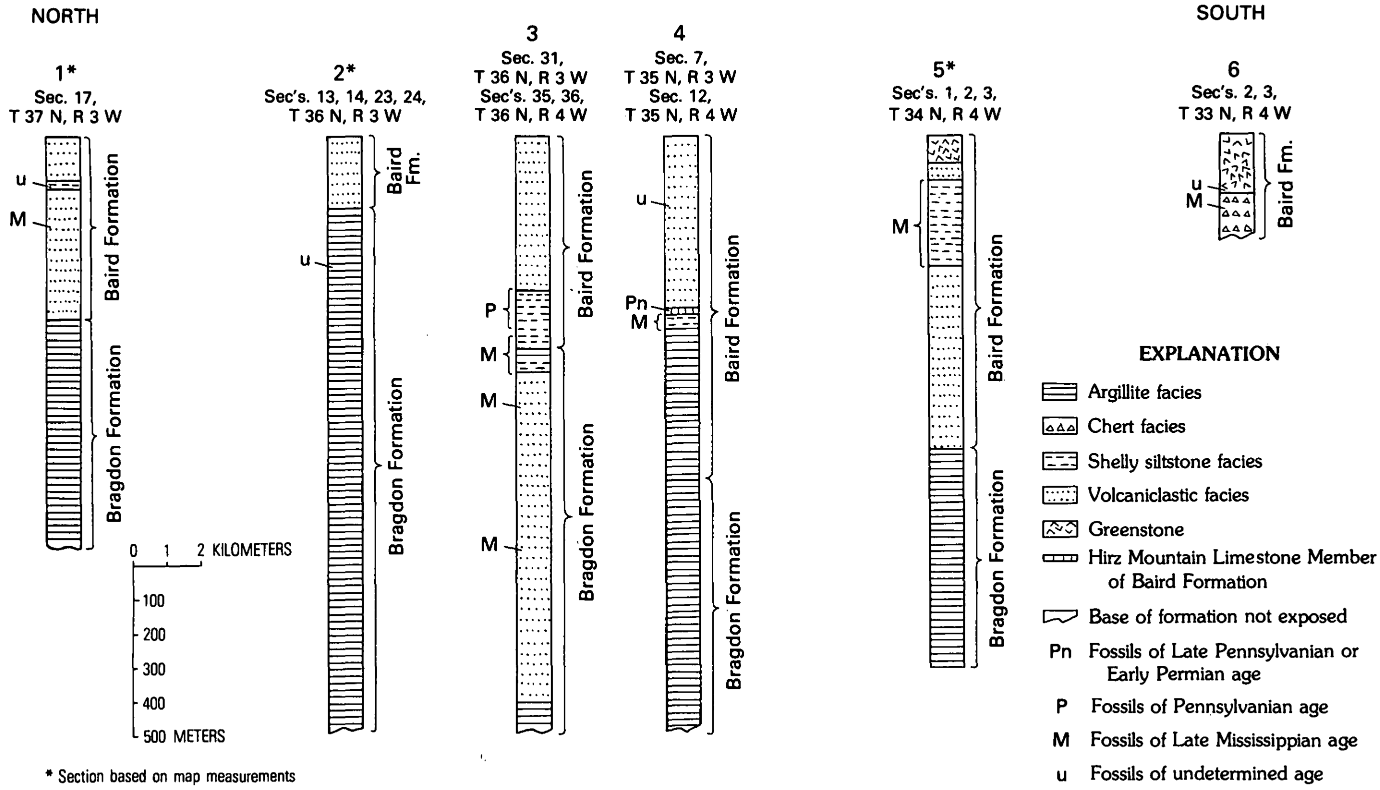

FIGURE 18.—Columnar sections of Carboniferous rocks in the eastern Klamath Mountains, California (see text for description of facies). 1, Tow Dow Creek; 2, High Mountains and headwaters of Middle Salt Creek; 3, east slopes of Hanland Peak and Kabyai Creek; 4, Hirz Mountains; 5, Ycotti Creek and Greens Creek; 6, south of Pitt River Bridge. Age assignments also represent all observed occurrences of shelly fossils. Thickness of section 5 is based on estimates by Albers and Robertson (1961).

cies of the Bragdon by the lack of conglomeratic wacke beds. The Baird Formation also contains a chert facies in the Greyrock section south of Pitt River Bridge. This facies consists of thin tabular beds of chert and cherty argillite, minor bioclastic limestone, and no apparent bioturbation. The shelly siltstone facies has several disjunct occurrences within the Baird, as shown in figure 18, and is the only part of the formation that has bioturbation and common shelly fossils. Fifteen meters of the facies was examined in detail near Greens Creek. It consists of: 50.0 percent very tuffaceous siltstone and fine sandstone that has flat parallel lamination, rare low-angle cross lamination, and rare mollusk shells; 49.9 percent siltstone and silty mudstone that has intense bioturbation and scattered mud-supported shells of many phyla; 0.1 percent closely packed shell debris in thin tabular beds. The shelly siltstone facies also contains limestone units near Kabyai Creek, which have been described by Watkins (1973). The major exposure of the shelly siltstone facies is along the McCloud River between Campbell Creek and Potter Creek. In most years, this area is covered by Shasta Lake.

The volcaniclastic facies is the only unit in the Baird Formation that is continuous along strike (fig. 18). Twenty-five meters of strata was measured in this facies on the southwest side of High Mountain; it consists of: 85.4 percent purple mudstone and silty mudstone, massive, containing floating volcanic lithoclasts; 9.7 percent arenite, in beds graded from coarse sand at base to fine sand at top, sometimes having low-angle cross lamination in upper part; 4.8 percent arenite containing pebbles at base of beds. The arenite beds are tabular, laterally continuous, and have sharp flat contacts. They are composed mainly of volcanic-rock fragments. The fine-grained purple sedimentary rocks are probably a volcanic mudflow, as some of their lithoclasts are intruded by the muddy groundmass of the rock. Tuff and minor fossiliferous limestone are also present in this facies, as described by Watkins (1973). A unit of greenstone forms the upper part of the Baird Formation in the southern part of its outcrop area

(Albers and Robertson, 1961) and is laterally equivalent to the volcaniclastic facies (fig. 18).

Age relations of the Baird Formation are shown in figure 18. The lower part of the Baird includes the brachiopods *Dorsoscyphus, Semicostella,* and *Striatifera,* which indicate a Late Mississippian age (Watkins, 1973, 1974). The higher part of the Baird Formation is usually barren of fossils, but on the east slopes of Hanland Peak it contains fusulinids and brachiopods of Pennsylvanian age (Skinner and Wilde, 1965; Watkins, 1973). Stratigraphic sections are not fossiliferous enough to locate a Mississippian-Pennsylvanian boundary precisely, and sedimentation appears to have been continuous across this interval. The youngest fauna in the Baird Formation is in the Hirz Mountain Limestone Member of Watkins (1973). This limestone crops out for less than 2 km along strike. It contains *Diplanus* and other brachiopods, which suggest a latest Pennsylvanian or Early Permian age. Where contacts have not been faulted or intruded by diorite, the Baird is conformably overlain by the Lower Permian McCloud Limestone. Watkins (1973) summarized evidence for the partial lateral equivalency of the upper part of the Baird and the lower part of the McCloud.

SEDIMENTARY ENVIRONMENTS

The Bragdon and Baird Formation record a general shoaling in the eastern Klamath area throughout Carboniferous time. The argillite facies of the Bragdon is a basin-and-slope deposit. Laminated argillite facies represent slow clay deposition under abiotic conditions, and wacke beds represent turbidite deposition. Upward increase in thickness and complexity of wacke beds in this facies indicates a change from distal to proximal turbidite deposition, and many of the thick conglomeratic wackes in the upper part of the Bragdon probably represent deposits of submarine channels near the top of a slope system. The beginnings of bioturbation and rare transported shelly fossils at the top of the argillite facies also suggest shoaling from basin towards shelf conditions.

The volcaniclastic facies and shelly siltstone facies are shelf deposits, which are found at the top of the Bragdon and throughout the Baird Formation. The volcaniclastic facies represents submarine tuff deposits, volcanic mudflows, and shelf-type turbidite deposits derived from volcanic source areas. Its variations in age, thickness, and stratigraphic relations (fig. 18) suggest episodic deposition from local changing source areas such as volcanic islands. The shelly siltstone facies is a typical bioturbated offshore shelf deposit containing a rich fauna of invertebrates. Environmental variation within the facies is indicated by the presence of several types of benthic invertebrate communities (Watkins, 1973, 1974).

Shoaling in the eastern Klamaths culminated in Early Permian times with the spread of shallow shelf carbonate deposits across the area. This environmental transition is best preserved in sections along the middle part of Nawtawaket Creek. Here, the upper part of the volcaniclastic facies of the Baird Formation is interbedded with graded beds of redeposited limestone, and the entire sequence is overlain by thick massive carbonate rocks of the McCloud Limestone. These relations suggest that the McCloud was deposited at shallower depths than the volcaniclastic facies of the Baird and that downslope transport of carbonate sediment into terrigenous and volcanic deposits took place. The petrography of the McCloud also indicates very shallow shelf conditions (Demirmen and Harbaugh, 1965).

FAUNAL AFFINITIES

The occurrence of invertebrate faunas in the Bragdon and Baird Formations is shown in figure 18. Horizons containing fossils are relatively few and are separated by hundreds of meters of barren sedimentary rocks. The faunas are dominated by brachiopods, crinoids, bryozoans, mollusks, and corals; extensive species lists are contained in Smith (1894) and Watkins (1973). Taxonomic treatment of this fauna is limited, however, to descriptions of a few brachiopod, bivalve, and trilobite species by Muir-Wood and Cooper (1960), Dutro (1955), Wheeler (1935), and Watkins (1974, 1975).

Biogeographically, the fauna of the Bragdon and Baird Formations is part of an Asiatic province and has few species in common with typical Carboniferous faunas of North America (Smith, 1894; Watkins, 1974). Watkins (1973, 1974) described eight invertebrate communities from the Bragdon and Baird that are characterized by differences in species composition, species diversity, trophic relations, and sedimentary occurrence. Additional undescribed communities in the Baird Formation are found southwest of Tater Hill and along Tom Dow Creek.

CARBONIFEROUS FORMATIONS IN OREGON

By Ewart M. Baldwin

ABSTRACT

Carboniferous strata are known in Oregon only in the Suplee area near the head of Crooked River, about 25 km southeast of Paulina. Two formations have been named. The older is the middle and upper Meramecian and Chesterian Coffee Creek Formation, composed mainly of limestone, and the younger is the Morrowan Spotted Ridge Formation, composed of conglomerate, sandstone, and plant-bearing mudstone.

Beds within the Elkhorn Ridge Argillite in the Durkee quadrangle contain a fusulinid fauna assigned to the Desmoinesian stage of the Pennsylvanian. All Carboniferous strata are probably allochthonous and were emplaced during plate movements.

INTRODUCTION

Carboniferous strata are known only from central Oregon, with a possible occurrence in northeastern Oregon. Early discoveries of Carboniferous strata are usually credited to Thomas Condon (1902). E. L. Packard (1928, 1932) and his students found and described many other localities. The beds crop out at the head of Crooked River near Suplee (fig. 19). The type sections of the Carboniferous formations are approximately 25 km southeast of Paulina in areas only partly covered by detailed topographic maps.

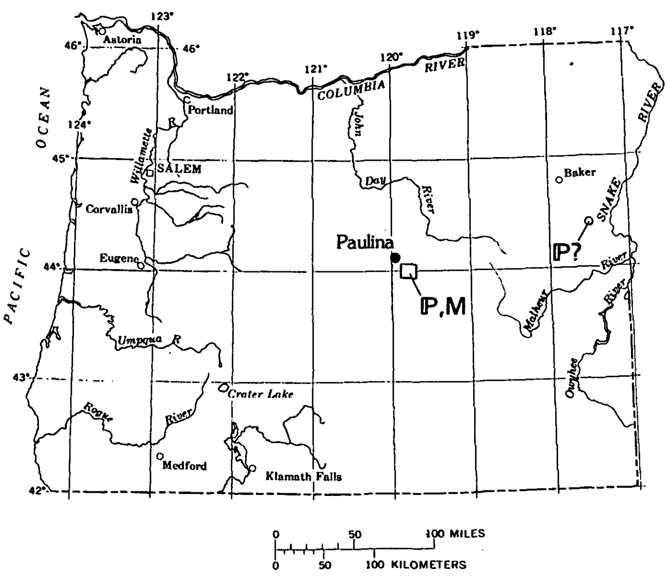

FIGURE 19.—Index map of Oregon showing locality of the Elkhorn Ridge Argillite, Pennsylvanian? (ℙ?); Spotted Ridge Formation, Pennsylvanian (ℙ); and Coffee Creek Formation, Mississippian (M).

Merriam and Berthiaume (1943) mapped and described the Mississippian, Pennsylvanian, and Permian rocks along upper Grindstone Creek, a tributary of Crooked River. They gave the name Coffee Creek Formation to the Mississippian limestone beds containing the large brachiopods noted by Packard. The Pennsylvanian plant-bearing beds associated with the conglomerate and sandstone were named the Spotted Ridge Formation, and the overlying Permian limestone was called the Coyote Butte Formation. Middle Devonian beds nearby were reported later by Kleweno and Jeffords (1962), but they have not been formally named.

The most detailed mapping of the Paleozoic units has been by H. J. Buddenhagen, whose map is on open file at the State of Oregon Department of Geology and Mineral Industries in Portland. Figure 20 is a part of that map. Buddenhagen (1967) published a summary of his work. The Suplee area was mapped by Dickinson and Vigrass (1965) who dealt mainly with the Mesozoic strata. Paleozoic formations are shown on the geologic map of eastern Oregon east of the 121st meridian (Walker, 1977).

Structural relationships are difficult to determine because of intense deformation and because the rolling grass-covered terrain presents an outcrop area of only about 10–15 percent. Thus, relationships between units must be inferred. A recent summary of the Paleozoic units of central and northeastern Oregon by Vallier, Brooks, and Thayer (1977) discusses lithology, stratigraphic units, and their ages. They conclude that contacts are commonly thrust planes and that faunas of Tethyan and North American affinities are intermixed because of plate movements.

COFFEE CREEK FORMATION

The Coffee Creek Formation was named by Merriam and Berthiaume (1943) for a small tributary of Grindstone Creek, southeast of Wade Butte (fig. 20). The type section is in sec. 30, T.18 S., R. 25 E. The area had been covered in reconnaissance by Packard (1928, 1932), but the most detailed mapping is by Buddenhagen (fig. 20).

The Coffee Creek Formation is in fault contact with the Spotted Ridge, Coyote Butte, and Triassic formations. No depositional contacts are known, but

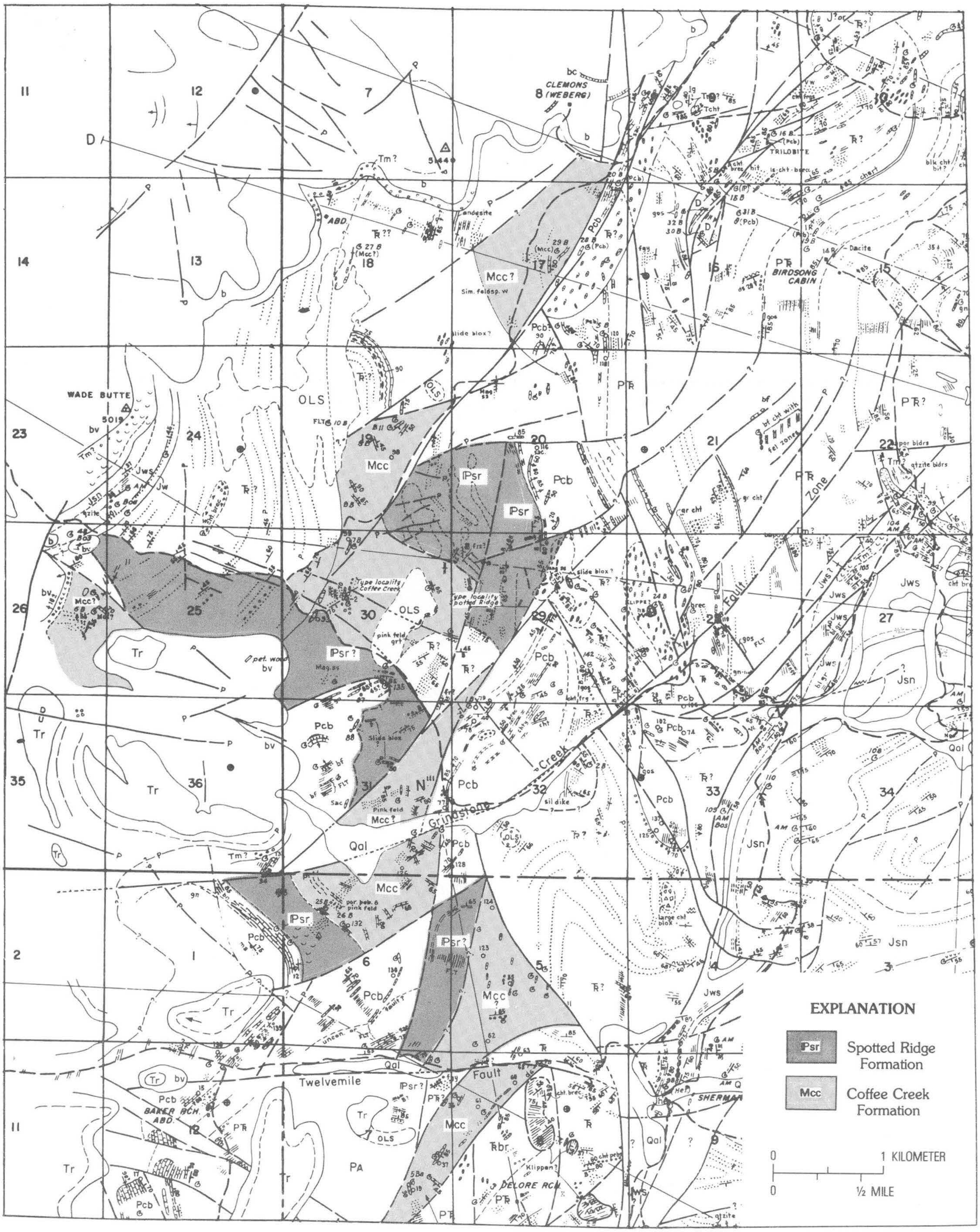

FIGURE 20.—Carboniferous formations in central Oregon. Areas of Carboniferous rocks are shaded. Geology by H. J. Buddenhagen.

the Triassic beds in Wade Butte appear to rest on the Mississippian limestone on the west. No underlying strata are known in juxtaposition.

Merriam and Berthiaume described the type section as follows:

A conspicuous linear outcrop at the type section shows about 75 feet of strata dipping about 53° NW. Within the upper 40 feet very dark gray to black argillaceous and carbonaceous limestones are in places rather thin-bedded. These deposits are locally packed with the brachiopod *Striatifera.* Forty-five feet from the top of the limestone section is a bed of very sandy limestone with *Striatifera* immediately below which lies the main *Gigantella* horizon. The usually large productid *Gigantella* is very profuse here in a fine-grained and rather pure limestone of deep neutral gray; the large brachiopod is apparently restricted at this point to a bed varying in thickness from one to two feet. Within and below the *Gigantella* bed solitary and compound rugose corals are abundant; these are embedded in gray limestone or immediately below in argillaceous limestones weathering to a light grayish brown. In the lower 30 feet of this section these argillaceous layers are interbedded with purer gray limestones, while toward the base sandy limestones and calcareous sandstone predominate.

The Formation is traceable along the strike for approximately 2 km, and Merriam and Berthiaume (1943, p. 151) estimated a thickness of 275–300 m.

The Coffee Creek Formation was assigned a middle and upper Meramecian and Chesterian age by Poole and Sandberg (1977) who suggested that it was deposited in an inner-arc basin between the Antlers orogenic highland to the east and the Klamath-North Sierran island arc to the west. Poole and Sandberg (1977, p. 82) stated:

Our Mississippian model involves continued Benioff-type subduction of Pacific oceanic crust beneath an island-arc complex above an east-dipping subduction zone separated from the continental slope and shelf by an inner-arc basin.

SPOTTED RIDGE FORMATION

The Spotted Ridge Formation is in juxtaposition with the Coffee Creek Formation and is separated from it and from nearby Permian and Triassic formations by faults. Cenozoic erosion has produced a rolling grass- and sagebrush-covered terrain containing shallow ravines. Outcrops vary according to the type of sedimentary rocks; the conglomerate and sandstone are more in evidence than the mudstone.

The Spotted Ridge Formation was named by Merriam and Berthiaume (1943). They included in it approximately 300–500 m of beds ranging from coarse conglomerate and crossbedded sandstone to compact mudstone and noted that the formation was variable both vertically and laterally. Minor amounts of chert and magnetite sandstone are present.

The conglomerate has been examined by Taubeneck (1969), who found that most granitic clasts are quartz diorite. Some clasts are granodiorite; one is quartz monzonite. The only clast having a gneissoid texture is quartz monzonite. The source of the conglomerate is unknown.

A medium-grained sandstone was examined by Sam Boggs, Jr., of the University of Oregon. This rock contained 48 percent volcanic rock fragments, 8 percent quartz, 4 percent polycrystalline quartz, 1 percent chert, 15 percent plagioclase feldspar and no evidence of potassium feldspar, 8 percent authigenic chlorite, (?)3 percent authigenic hematite and more than 10 percent matrix consisting mainly of chlorite. He stated:

Some of the softer rock fragments have been squeezed and flattened by compaction and many of the fragments are partially to completely altered to chlorite. Most of the plagioclase grains show some degree of alteration to sericite. Using Gilbert's classification, this sandstone is a lithic (volcanic) wacke.

The inconsistent character of clastic sedimentary deposits and the intermixture of continental and marine beds indicate that the formation was deposited near the shoreline in a rapidly filling basin during a time of active tectonism.

Carbonaceous sandstone and siltstone are present throughout the formation, but the upper part of the formation includes mudstone containing an abundant flora. Merriam and Berthiaume (1943, p. 152) described the location (locality 115) as follows:

The lenticular mudstones and siltstones are of an unusual shade of medium-grayish olive green. Lamination is usually not well defined, and fissility is undeveloped. The sediment is very compact and brittle. In general, the plants lie more or less parallel to bedding though occasional leaves and stems are decidedly oblique in position. The plants occur as coaly films; stems are much flattened but retain evidence of vascular structure. * * * While certain tongues or lenses within the plant-bearing beds are marine or brackish it is believed that most of the sediments in this facies are landlaid, though an estuarine origin is not unlikely. The plants do not appear to have suffered transportation; in fact some calamite stalks with whorls of twigs appear to be essentially in position of growth.

Read and Merriam (1940) examined the flora, but a more detailed study was made by Mamay and Read (1956), who found additional material. They considered the flora to be Early Pennsylvanian, somewhat equivalent to the Pottsville of Pennsylvania.

A marine fauna is found in a ravine just north of the road that leads eastward from Coffee Creek over a low ridge and descends into Grindstone Creek. The fauna, exposed just north of the south line of sec. 30, T. 18 S., R. 25 E., is in gritty sandstone and fine conglomerate, that has clasts commonly as much as a centimeter in diameter. The writer visited the locality twice, accompanied by Gregory Miles, Jack Messe, Thomas Sharp, and William Gandera. Daniel E. Penttila contributed fossils from the same outcrops.

The fauna was identified by MacKenzie Gordon, Jr., of the U.S. Geological Survey in consultation with G. A. Cooper, National Museum of Natural History; and John Pojeta, Jr., and E. L. Yochelson, U.S. Geological Survey. The fauna is assigned by them to the Morrowan stage of the Pennsylvanian, with suggested correlations with other units. The following genera were identified:

Corals:	
Lophophyllidium sp.	1
Bryozoans:	
Fenestella sp.	1
Brachiopods:	
Schizophoria? sp. indet.	fragment
Derbyia? sp. indet.	fragment
Rugoclostus? sp. indet.	fragment
Rhynchonelloid, gen. and sp. indet.	fragment
Neospirifer sp.	3
Pelecypods:	
Nuculopsis sp.	1
Phestia sp.	14
Permophorus sp.	3
Edmondia sp.	1
Solemya sp.	1
Rostroconchs:	
Bransonia sp.	1
Gastropods:	
Sinuatina sp.	3
Euphemites sp.	2
Bellerophon (*Bellerophon*) sp.	9
Retispira sp. indet.	9
Glabrocingulum (*Ananias*) sp.	8
Pleurotomariacean?, gen. and sp. indet.	2
Shansiella? sp.	2
Girtyspira sp.	1
Meekospira? sp. indet.	2
Ammonoids:	
Syngastrioceras? sp. indet.	fragment
Cancelloceras cf. *C. cancellatum* (Bisat)	20 (mostly fragments)
Vermes?:	
Worm tube or scaphopod	1
Echinoderms:	
Crinoid columnals	

MacKenzie Gordon, Jr. (written commun., 1977), notes:

The preservation of the ammonoids permits identification of the genus *Cancelloceras*, a reticulate form of *Gastrioceras*, known in northwest Europe at the base of the Lower *Gastrioceras* (*G.*) Zone and recently recognized in the Hale Formation of Arkansas. This permits an Early Pennsylvanian age to be assigned to the Spotted Ridge Formation, from which these fossils were collected.

Whether or not beds of later than Morrowan Pennsylvanian age are present in the Spotted Ridge Formation is not presently clear. A single well-preserved specimen from float in Crook County, Oregon, was described in 1940 as *Eoasianites merriami* Miller and Furnish. Now known as *Somoholites merriami* (Miller and Furnish), it has been recognized in the Barnett Hill Member of the Atoka Formation in Oklahoma and the Hare Fiord reef deposits of Ellesmere Island, Arctic Canada, by Saunders (1971, p. 109). To this I can add the well known Smithwick Shale locality, of Atokan age, in McCulloch County, Texas. Whether this Atokan *Somoholites* from central Oregon came from some part of the Spotted Ridge Formation or whether it is from an as yet unrecognized stratigraphic unit in that vicinity is a matter yet to be determined.

ELKHORN RIDGE ARGILLITE

The Elkhorn Ridge Argillite was named by Gilluly (1937) for exposures in Elkhorn Ridge a short distance west of Baker. The formation consists of approximately 1,750 m of silicified highly contorted argillite, tuff, chert, and some limestone and greenstone. The argillite, which makes up the greater part of the formation, is fine grained, thinly laminated, and consists of quartz, andesine, muscovite, chlorite, and black carbonaceous material.

The age and position of the Elkhorn Ridge Argillite has been discussed by Vallier, Brooks, and Thayer (1977). Fusulinids within the unit have been assigned to the Leonardian and the unit considered essentially equivalent to the Coyote Butte Formation of central Oregon, which overlies the Spotted Ridge Formation. Fusulinids from the Virtue Hills east of Baker were found by Bostwick and Koch (1962) and identified as *Yabeina* and *Schwagerina*. They were large, well advanced, and probably Ochoan. Thus, the preponderant evidence points to Middle and Late Permian age, with a possibility of some Triassic beds.

However, D. A. Bostwick (written commun., 1977) reported fusulinids of Middle Pennsylvanian (Desmoinesian) age in beds assigned to the Elkhorn Ridge Argillite in the lower half of sec. 2, T. 10S., R. 43 E., in the Durkee quadrangle. Apparently the Elkhorn Ridge Argillite ranges into the Middle Pennsylvanian as well unless rafted blocks of this age resembling the Elkhorn Ridge have been brought into juxtaposition during plate movements.

Vallier, Brooks, and Thayer (1977, p. 461) discussed the age and setting for the emplacement of the Paleozoic units of northeastern Oregon. They stated:

The rocks have undergone severe deformation, which placed supracrustal and basement rocks in structural contact to form a melange-like chaotic assemblage. Tectonic shuffling may have taken place on a major scale. For example, limestone bodies containing Tethyan fusulinid faunas are relatively close to those with non-Tethyan affinities (Bostwick, personal commun., 1977) * * * Known ages of the rocks in this terrane range from Pennsylvanian to Mesozoic, indicating that the chaotic mixture formed throughout a long time period, perhaps with last major movement in the Late Triassic or Early Jurassic.

CARBONIFEROUS ROCKS IN WASHINGTON

By Ernest H. Gilmour and Wilbert R. Danner

ABSTRACT

Scattered Carboniferous rocks in both western and eastern Washington range in age from Chesterian to Desmoinesian and possibly Missourian. Some of the Carboniferous rocks in western Washington contain Tethyan faunas and are believed to have been tectonically transported north and east from the Permian equatorial area. Fossils in the Carboniferous rocks of northeastern Washington are thought to have eastern North American or Rocky Mountain affinities.

INTRODUCTION

Carboniferous rocks are found in three areas of Washington: the San Juan Islands in northwest coastal Washington; the Northern Cascade Mountains; and scattered localities in northeast Washington (fig. 21). In western Washington, most fossils are found in limestone lenses that are interbedded

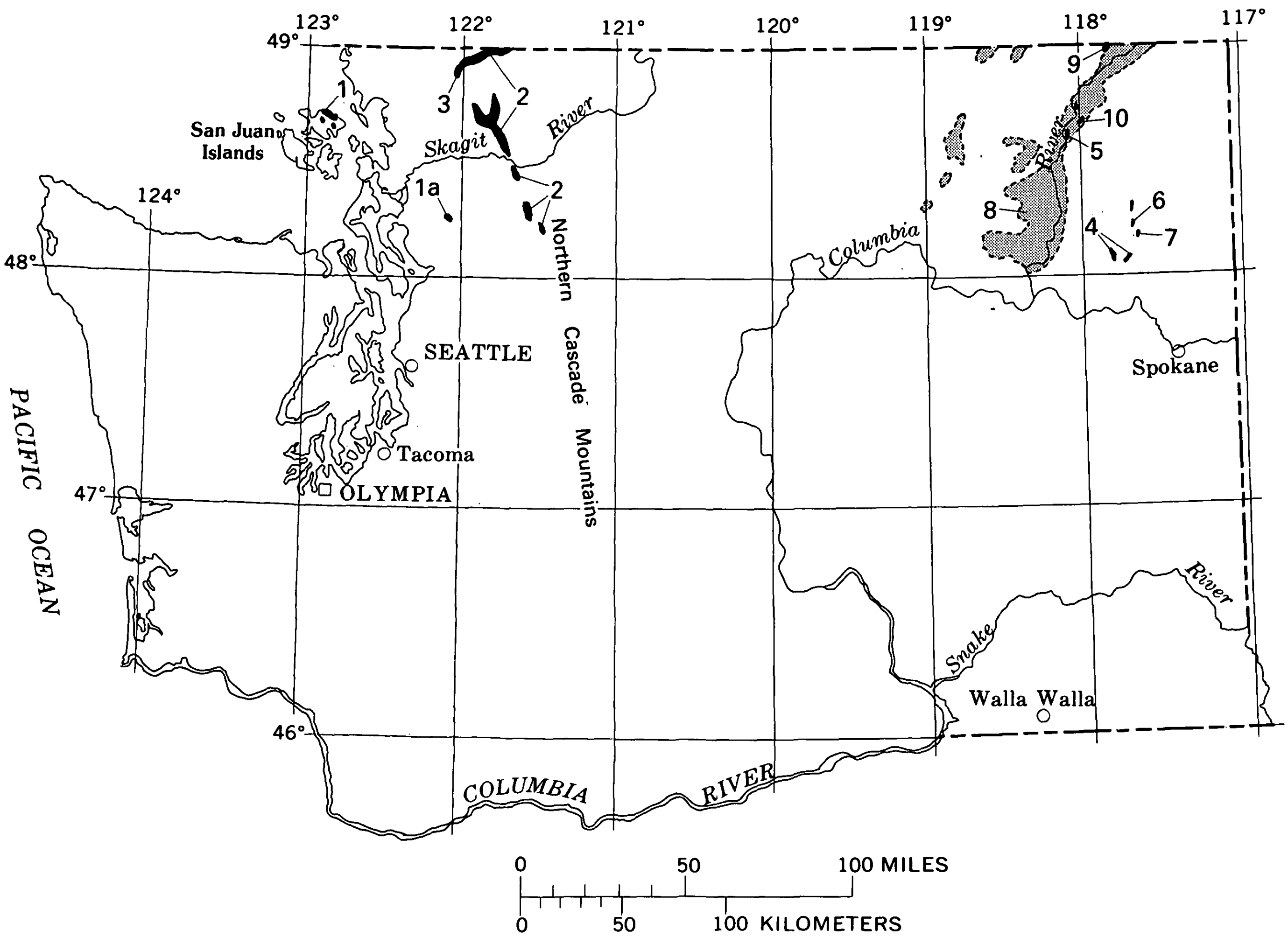

FIGURE 21.—Carboniferous rocks and rocks of questionable Carboniferous age exposed in Washington. Locality numbers are referred to in text. Solid areas are known Carboniferous rocks and shaded areas are upper Paleozoic undifferentiated.

with thick detrital sedimentary sequences and associated volcanic rocks. In eastern Washington, both Mississippian and Pennsylvanian fossiliferous limestone bodies have been described as biohermal buildups deposited in relatively shallow water in the eastern part of the "Cordilleran eugeosyncline." The western Washington occurrences are well exposed in beach cliffs on the San Juan Islands and in the alpine regions (fig. 22) of the Northern Cascade Mountains but are otherwise mantled by a thick sequence of Pleistocene and recent sediments and by dense vegetation. They have been folded and faulted and subjected to low-temperature—high-pressure metamorphism of the prehnite-pumpelleyite facies. Volcanism, plutonism, folding, and block faulting (Yates, 1970) have disturbed and metamorphosed much of the Carboniferous strata in eastern Washington. Also in eastern Washington, glaciofluvial, lacustrine, and alluvial deposits cover much of the bedrock.

SAN JUAN ISLANDS

Rocks of Early and Middle Pennsylvanian age in the San Juan Islands make up the middle part of the Trafton Group. On Orcas Island (fig. 21, loc. 1), they are estimated to be 150 to 300 m thick (Danner, 1966). Small limestone pods and lenses containing Pennsylvanian fossils are interbedded with thin-bedded shale, siltstone, graywacke, and volcanic flows and breccia. At least one thin seam of poor quality coal is found on the northeast coast of Orcas Island.

The age of the Trafton Group ranges from Devonian to at least Late Permian, but until recently, no Mississippian strata have been conclusively identified, although brachiopods assigned to that age are found on Orcas Knob on Orcas Island associated with other faunas thought to be of Early Pennsylvanian age. *Eostaffella, Nankinella, Millerella?, Tetrataxis,* endothyrid Foraminifera, *Komia,* corals,

FIGURE 22.—Lower Pennsylvanian Limestone of Red Mountain Subgroup exposed at the base of Washington Monument Peak, Skagit County, Washington, Northern Cascade Mountains. View looking northwest. Limestone is underlain by argillaceous rocks and overlain by coarse clastic rocks of Pennsylvanian or Early Permian age.

bryozoans, brachiopods, and one species of trilobite have been collected from the Carboniferous strata (Danner, 1966; Sada and Danner, 1973).

The Trafton Group rocks are considered to be an exotic belt of Paleozoic rocks differing lithologically from other Paleozoic rocks in western Washington and containing a Tethyan fauna (Danner, 1977). This western Cordilleran Tethyan belt must have been tectonically transported north and east from the Permian equatorial area to its present location.

The Trafton Group extends southeastward into the foothills and mountains of the southern part of the Northern Cascades, but until recently, only Permian faunas had been found in this area. One outcrop of radiolarian chert of this group (fig. 21, loc. 1a), previously assigned a Permian age, is now reported to contain conodonts and radiolarians of Mississippian age (J. Whetten and D. Jones, oral commun., 1977).

NORTHERN CASCADE MOUNTAINS

The Chilliwack Group (Daly, 1912; Moen, 1962; Smith, 1962; Monger, 1966; Danner, 1957, 1966, 1970, 1977) contains rocks of Pennsylvanian age exposed in a northwest-southeast band (fig. 21, loc. 2) extending from near Harrison Lake in southwestern British Columbia south across the border into Washington State through Whatcom, Skagit, and Snohomish Counties in the Northern Cascade Mountains. The Chilliwack Group is composed of thin-bedded argillite, siltstone, and graywacke together with interbedded massive to well-bedded argillaceous limestone, conglomerate, minor radiolarian chert, and andesitic and dacitic volcanic rocks. The Pennsylvanian part of the group (Red Mountain Subgroup) is composed largely of thin-bedded argillite and cherty argillite, siltstone, graywacke, and argillaceous limestone. The limestone is near the top of the subgroup and is overlain by sandstone and conglomerate of either Pennsylvanian or Permian age which contain plant debris. *Lepidodendron* and *Calamites* (identified by G. E. Rouse) occur in this sandstone on Red Mountain.

Thickness of the Pennsylvanian rocks is estimated to be 300 to 600 m. Lenticular beds of limestone at least 120 m thick crop out on Red Mountain at the type area of the subgroup (fig. 21, loc. 3). Much of the limestone is organoclastic and oolitic.

The age of the Chilliwack Group ranges from at least Devonian to middle Permian. Some of the limestone contains brachiopods similar to *Gigantoproductus* and may be of Mississippian age, but the same limestones contain a microfauna identified as being of Early Pennsylvanian age and corals thought to be of Permian age. Other limestones containing abundant *Endothyra* may be of Mississippian age. Large crinoid columnals (fig. 23) 40 mm or more in diameter are characteristic of many of the Pennsylvanian limestones and are considered to be good index fossils for field identification of these Lower Pennsylvanian rocks.

The faunas of the Chilliwack Group are non-Tethyan but of a distinctive non-American aspect; the Chilliwack Group is believed to have formed as an island arc in a subtropical climate in the Paleozoic Pacific Ocean and to have been rafted in against North America during mid-Mesozoic time. The Red Mountain Subgroup correlates with the Harper Ranch Group near Kamloops to the north in British Columbia and with the upper part of the Baird Formation in northern California.

NORTHEASTERN WASHINGTON

Rocks of known Carboniferous age and rocks of questionable Carboniferous age are found in northeastern Washington. Some rocks mapped as Permian may include Carboniferous strata. Outcrops are usually limited in extent, tectonically disturbed and metamorphosed (Yates, 1970; Mills, 1977); consequently, it is difficult to correlate stratigraphic units any distance with much confidence.

Documented Carboniferous rocks have been reported by various workers in several places. Enbysk (1956) reported both Mississippian and Pennsylvanian fossils near Kulzer (fig. 21, loc. 4) in southeastern Stevens County. The Upper Mississippian (Chester) rocks include the ostracodes *Graphiodactylus tenuis, Jonesina carterigera,* and *Cavellina* aff. *C. coryelli;* coral *Triplophyllum* sp.; brachiopods *Spirifer* cf. *S. pellaensis* and *Chonetes* sp. Pennsylvanian fossils were identified (Enbysk, 1956) from a limestone near Springdale in southeastern Stevens County. These fossils include *Spirifer* aff. *S. rockymontanus, Rhombopora nitidula,* and *Lophophyllidium* cf. *L. proliferum.* Six genera of Foraminifera were also identified, including *Rhabdammina, Ammobaculites, Endothyra, Millerella, Globivalvulina,* and *Trochammina.* Enbysk reports these species as having eastern North American or Rocky Mountain affinities. Recently, Bruce Wardlaw (oral commun., 1977) has reported Pennsylvanian (Desmoinesian or younger) fossils from the Kettle Falls quadrangle (fig. 21, loc. 1) in Stevens County. The fauna consists of brachiopods, pelecypods, trilobites, and dendroid and fenestrate bryozoans. Some of the fossils described are *Cyrtorostra* sp., *Costachonetes* sp.,

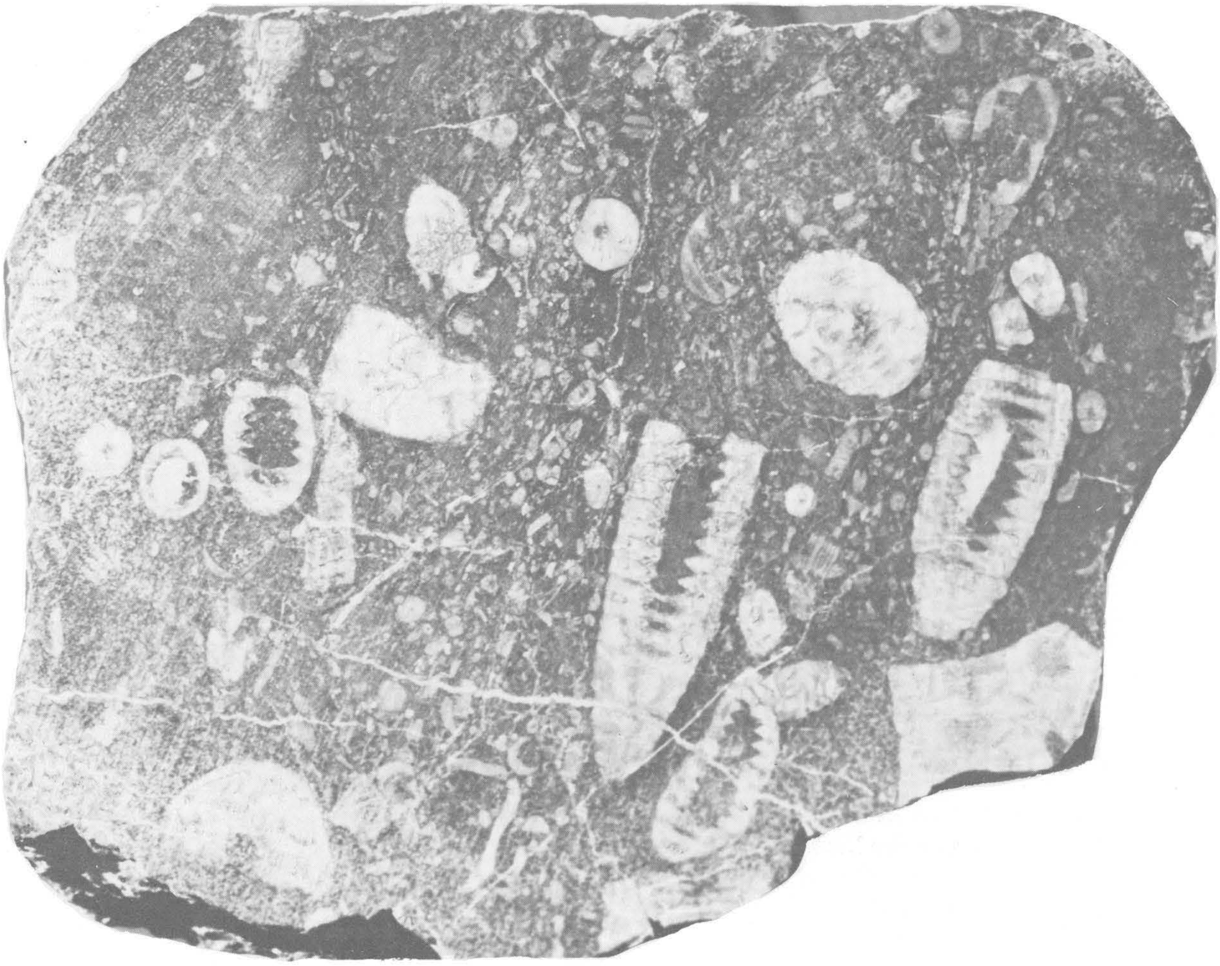

FIGURE 23.—Cut slab of crinoidal limestone of Early Pennsylvanian age, Red Mountain, Whatcom County, Washington.

Neospirifer cf. *N. triplactus?* (Hall), *Rhynchopora* sp., *Waagenoconcha* sp., and *Kutorginella* sp. Miller (1969) reported at least 180–210 m of limestone containing abundant chert north of Valley (fig. 21, loc. 6) that contains Mississippian fossils. McLaughlin and Simons (1951) described Late Mississippian (Chester) ostracodes from a limestone 2.4 km east of Valley (fig. 21, loc. 7), and other workers have identified Mississippian brachiopods, corals, and conodonts.

An extensive area of sedimentary rocks including strata of Carboniferous (?) age has been referred to as upper Paleozoic undifferentiated or upper Paleozoic and Triassic (Mills, 1962) in northeastern Washington (fig. 21, loc. 8). This area extends southward from the Canadian border and parallels the Columbia River to the confluence of the Columbia and Spokane Rivers. Some of these rocks are probably Pennsylvanian in age.

Yates (oral commun., cited by Irwin, 1975) believes that the limestone lenses and associated graywacke and argillite in northern Stevens County (fig. 21, loc. 9) are a southward continuation of fossiliferous rocks of the Mt. Roberts Formation of southern British Columbia. Little (1960) listed several fossils of "probable Pennsylvanian age" in the Mt. Roberts Formation of southern British Columbia. The Mt. Roberts Formation consists of fine clastic rocks intercalated with pyroclastic debris and mafic flow rocks and fossiliferous limestone lenses. Yates (1970) considered these rocks to have been deposited in the Cordilleran eugeosyncline.

Some of the rocks previously mapped as Permian in northeastern Washington have recently been identified as Pennsylvanian (?). West (1976) reported *Pseudoendothyra* sp. from a biohermal bank 5 km northeast of Kettle Falls (fig. 21, loc. 10). This limestone is in close proximity to similar limestones containing Permian fusulinids. As additional detailed studies are conducted, more of the area identified as upper Paleozoic undifferentiated will probably be found to be of Carboniferous age.

ECONOMIC PRODUCTS

Limestone has been the principal economic product from Carboniferous rocks in Washington. Before 1900, Carboniferous limestone was quarried and used for building stone and the production of burned lime. Since 1900, large quantities of limestone have been used for portland cement, in the paper pulp industry, and for agricultural purposes. Two excellent reports on the limestone resources of Washington have been published by Mills (1962) and Danner (1966).

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The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Alaska

By J. THOMAS DUTRO, JR.

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Historical review and summary of
areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in Alaska

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CONTENTS

ILLUSTRATIONS

THE MISSISSIPPIAN AND PENNSYLVANIAN (CARBONIFEROUS) SYSTEMS IN THE UNITED STATES—ALASKA

By J. Thomas Dutro, Jr.

ABSTRACT

The Carboniferous rocks of Alaska are more widely distributed than those of any other Paleozoic system. The Lisburne Group of northern Alaska contains a great variety of carbonate facies, predominantly shallow-water deposits, but including some outer shelf and slope deposits. In east-central Alaska, the deep-water Ford Lake Shale is succeeded by the Calico Bluff Formation, which was probably deposited in a slope environment. Scattered throughout central and southwestern Alaska are remnants of deep-water laminated chert and argillite deposits containing a few carbonate members that may be of shelf-margin or upper slope origin. In southeastern Alaska, Carboniferous deposition began with deep-water chert and volcaniclastic rocks, but most of the sequence reflects upward-shoaling shallow-water carbonate deposition, perhaps in an island archipelago.

Correlations among the various sequences in Alaska and adjacent areas are based mainly on the foraminiferal zonation of B. L. Mamet and on other fossils, among which the cephalopods are most significant. Corals, brachiopods, echinoderms, and mollusks are regionally important guides in certain parts of the section, and fusulinids are quite useful in the upper Carboniferous deposits. Fossil plants, although minor elements of the total biota, are important in the Lower Mississippian clastic sequences and have provided a spectacular breakthrough in the recognition of terrestrial Upper Pennsylvanian strata in the western Alaska Range.

The most important economic factor in the Carboniferous deposits of the State is the oil and gas potential of the Lisburne Group in the subsurface of northern Alaska. The abundance of carbonate rocks for potential industrial uses is an untapped resource for possible future development.

INTRODUCTION

Rocks of Carboniferous age are widely distributed in Alaska (fig. 1), representing more outcrop area than those of any other Paleozoic system. Carbonate rocks of the Lisburne Group compose most of the thrust sheets and folded structures in the central and northern Brooks Range. This sequence in northern Alaska was deposited by the sea onlapping older terranes, the first rocks in the sequence are shore and nearshore clastic rocks. The succeeding Mississippian and Lower Pennsylvanian carbonate strata were deposited in a complex of predominantly shallow-water marine environments. Similar carbonate sections are also present at the western tip of the Seward Peninsula and on St. Lawrence Island in the Bering Sea. The Carboniferous of east-central Alaska begins with deep-water dark chert and shale of the Ford Lake Shale; the Ford Lake is overlain by slope deposits of limestone and shale, the Calico Bluff Formation, part of which is as young as Early Pennsylvanian. Although the Carboniferous is sparsely represented in central, southern, and southwestern Alaska, deep-water clastic and volcaniclastic rocks are present in the Alaska Range, and bedded chert and argillite are present in the Medfra quadrangle. Recently discovered terrestrial plant-bearing conglomerates of Pennsylvanian age crop out on Mt. Dall in the Talkeetna quadrangle, southwest of Mt. McKinley. This is the first positive evidence of nonmarine late Carboniferous deposits in Alaska. In southeastern Alaska, an essentially carbonate sequence of Mississippian and Lower Pennsylvanian rocks is represented by the Iyoukeen Formation at Freshwater Bay and by the Peratrovich and Klawak Formations and the Ladrones Limestone in the Prince of Wales Island area. In addition, the Saginaw Bay Formation, which contains a lower volcanic member, a black chert member, a chert and limestone member, and an upper silty limestone member, is present in the northern part of Kuiu Island and on the Keku Islets. The Saginaw Bay Formation appears to include correlatives of the Peratrovich, Ladrones, and Klawak.

Most of the exposures in northern Alaska are north of the tree line and, consequently, are magnificent. Except where covered by deposits of mountain glaciers and recent talus cones and colluvium, Carboniferous strata are totally exposed, and opportunities for unravelling the stratigraphy, structure, and geologic history of the region are unexcelled. Much of the lowland region of central and southern Alaska is tree covered, and exposures are limited to

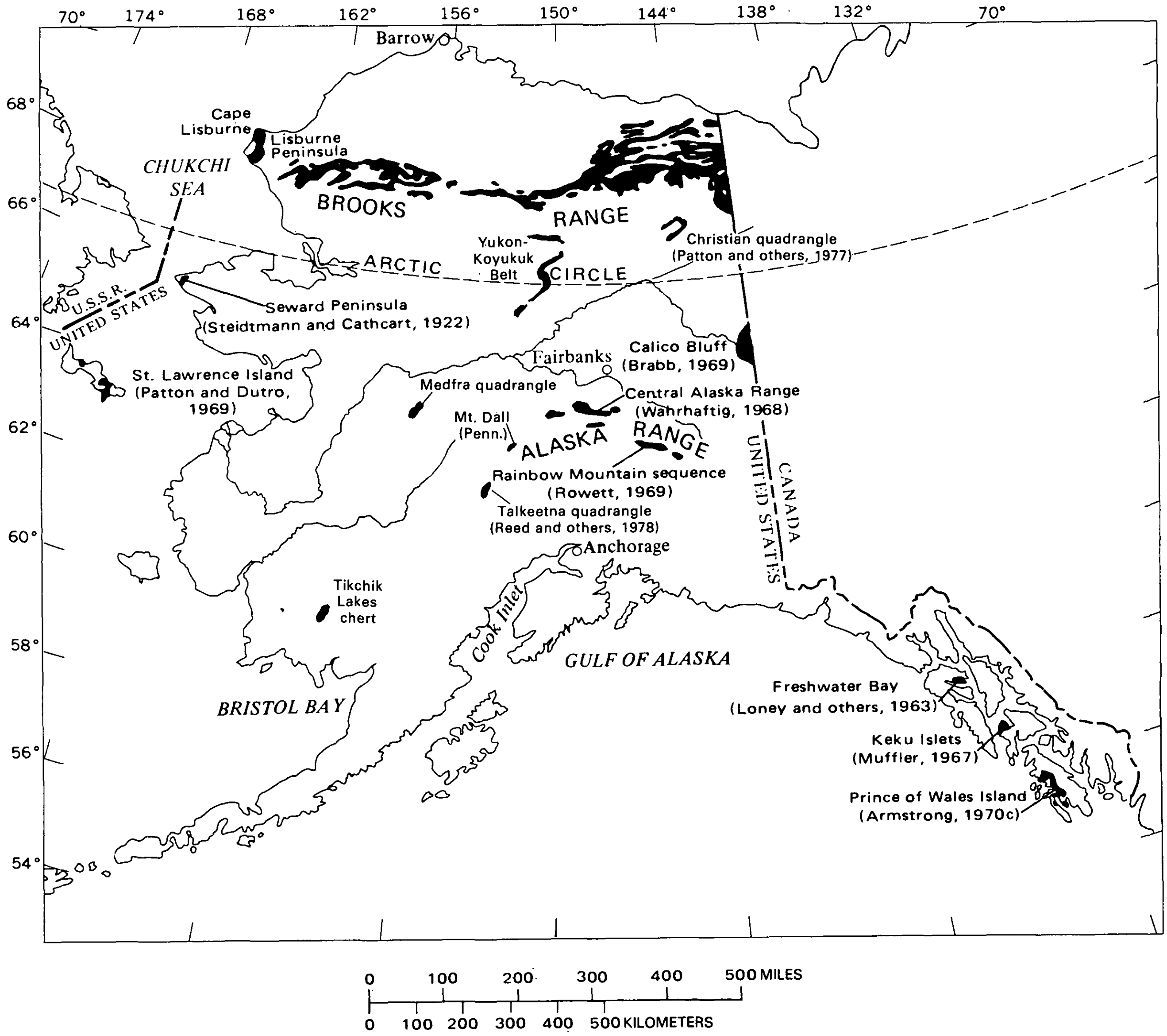

FIGURE 1.—Map showing distribution of Carboniferous rocks (shaded areas) in Alaska. Modified from Armstrong (1975).

river bluffs and isolated hilltop outcrops. Some Carboniferous strata crop out above the timberline in the Alaska Range, and exposures in southeastern Alaska are superb along the island coastlines where the great tidal range provides hectares of fine outcrop from low tide to midtide.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the State of Alaska Department of Natural Resources, Division of Geological and Geophysical Surveys.

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In addition, the work of K. J. Bird and C. F. Jordan has been relied on, in large part, for the analysis of the oil and gas potential of the Lisburne Group in northern Alaska.

Finally, I am grateful to a group of supervisors who had the vision and faith in our work to support projects, at times rather esoteric, which permitted all of us to explore together the fascinating Carboniferous geology of Alaska. These farsighted chiefs include George Gryc, Preston E. Cloud, Jr., Wilmot H. Bradley, George L. Gates, Ralph L. Miller, Charles L. Anderson, and Thomas G. Ovenshine.

HISTORY

The first report of Carboniferous rocks from Alaska is probably that of Capt. F. W. Beechey (1831) who, along with other early coastal explorers, was awed by the spectacular cliff exposures near Cape Lisburne. The first mention of Carboniferous fossils is by W. Buckland (1839, p. 171) who reported that fossils collected from Cape Lisburne by the Beechey party are indistinguishable from those of the Derbyshire Limestone and listed "producta Martini, other productae, lithostrotion, flustrae and trilobites." In the latter half of the 19th century, several other writers referred to the occurrence of Carboniferous fossils in Alaska, most of them relying on Buckland's earlier statement. By the end of the century, scattered information on Paleozoic fossils, including those of the Carboniferous, was summarized by Charles Schuchert, W. H. Dall, and F. H. Knowlton (*in* Dall, 1896, p. 864–865, 876–906) in the 17th Annual Report of the U.S. Geological Survey.

During the first decade of this century, geologic exploration related to the search for gold and other mineral deposits added substantial knowledge concerning the distribution of Carboniferous rocks in Alaska. The Lisburne formation was named by F. C. Schrader (1902), who provisionally assigned it to the Devonian. By 1906, however, G. H. Girty and David White (*in* Collier, 1906) had correctly assigned the Lisburne fossils to the early Carboniferous.

Carboniferous rocks were found during the geological reconnaissance conducted in conjunction with the work of the International Boundary Survey in 1911 and 1912. The part of the boundary between the Yukon and Porcupine Rivers was studied by the Geological Survey of Canada (Cairnes, 1914). The part from the Porcupine River northward to the Arctic Coast was examined by the U.S. Geological Survey (Maddren, 1912). Cairnes recognized three major rock units—a shale group, a limestone-chert group, and the Nation River Formation. Subsequent work has shown that the Nation River is Upper Devonian and that much of the shale group is lower Paleozoic, but the limestone-chert group is probably what is now recognized as the Calico Bluff Formation. G. H. Girty of the U.S. Geological Survey identified many fossil collections and showed that these rocks are, indeed, of Carboniferous age, although many subsequent refinements in the ages and correlations have been made. North of the Porcupine River, Maddren (1912) found large areas of Carboniferous limestone when he crossed the Brooks Range. He reported that these rocks stretch for nearly 64 km north of Ammerman Mountain and that they contain Carboniferous fossils. Unfortunately, Maddren's maps and complete report were never published, but much of the information has been incorporated into later U.S. Geological Survey maps and papers.

A flurry of activity in the 1920's related to the exploration of Naval Petroleum Reserve (NPR) No. 4 greatly increased the understanding of the geologic history of northern Alaska. Specific details of Carboniferous geology and an excellent summary of all earlier exploration were given by Smith and Mertie (1930). During the same decade, detailed exploration in southeastern Alaska resulted in the first regional geologic synthesis of the panhandle by Buddington and Chapin (1929). The Yukon-Tanana region, a third major area of Carboniferous rocks, was examined primarily by J. B. Mertie, Jr., in the 1920's and early 1930's. Mertie's conclusions regarding the stratigraphy, structure, and general geology of this vast region were presented in two major papers (1930 and 1937). The general state of geologic knowledge of Alaska before World War II was summarized by P. S. Smith (1939), who discussed the Carboniferous deposits in each region and presented a very generalized correlation chart.

Until World War II, practically all the paleontologic and biostratigraphic information on the Carboniferous of Alaska was the result of the work of G. H. Girty, who provided the extensive faunal lists, age designations, and correlations that appeared in the reports of the U.S. Geological Survey during the first four decades of the century.

Work in NPR No. 4 was renewed during the latter years of World War II. From 1944 to 1953, more than 300 man-months of geologic fieldwork greatly increased the knowledge of the geology of Carboniferous deposits in about 54,390 km^2 in northern Alaska and culminated in a series of regional papers and maps (Reed, 1958).

Modern stratigraphic and paleontologic studies began with the work related to the exploration of NPR No. 4. New formational units were named, and a preliminary biostratigraphic zonation was established (Bowsher and Dutro, 1957). Several groups of fossils were described in some detail (Gordon, 1957; Yochelson and Dutro, 1960; Sohn, 1971).

In the early 1960's, major oil companies instituted massive regional studies and examined the possibility that the widespread Carboniferous carbonate rocks in northern Alaska would become oil and gas producers. Detailed regional, stratigraphic, facies, petrographic, and depositional analyses were made. A. K. Armstrong, now with the U.S. Geological Survey, contributed significant papers on coral biostratigraphy, carbonate petrography, and environmental reconstructions of the Lisburne rocks (1970a, b, c; 1972a, b; 1975).

Major advances in understanding the Carboniferous of southeastern Alaska resulted from detailed mapping and stratigraphic work of U.S. Geological Survey parties during the past 20 years. The Peratrovich and Klawak Formations and the Ladrones Limestone on Prince of Wales Island were described by Eberlein and Churkin (1970); and the Iyoukeen Formation of Freshwater Bay was named by Loney, Condon and Dutro (1963). Fusulinids from the Ladrones and Klawak indicate that they are early Middle Pennsylvanian (Douglass, 1971), Mississippian corals and biostratigraphy of the Peratrovich were described by Armstrong (1970c). Muffler (1967) described the Saginaw Bay Formation in the Keku Islets and nearby northern Kuiu Island; the Saginaw Bay Formation contains fossils of Mississippian and Pennsylvanian ages.

A recent summary of the Carboniferous of most of Alaska, except the southeastern panhandle, was prepared by Brosgé and Dutro (1973). Biostratigraphic studies emphasizing the calcareous microfossils have been an important part of all recent work in the Carboniferous. Microfaunal zonations established by B. L. Mamet are used for general correlations in Alaska as well as in most other parts of North America (fig. 2). Papers documenting this zonation include those by Armstrong, Mamet, and Dutro (1970, 1971), Mamet and Armstrong (1972), and Armstrong and Mamet (1975, 1977).

GEOLOGIC SETTING

Although in several areas no break in deposition is apparent between Lower Mississippian rocks and uppermost Devonian strata, at most places the Carboniferous rocks lie unconformably on older rocks (Churkin, 1973). In parts of the western and central Brooks Range, no structural break separates rocks as old as Mamet's Foraminifera Zone 6 from the underlying marine Devonian beds of Famennian age. In the central Brooks Range, where the highest Devonian unit is nonmarine, a basal Mississippian littoral sandstone lies disconformably on the upper Kanayut clastic sequence. In the eastern Brooks Range, Mississippian strata are regionally unconformable on rocks as old as Precambrian. Broad anticlinoria characterize the eastern Brooks Range and extend as far west as the Doonerak anticlinorium in the northeastern Wiseman quadrangle, where the Carboniferous rests with angular unconformity on rocks as old as Ordovician (Grybeck and others, 1977). In the extreme northeastern part of the State, Mississippian beds were deposited across a varied older terrane that ranges in age from Precambrian through early Middle Devonian (Reiser, 1970).

In east-central Alaska, where all the Upper Devonian and Carboniferous strata were deposited in relatively deep water, no depositional break is apparent at the boundary. In southeastern Alaska, the Mississippian marine shale and carbonate rocks are disconformable on Upper Devonian marine volcanic sequences, but the precise amount of time that is not represented by rocks is unknown.

The geology of pre-Carboniferous rocks is best shown in northern Alaska where Precambrian through lower Middle Devonian strata show the effects of at least three orogenic episodes. The older Precambrian is a sequence of quartz-mica schist and related metamorphic rocks that is succeeded by weakly metamorphosed marine deposits of late Precambrian, Cambrian, and Ordovician age. Middle Devonian clastic rocks lie with angular unconformity on the Ordovician in the southwestern part of the Demarcation Point quadrangle. The onlapping Carboniferous deposits cover all older structures. Granitic intrusive bodies in the Mt. Michelson–Jago River area are of two ages, 430 m.y. (million years) and 360 m.y., and volcanic rocks of similar ages have been dated from the core of the Doonerak anticlinorium (430 m.y. and 370 m.y.–380 m.y.). Equivalents of both the Taconic and Acadian mountain-building episodes of eastern North America clearly are a part of the early history of northern Alaska.

In most regions, Carboniferous rocks are overlain unconformably by Permian or younger sequences. An exception is the east-central Alaska Range where volcaniclastic rocks of the Pennsylvanian Slana Spur Formation grade upward into limestone and sandy volcaniclastic limestone of the Wolfcampian Eagle Creek Formation.

Carboniferous and early Mesozoic strata were subsequently folded and faulted beginning in Late Jurassic time. In northern Alaska, deformation is most intense in the core of the Brooks Range where long-distance thrust faulting and imbricate thrust patterns are ubiquitous. Further north, as seen in the northeastern part of the range, tectonic style tends to more open folds and high-angle reverse faulting. Broadly folded Neogene strata at the Arctic Coast near Barter Island indicate that tectonism continued until relatively recent times.

LITHOSTRATIGRAPHY

Most of the major stratigraphic units in the Alaskan Carboniferous rocks have been examined in at least a broad reconnaissance way, and most units examined have been given formal names.

In northern Alaska, the Lisburne Group is composed of predominantly carbonate rocks of Mississippian and Early Pennsylvanian age. Various formation and informal member names are applied to major mapped units, most of which reflect regional facies variations (Bowsher and Dutro, 1957; Sable and Dutro, 1961). Figure 2 shows the correlation of the Lisburne Group in sections whose locations are shown on figure 3.

Lower Mississippian clastic rocks, including the Kayak Shale, Kekiktuk Conglomerate, and Itkilyariak Formation (Mull and Mangus, 1972), have been mapped on small-scale regional maps, together with Devonian clastic units, as the Endicott Group. This is a convenient expedient because of the onlapping nature of the deposition from the Late Devonian into the Middle Mississippian.

In east-central Alaska, the Upper Devonian and Lower Mississippian Ford Lake Shale apparently grades upward into the Calico Bluff Formation of Late Mississippian and Early Pennsylvanian age.

Throughout a broad belt in central and southwestern Alaska, fragments of a deep-water argillite-chert-volcanic rock terrane have recently yielded radiolarians and conodonts of Carboniferous age. In the past, some of these outcrops have been mapped as Circle Volcanics, Rampart Group, or Livengood Chert. Although the type Livengood Chert has subsequently been shown to be of Ordovician and, perhaps, Silurian age, some exposures in the Medfra, Ruby, and Kantishna quadrangles have yielded Carboniferous fossils (Patton and others, 1977).

In the western Alaska Range (Talkeetna quadrangle), pods of limestone in deep-water graywacke-argillite sequences have yielded both Chesterian and Atokan Foraminifera (Reed and others, 1978). In the east-central Alaska Range, the Rainbow Mountain sequence is known to be of Middle or Late Pennsylvanian age (Rowett, 1969). Further to the east, Atokan and younger volcaniclastic strata containing thin limestone beds are included in the Slana Spur Formation. This formation is underlain by the Tetelna Volcanics, a thick pile of marine flows and associated volcanic rocks that may be of Early Pennsylvanian or Mississippian age (Richter and Dutro, 1975; Richter, 1976).

Although Carboniferous rocks are found in only a few places in southeastern Alaska, they are very fossiliferous and are predominantly shallow-water carbonates. The Peratrovich Formation, which is more than 300 m thick, includes relatively deep water slope deposits of shale, chert, and volcanic rocks at its base and grades upward through shelf limestones to shoal-water carbonates near its top (Armstrong, 1970c). The Peratrovich is disconformably overlain by two formations, both probably of Early or early Middle Pennsylvanian age. The Klawak Formation is composed of 150–300 m of calcareous sandstone and siltstone and minor beds of limestone and chert-pebble conglomerate. The Ladrones Limestone, on the other hand, is nearly pure massive-bedded lime mudstone containing some dolomitic beds and is as much as 300 m thick. No great time gap existed between deposition of the Peratrovich and deposition of the overlying Pennsylvanian strata as all Mamet's Foraminifera Zones

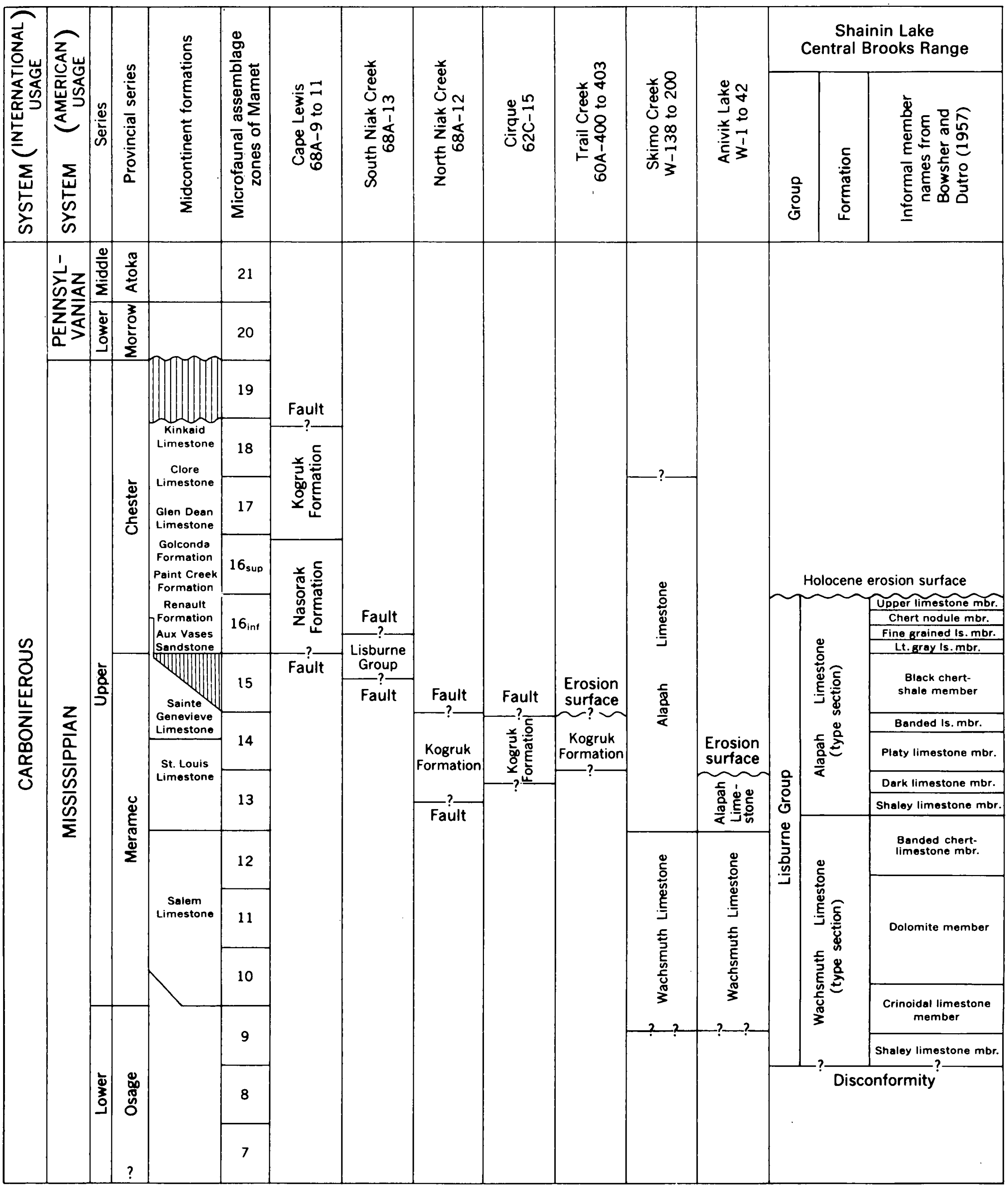
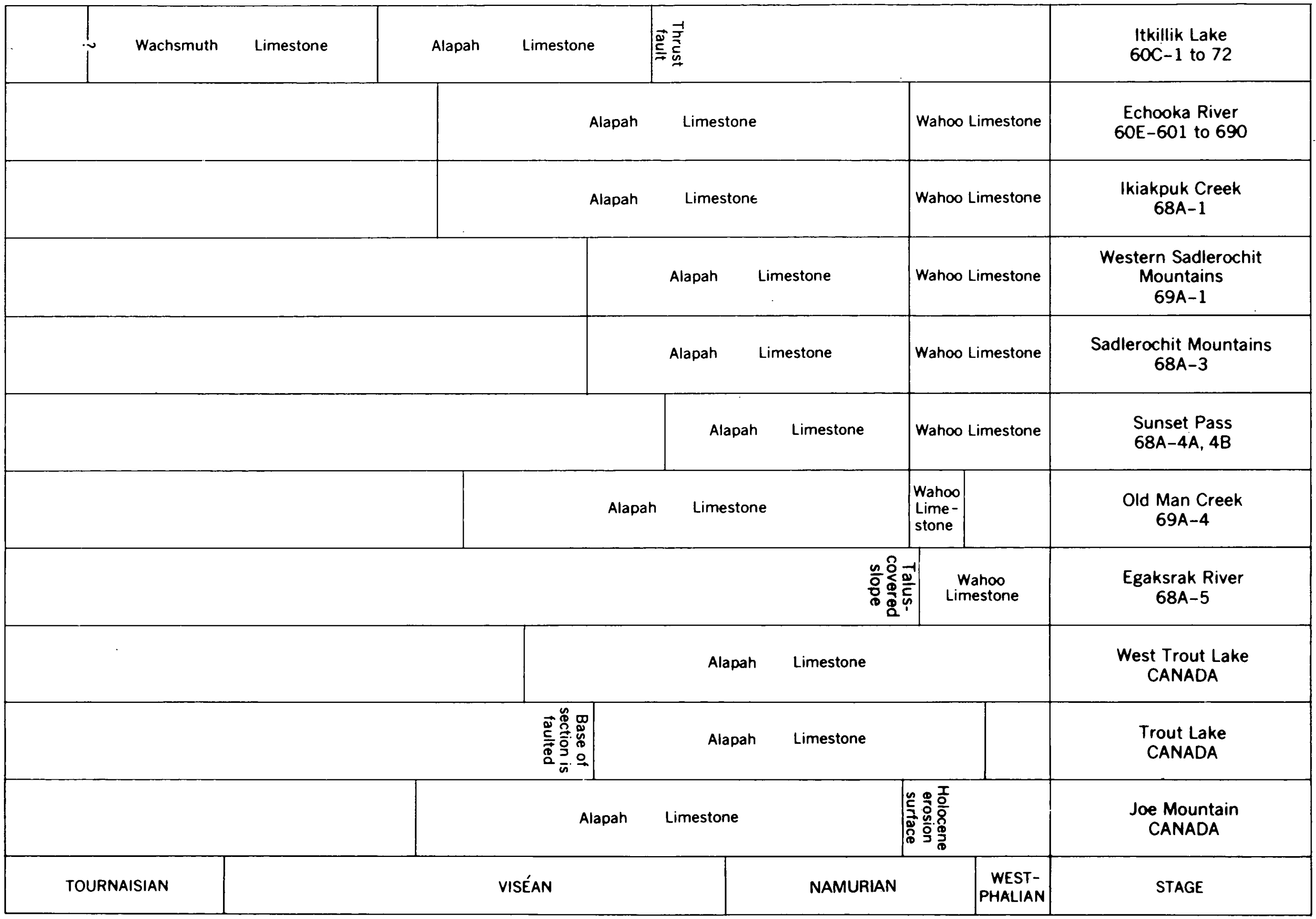


FIGURE 2.—Regional stratigraphic correlation chart of the Lisburne Group, northern Alaska and adjacent Canada.



Locations of sections are shown on figure 3. Modified from Armstrong and Mamet (1977, fig. 3).

are present, including Zone 19, which is the highest Mississippian zone in the Mamet classification.

ENVIRONMENTS OF DEPOSITION

The Carboniferous of northern Alaska was a time of northward and eastward onlap across an older complex geological terrane. Initial deposits are now littoral sandstone and shallow-water shale and shaly carbonate rock. Although many fluctuations are represented in the thick carbonate sequence, the general picture is one of regional shallow-water deposition throughout the Mississippian. In Meramecian time, platform carbonates graded southward into deeper water dark limestone and chert in some areas. In the northeast, nearly continuous deposition across the Mississippian-Pennsylvanian boundary extended the regionally shallowing pattern into the early Middle Pennsylvanian rocks of Atokan age (Wood and Armstrong, 1975).

Throughout most of the rest of the State, the Mississippian deposits were laid down on the outer shelf or upper slope. In a few places, limestones probably of shelf-edge origin contain shallow-water fossils that can be correlated with shallow-shelf deposits elsewhere.

Several of the deeper water Carboniferous deposits have been interpreted recently as parts of an extensive ophiolite terrane in the central and southwestern parts of the State (Patton and others, 1977). Although the mafic belts are thought to have been emplaced in the Jurassic, associated laminated cherts and volcaniclastic rocks have yielded Mississippian, Pennsylvanian, and Permian fossils in at least five places.

The upward-shoaling sequence in southeastern Alaska, because of its regional setting and local distribution, may well represent carbonate deposition in a volcanic archipelago (Churkin and Eberlein, 1977).

BIOSTRATIGRAPHY

All correlations of the Carboniferous of Alaska are integrated by using the scheme of foraminiferal zones proposed by B. L. Mamet (Mamet and Skipp, 1970; Sando and others, 1969). In this numbered sequence of zones, 6 through 9 are Tournaisian, 10 through 16 are Viséan, 17 through 20 are Namurian, and 21 is early Westphalian. The Mississippian-Pennsylvanian boundary is about at the boundary between Zones 19 and 20. In terms of midcontinent stratigraphy, the Chesterian-Meramecian boundary lies at about the boundary between Zones 16_{inf} and 15, and the Meramecian-Osagean boundary is approximately at the boundary between Zones 10 and 9. Beds of Zone 6 and pre-7 ages are equivalent to the Kinderhookian, but the Kinderhookian-Osagean boundary has not been precisely set.

The first generalized sequence of megafossil zones was discussed by Bowsher and Dutro (1957, p. 5–6) and was later slightly modified by Yochelson and Dutro (1960, p. 114–115). These assemblage zones are most useful for identifying informal subdivisions of the Wachsmuth and Alapah Limestones in the central Brooks Range.

Cephalopods from the Carboniferous of Alaska were described by Gordon (1957), who placed the several occurrences of nautiloids and goniatites in the general zonal scheme that he had devised for North America. Correlations with the worldwide zones were also shown.

Gastropods were described by Yochelson and Dutro (1960), and their relationship to the megafaunal zonation was indicated. The systematic descriptions of corals that have been published during the last decade by Armstrong (1970c; 1972a, b) were related by that author (1975) to the biostratigraphic scheme for the Lisburne Group, and a coral zonation was established. New ostracodes were described by Sohn (1971), and, although they appear to be of limited biostratigraphic use, they provide valuable paleontologic documentation for the Mississippian fossil assemblages in Alaska.

Few Pennsylvanian megafossils have been described, but calcareous Foraminifera were described by Armstrong and Mamet (1977), and Pennsylvanian fusulinids of southeastern Alaska were described by Douglass (1971).

Mississippian biostratigraphic zonations and occurrences of the several invertebrate groups that have been studied in more detail are shown on figure 4. As a result of discussion with Mackenzie Gordon, Jr., and A. K. Armstrong (oral communs., 1978), positions of some of the cephalopod and coral zones are adjusted slightly from those originally published.

Fossil plants are not abundant in the Carboniferous of Alaska, and no coal beds of any significance are known. Nevertheless, plant fragments are present in the dark shales near the base of the Mississippian sequence throughout the Brooks Range. Many of these collections have been studied by S. H. Mamay during the past two decades. A combined list of this Mississippian floral assemblage includes *Lepidodendropsis* sp., *Calamites* sp., *Stigmaria* cf. *S. ficoides* Sternberg, *Lepidodendron* cf. *L. veltheimii*

Sternberg, *Lepidophyllum* sp., *Lepidostrobus* sp., *Rhodea vespertina* Read, and *Triphyllopteris* sp. According to Mamay (written commun., 1970), this assemblage represents Floral Zone 2 of Read and Mamay (1964).

Plants were collected years ago from carbonaceous shale in the exposures along the sea coast south of Cape Lisburne (David White, *in* Collier, 1906, p. 22). In a review of these floras, Mamay (written commun., 1964) identified several forms among new collections that permitted him to recognize both Zones 2 and 3 of Read and Mamay (1964). The oldest marine fossils from beds lying above the plant-bearing strata are now known to be assignable to Mamet's Microfaunal Assemblage Zone 13 (middle Meramecian); thus, the Early Mississippian age traditionally assigned to Floral Zones 1 and 2 of Read and Mamay (1964) seems to be justified, and Floral Zone 3 must begin as early as Meramecian.

Until very recently, no known fossils of Pennsylvanian plants had been reported from Alaska or, for that matter, from any part of the western Cordillera north of Washington. In 1976, however, Bruce Reed of the U.S. Geological Survey collected an assemblage of plants that S. H. Mamay (written commun., 1977) reported to be "Middle Pennsylvanian, probably no lower than Alleghenian * * * possibly Floral Zone 10 of Read and Mamay, 1964." The assemblage includes *Cordaites* sp., ?*Calamites* sp., *Neuropteris scheuchzeri* Hoffman, *Pecopteris* cf. *P. arborescens* Schlotheim, and *Pecopteris unita* Brongniart. These plants come from the upper half of the Mt. Dall Conglomerate in the Talkeetna C–5 quadrangle, about 5.23 km N. 76°E. from Mt. Dall (Reed and others, 1978).

IGNEOUS AND METAMORPHIC ROCKS

Igneous and metamorphic rocks of Carboniferous age are found in only a few places in Alaska. In northern Alaska, mafic dikes and sills are found within the Lisburne Group in the northeastern part of the Philip Smith Mountains quadrangle. Just to the north, in the southeastern Sagavanirktok quadrangle, volcaniclastic rocks and associated andesitic flows are present in about the same stratigraphic position. Fossils in the enclosing rocks suggest that these volcanic rocks are Late Mississippian in age.

In the east-central Alaska Range, the Mankomen Group is underlain by a predominantly mafic marine volcanic sequence, the Tetelna Volcanics of probable Carboniferous age. As the oldest fossils in the overlying dominantly volcaniclastic Slana Spur Formation are of Atokan age, the Tetelna must be Early Pennsylvanian or older.

In southeastern Alaska, both intrusive igneous rocks and volcanic rocks are known. In the Kuiu-Kupreanof Islands area, Muffler (1967) reported basaltic to andesitic pyroclastic rocks and pillow lavas, apparently of Early Mississippian age, in the lower part of the Saginaw Bay Formation. On Prince of Wales Island, granitic rocks about 280 m.y. old intrude the Klawak Formation of Early and Middle Pennsylvanian age (Eberlein and Churkin, 1970). In the Nabesna quadrangle, the Ahtel pluton is also of latest Pennsylvanian age. Richter (1976) reported its age to be about 280 m.y. With the exception of these very late Pennsylvanian intrusive rocks, no plutonic rocks of Carboniferous age are known in Alaska.

Metamorphic rocks of Carboniferous age are very poorly known; the only possible candidate is the Totatlanika Schist of the central Brooks Range. Wahrhaftig (1968) reported that fossils of probable Mississippian age are found in this terrane which occupies a northeast-trending belt in which Permian rocks lie unconformably on early Paleozoic or older rocks. Significantly, this belt also includes the only known nonmarine Carboniferous plant-bearing strata in the State.

ECONOMIC PRODUCTS

The major commodities in the Carboniferous of Alaska are oil and gas. The carbonate rocks of the Lisburne Group in the subsurface of northern Alaska provide one of the main reservoir zones for storage of hydrocarbons. In the Prudhoe Bay field, porosity zones in the upper Lisburne are primary targets for drilling. Facies trends indicate that Chesterian (upper Alapah Formation) strata contain good to excellent dolomite porosities in a broad band that extends generally west from the Prudhoe area to the northwestern Arctic Coast in the vicinity of Wainwright (Armstrong and Mamet, 1970). Unfortunately, most of the part of this band that is west of the Colville River is probably too deep for exploration. However, in the region east of the Colville, from the foothills north to the coast, good possibilities exist for reaching the Lisburne in a number of structural settings.

An analysis of the discovery well in the Prudhoe field, Prudhoe Bay State 1, by Armstrong and Mamet (1974) indicates that the Lisburne Group is 540 m thick and ranges in age from Mamet's Foraminifera Zone 16_{sup} to Zone 21. The Alapah Limestone is

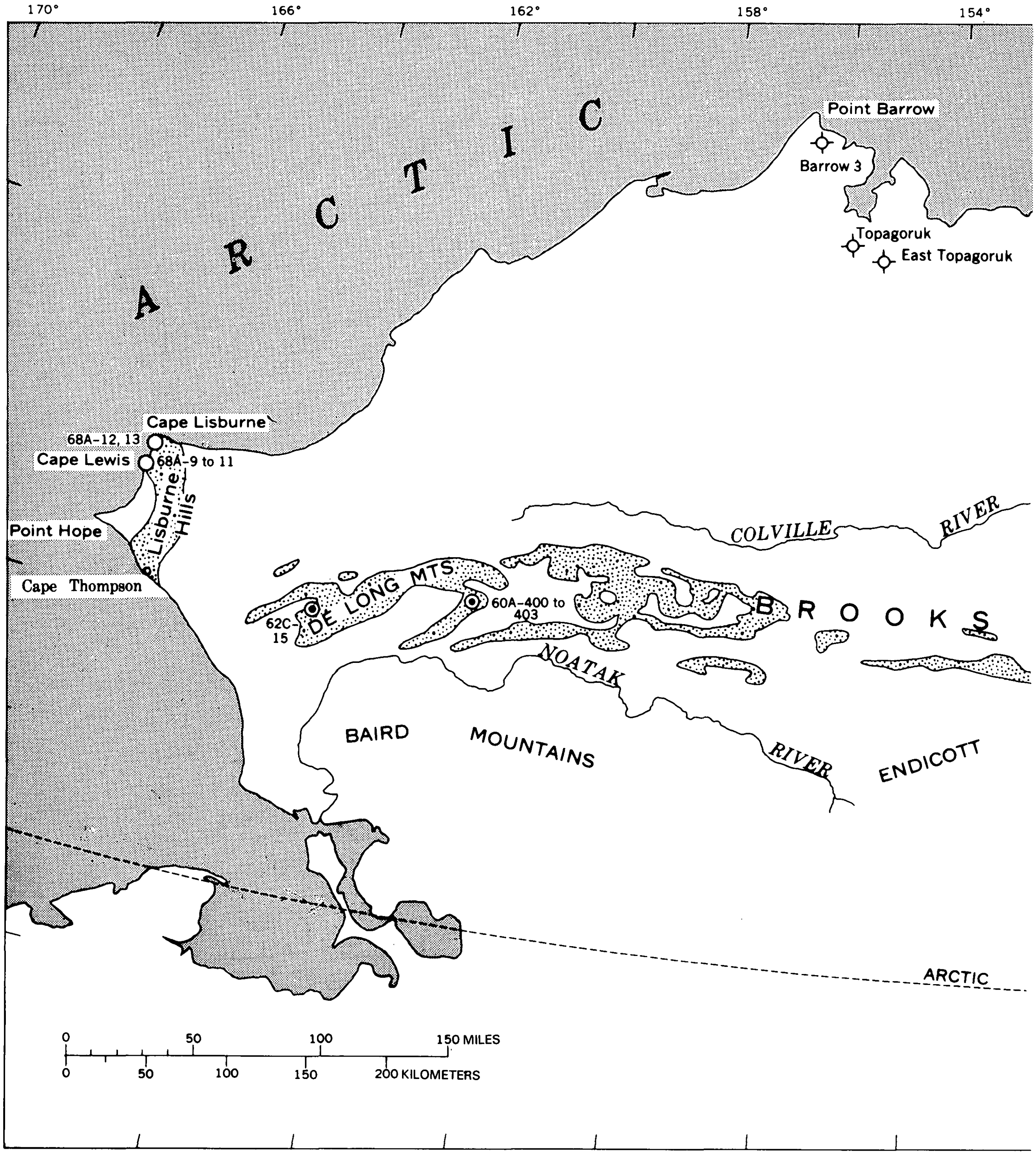
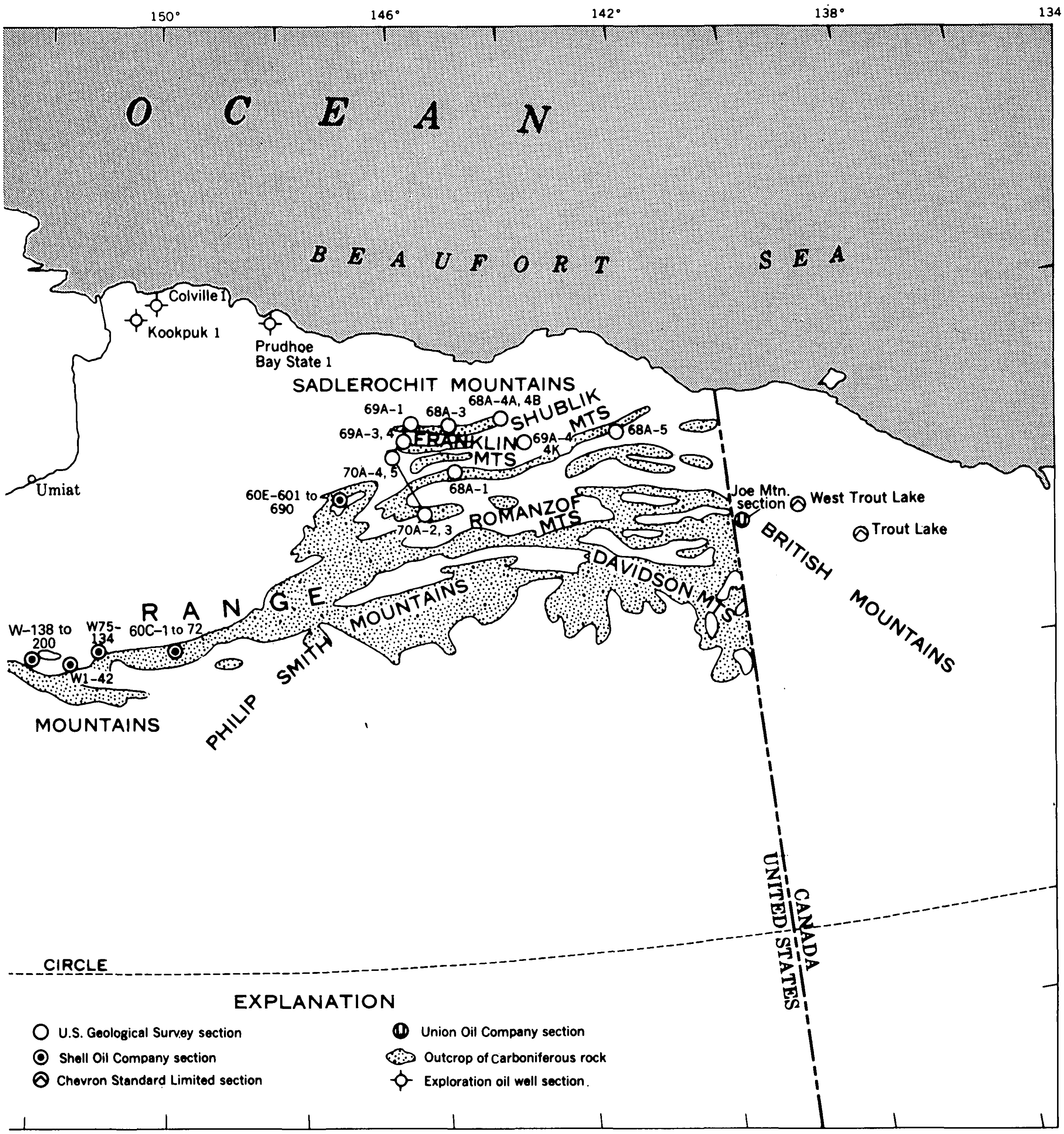


FIGURE 3.—Index map of arctic Alaska and adjacent Canada showing locations of outcrops of Carboniferous rock in Alaska

320 m thick and is composed of arenaceous packstone and wackestone, lime mudstone, and gypsiferous microdolomite. The Alapah is entirely Chesterian in age, ranging from Zone 16_{sup} through Zone 18. The overlying Wahoo Limestone, 219 m thick, is a series of cycles composed of packstone and wackestone capped by lime mudstone and arenaceous microdolomite. These deposits were interpreted by Armstrong and Mamet (1974) to represent tidal-flat lime muds directly above ooid tidal bars. The



and locations of sections, some of which contain the Lisburne Group (fig. 2). From Armstrong and Mamet (1977, fig. 1).

Wahoo Limestone is entirely Pennsylvanian (Foraminifera Zones 20 and 21). The Alapah Limestone is more arenaceous and dolomitic in the Prudhoe Bay State 1 well than are its outcrop equivalents in the Franklin and Sadlerochit Mountains. On the other hand, the Wahoo Limestone is very similar to outcrop sections in the Sadlerochits and represents a similar depositional environment.

The oil and gas potential of the Lisburne Limestone in the subsurface of northern Alaska was re-

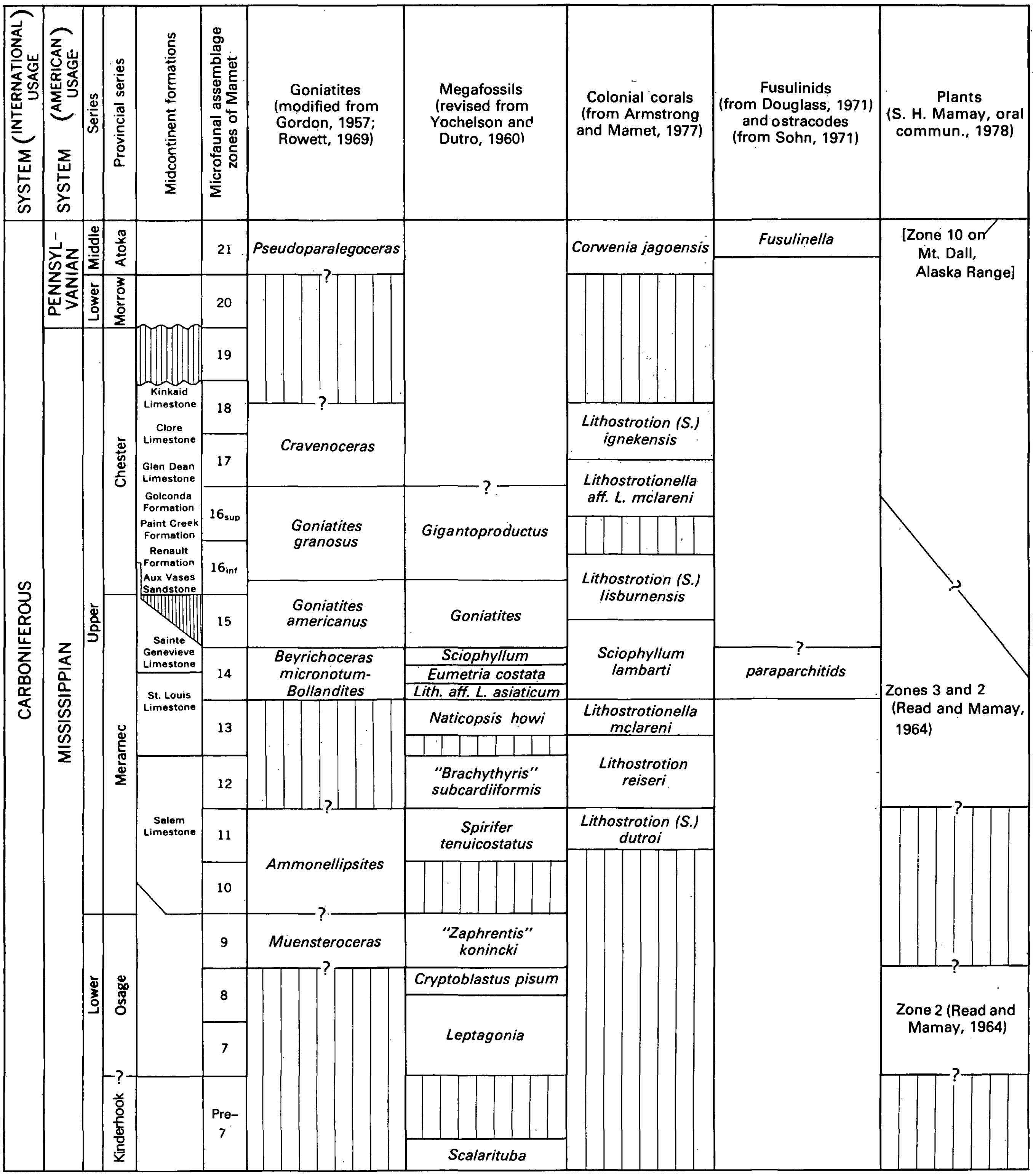


FIGURE 4.—Biostratigraphic zonations in the Carboniferous deposits of Alaska.

cently analyzed by Bird and Jordan (1977) who reviewed the regional setting and reservoir potential of the Lisburne Group. According to Bird and Jordan, although the Lisburne had been considered a probable petroleum reservoir for many years, these carbonate rocks were not penetrated by drilling until 1966. In that year, both the Union Kookpuk No. 1 and the Sinclair Colville No. 1 reached the

Lisburne; both wells had porosity and oil shows. The Arco-Humble Prudhoe Bay No. 1, discovery well for the Prudhoe field in 1968, also penetrated the entire Carboniferous section. Records and samples from 46 wells reaching the Lisburne are now available to the public and were used in the Bird and Jordan study.

An isopach map of the Lisburne in the eastern part of the Arctic Coastal Plain (fig. 5) shows that the thickest section is near the front of the Brooks Range where more than 1,200 m (4,000 ft) of strata are present. These carbonate deposits thin northwestward to less than 150 m (500 ft) over a broad northwest-trending arch that is a major subsurface structural feature. To the northeast, near the mouth of the Canning River, the top of the Lisburne has been truncated below the ground surface, and the Carboniferous carbonate deposits are completely absent near the Arctic Coast.

Porosity in the Lisburne strata is of three kinds. Microdolomite has porosity values as high as 23 percent from laboratory measurements and 27 percent as calculated from well logs. This is mostly intergranular porosity, but, in addition, the presence of vugs and small vertical fractures enhances primary porosity. The fractures, more importantly, also provide increased permeability. Bird and Jordan (1977) suggested that an average porosity of at least 10 percent may be expected in the subsurface.

Oolitic grainstones are common in the Lisburne, and, although the open-framework packing has nearly always been found filled with secondary calcite, this facies must be considered for its potential contribution to the overall porosity of the carbonate deposits.

Although the extent of these strata is unknown, sandstone porosity may be important in offshore wells in the Beaufort Sea. The dolomite thicknesses and the possible distribution of sandstone in the Lisburne Group are shown in figure 6.

Hydrocarbons have been reported from the Lisburne in three wells of the Prudhoe Bay field and from another well southeast of the field. Other wells have yielded saltwater from the Lisburne part of the column. Oil and gas flow rates from the Lisburne have not been disclosed as yet; the saltwater flow

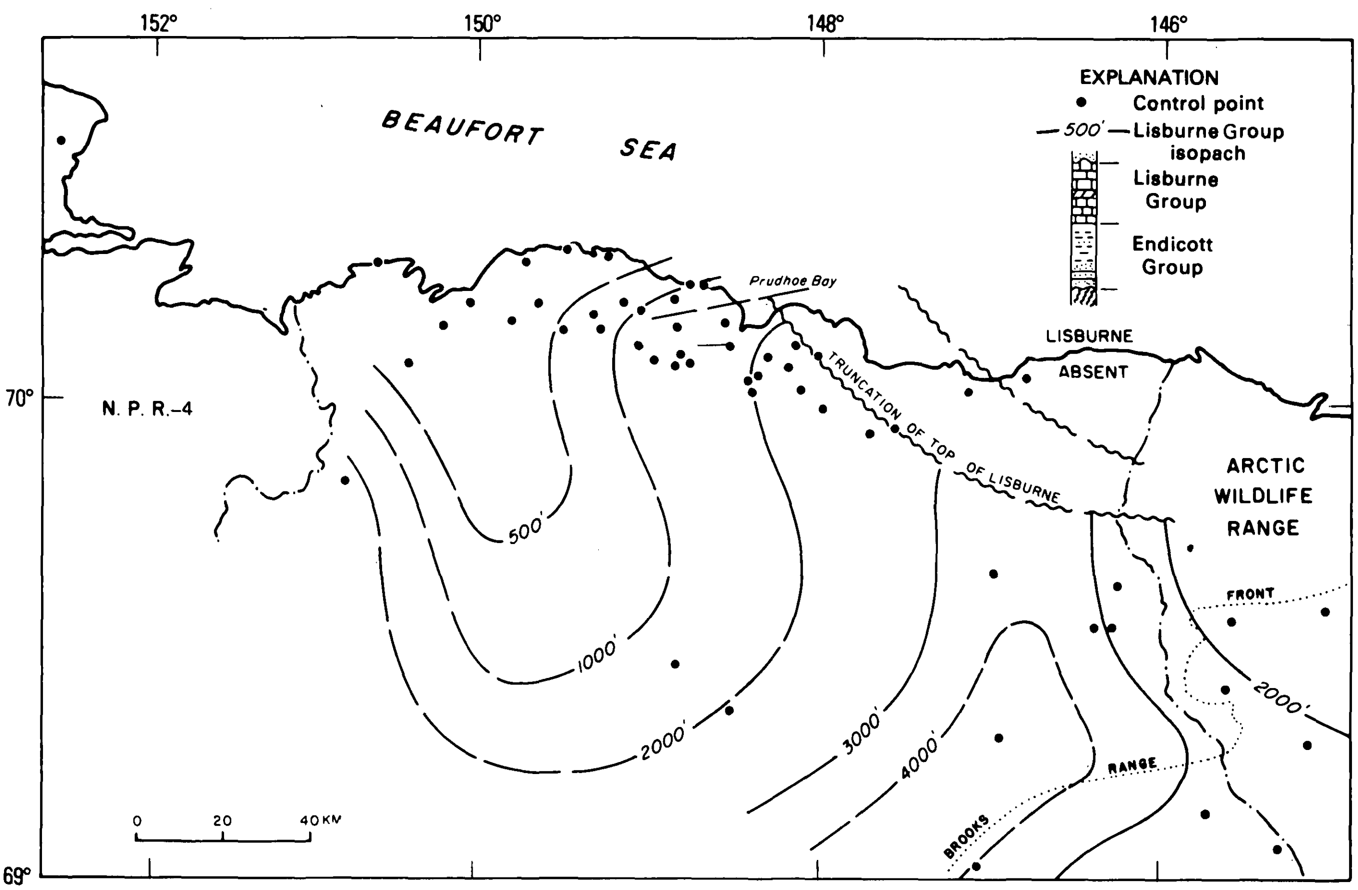


FIGURE 5.—Isopach map of Lisburne Group in the subsurface of the eastern Arctic Slope, Alaska. Modified from Bird and Jordan (1977).

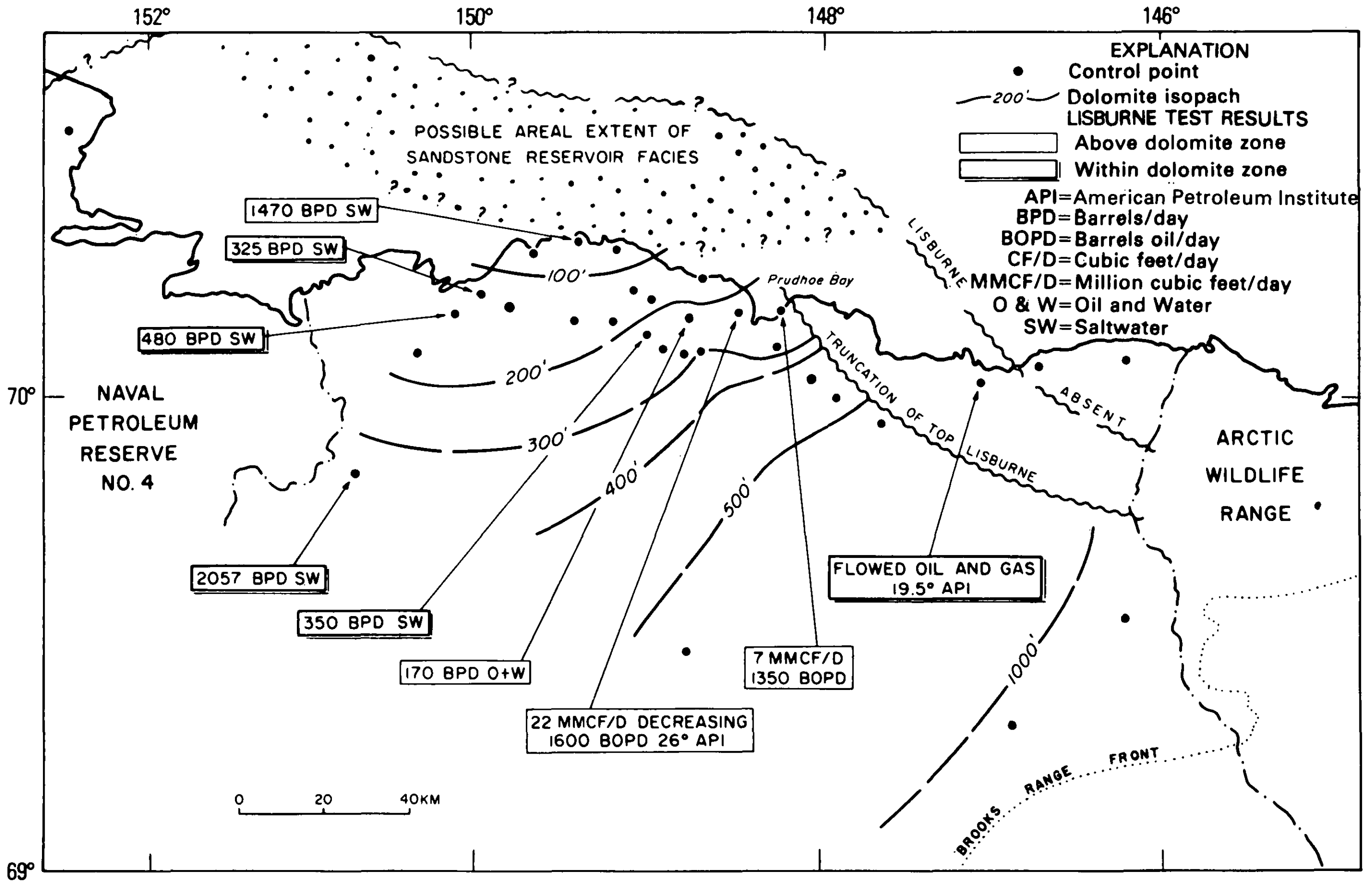


FIGURE 6.—Reservoir data summary for the Lisburne Group in the subsurface of the eastern Arctic Slope, Alaska. Modified from Bird and Jordan (1977).

rate from the dolomite in Mobil Mikkelson Bay No. 1 is 2,057 BPD (barrels per day) and from sandstone in the same well is 1,470 BPD.

Bird and Jordan (1977) suggested that, although the most likely source beds are Jurassic and Cretaceous shale, shale within the Lisburne Group and the Kayak Shale are possible contributors as well. They also indicated that coal in the basal Kekiktuk Conglomerate may be a source for dry gas in the Lisburne. Various estimates indicate that, although this coal in the Prudhoe Bay area has not been buried deep enough or for a long enough period to generate gas, downdip correlatives to the south may have generated gas.

Several formations provide adequate sealing beds at the top of the Lisburne Group throughout the subsurface area. Stratigraphic and structural traps and combinations of both are present in northern Alaska. Along the Barrow arch, traps of the Prudhoe Bay type may be expected. In the Prudhoe Bay field, the Sadlerochit pool was shown by Rickwood (1970) and Morgridge and Smith (1972) to be on a west-plunging anticlinal nose, faulted on both the north and south, and truncated and sealed by an unconformity to the east. The Lisburne pool, presumably, is similarly trapped.

Although nowhere commercially developed, the Carboniferous carbonate rocks of Alaska provide an immense potential resource for various industrial uses of limestone. In northern Alaska alone, thousands of cubic kilometers of limestone are exposed in the Brooks Range, either at the surface or within a hundred meters of the surface. No detailed chemical analyses have been made on these limestones, but the various carbonate facies that have been examined during field mapping and stratigraphic studies indicate a full range of shelf-carbonate types, ranging from pure chemical-grade lime mudstone through argillaceous and dolomitic limestone to pure dolomite. If industrial limestone is needed and if economic factors are favorable, the Carboniferous limestones of Alaska may become an invaluable asset.

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