Migration of the Acadian Orogen and Foreland Basin Across the Northern Appalachians of Maine and Adjacent Areas

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By Dwight C. Bradley, Robert D. Tucker, Daniel R. Lux,
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Migration of the Acadian Orogen and Foreland Basin Across the Northern Appalachians of Maine and Adjacent Areas

By Dwight C. Bradley, Robert D. Tucker, Daniel R. Lux, Anita G. Harris, *and* D. Colin McGregor[1]

Abstract

We reconstruct seven sequential positions of the Acadian deformation front and foreland basin to illustrate the northwestward migration of the orogenic belt across Maine and adjacent areas from Late Silurian to Middle Devonian time. The reconstructions are based on (1) U-Pb and $^{40}Ar/^{39}Ar$ ages of pretectonic, syntectonic, and posttectonic plutons; (2) conodont and palynomorph ages of key strata that predate or postdate the local age of deformation; and (3) interpretations of the depositional settings of these strata: far foreland, foreland basin, inner trench slope, and orogenic hinterland. Tight correlations between isotopically and paleontologically dated events are made possible by recent improvements in the Silurian-Devonian time scale.

During early Ludlow time (~423 Ma), the deformation front must have lain near the present midline of the Fredericton Basin, between the posttectonic Pocomoonshine pluton (423 Ma) to the southeast and the coeval graptolitic Smalls Falls Formation, 70 km across strike to the northwest. During early Lochkovian time (~417 Ma), the deformation front was near the present midline of the Central Maine Basin, as recorded by the northwestward advance of slope deposits of the Carrabassett Formation across axially transported turbidites of the Madrid Formation. The early Emsian (407–406 Ma) deformation front lay along or near the Lobster anticlinorium; its position is bracketed between the posttectonic Russell Mountain pluton (406 Ma) to the southeast and the coeval brachiopod-bearing Tomhegan molasse, 50 km across strike to the northwest. At the Emsian-Eifelian boundary (~394 Ma), the deformation front was located along the Pennington-Munsungun anticlinorium, northwest of the posttectonic Mapleton and Trout Valley Formations but southeast of the youngest nonmarine clastic rocks in the pretectonic Fish River Lake Formation. At the Eifelian-Givetian boundary (~387.5 Ma), the deformation front was probably somewhere near the midline of the Connecticut Valley-Gaspé Basin, northwest of a belt of posttectonic plutons in Québec, including the 384-Ma Scotstown pluton, but southeast of three occurrences of deformed Eifelian carbonates along the basin's northwestern margin. At the Givetian-Frasnian boundary (~382.5 Ma), the deformation front was somewhere to the northwest of these Eifelian carbonate outcrops. The orogen thus migrated northwestward about 240 km across strike (present distance) in about 40.5 m.y. Meanwhile, on the outboard (southeasterly) side of the orogen, a boundary between deformed and undeformed rocks advanced southeastward at least 50 km, probably during the Early and (or) Middle Devonian.

The migration pattern of the orogen and foreland basin suggests that during collision, a southeasterly plate that included the Acadian orogenic wedge and its Avalonian backstop overrode the Taconic-modified margin of North America. The implied minimum plate-convergence rate is about 6 mm/yr. The actual rate must have been considerably faster because this calculation was done on a nonpalinspastic base map; a more accurate estimate will have to await a careful assessment of Acadian shortening. If shortening reduced Maine to half its pre-Acadian width (a conservative estimate in light of the regional-scale tight to isoclinal folding), this situation would imply a plate-convergence rate of 12 mm/yr, at the slow end of the normal range of modern plate motions. The reconstructed positions of the orogen and foreland basin also constrain the setting of several Silurian and Devonian episodes of volcanism and plutonism, certain post-Acadian deformations in the Acadian orogenic hinterland, and certain pre-Acadian deformations in what was then the Acadian foreland.

Introduction

More than three decades after Wilson's (1966) first plate-tectonic interpretation of the Appalachians, a consensus on the plate geometry that led to the Acadian orogeny still has not emerged, though not for lack of trying. In this report, we present new results and review old evidence bearing on orogenic timing on a regional scale. By focusing on the syncollisional history, we can make new headway on the Acadian problem without getting bogged down in controversies regarding the precollisional plate geometry.

Donahoe and Pajari (1973) presented evidence that the Acadian orogeny was diachronous across strike, beginning in

[1]Geological Survey of Canada Contribution No. 1998178.

the Early Devonian along the Maine-New Brunswick coast and ending during the Middle Devonian in Québec, near the Gulf of the St. Lawrence. This finding was based on the age and distribution of fossiliferous strata known to predate and postdate deformation and on isotopic ages of posttectonic plutons. Five developments make it worthwhile to revisit this topic. First, the Silurian-Devonian time scale has undergone major revisions (Tucker and McKerrow, 1995; Tucker, and others, 1998). The new time scale is calibrated by U-Pb zircon ages of ash beds with good biostratigraphic control; most of the series and stage boundaries are 10 to 20 m.y. older than on the time scale available to Donahoe and Pajari (1973), and the durations of individual series and stages differ widely as well. Second, many Acadian plutons have now been reliably dated by U-Pb and $^{40}Ar/^{39}Ar$ methods. The new ages, with errors of 1 to 3 m.y., reveal that most of the Rb-Sr and conventional K-Ar ages used in Donohoe and Pajari's (1973) analysis were about as far off as the time scale. Third, certain key stratigraphic units that bear on the position of the deformation front over time have now been dated by using conodonts or spores with much greater precision and accuracy than was ever possible by using brachiopods, corals, and plants. Fourth, paleocurrent directions and depositional environments are now known for several key stratigraphic units, such as the Madrid and Carrabassett Formations (Hanson and Bradley, 1989, 1993; Bradley and Hanson, 1989); these data have helped us to locate the foreland basin during two time intervals—and, by inference, to locate the deformation front at the same times. Finally, as we will show, current understanding of orogenic wedges, flexural foreland basins, and orogenic backstops makes it possible to glean more information from conventional stratigraphic evidence about orogenic timing than was possible in the 1970's.

New U-Pb, $^{40}Ar/^{39}Ar$, conodont, and palynologic data, which were the basis for the regional tectonic conclusions by Bradley and others (1998) and Robinson and others (1998), are documented here. These new results, as well as previously published data, enable us to track seven sequential positions of the deformation front as the first wave of Acadian deformation migrated cratonward across the northern Appalachians during Late Silurian to Middle Devonian time. This information has implications for (1) the rates, trajectories, and amounts of plate convergence; (2) the syncollisional plate geometry; (3) strike-slip and thrust partitioning during collision; and (4) the relation between various suites of igneous rocks and the orogenic belt as it existed at the time of magmatism.

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Regional Geology

The study area is the Acadian-deformed belt of Maine, plus adjacent parts of Vermont, New Hampshire, the Eastern Townships of Québec, and New Brunswick. The regional geology was reviewed by Osberg and others (1989) and Robinson and others (1998). The area of figure 1 is subdivided into 10 map units, 7 of which are relevant to the timing of Acadian deformation. Middle Ordovician and older rocks are grouped together in figure 1 because they provide no useful constraints on the age of Acadian tectonism.

Deep-water Silurian strata define four basins: the Connecticut Valley-Gaspé Basin, the Aroostook-Matapedia Basin, the Fredericton Basin, and the Central Maine Basin (fig. 2), the last two of which have been identified by some workers (Bradley, 1983; Ludman and others, 1993) as the former site of an ocean whose closure resulted in the Acadian orogeny. Although this interpretation remains controversial, the nature of the depositional basement of these deep-water basins makes little difference for the present study of syncollisional tectonics.

Flanking the deep-water basins are four tracts where pre-Silurian rocks are now exposed. These belts—Avalonia, the Bronson Hill-Pennington anticlinorial belt, Miramichi, and the Taconic orogen (fig. 2)—either were undergoing active erosion during at least part of the Silurian or were the sites of shallow-water deposition. The shallow-water deposits include carbonate and locally derived siliciclastic rocks that range in age from Llandovery to Eifelian. One goal of our study was to obtain precise conodont ages from the youngest carbonate rocks in various key stratigraphic sections.

From Late Silurian to Middle Devonian time, much of the area of figure 1 was inundated by a thick succession of clastic rocks, which we regard as the fill of a migrating Acadian foreland basin (Bradley, 1983, 1987, 1997; Hanson and Bradley, 1989). The foreland-basin succession includes all of the Devonian flysch and Devonian molasse and part of the deep-water Silurian sequence. For want of sedimentologic studies, the Silurian deep-water deposits cannot yet be confidently subdivided into foreland-basin and pre-foreland-basin successions. A second focus of the present study was to obtain more tightly constrained fossil ages from key foreland-basin deposits. Palynology has proved to be most useful in this regard; several calcareous concretions from siliciclastic turbidites were processed for conodonts but were found to be barren.

Silurian and Devonian volcanic rocks occur in two broad belts (figs. 1, 2). Rocks of the Coastal volcanic belt were erupted into basement of the Avalonian terrane along the coast of Maine and New Brunswick. Farther north lies a second belt of Silurian-Devonian volcanic rocks that crop out along parts of the Bronson Hill-Pennington anticlinorial belt, the Aroostook-Matapedia Basin, the Miramichi anticlinorium, and the Connecticut Valley-Gaspé Basin. The northerly belt has been called the Piscataquis volcanic belt (or magmatic belt) in New

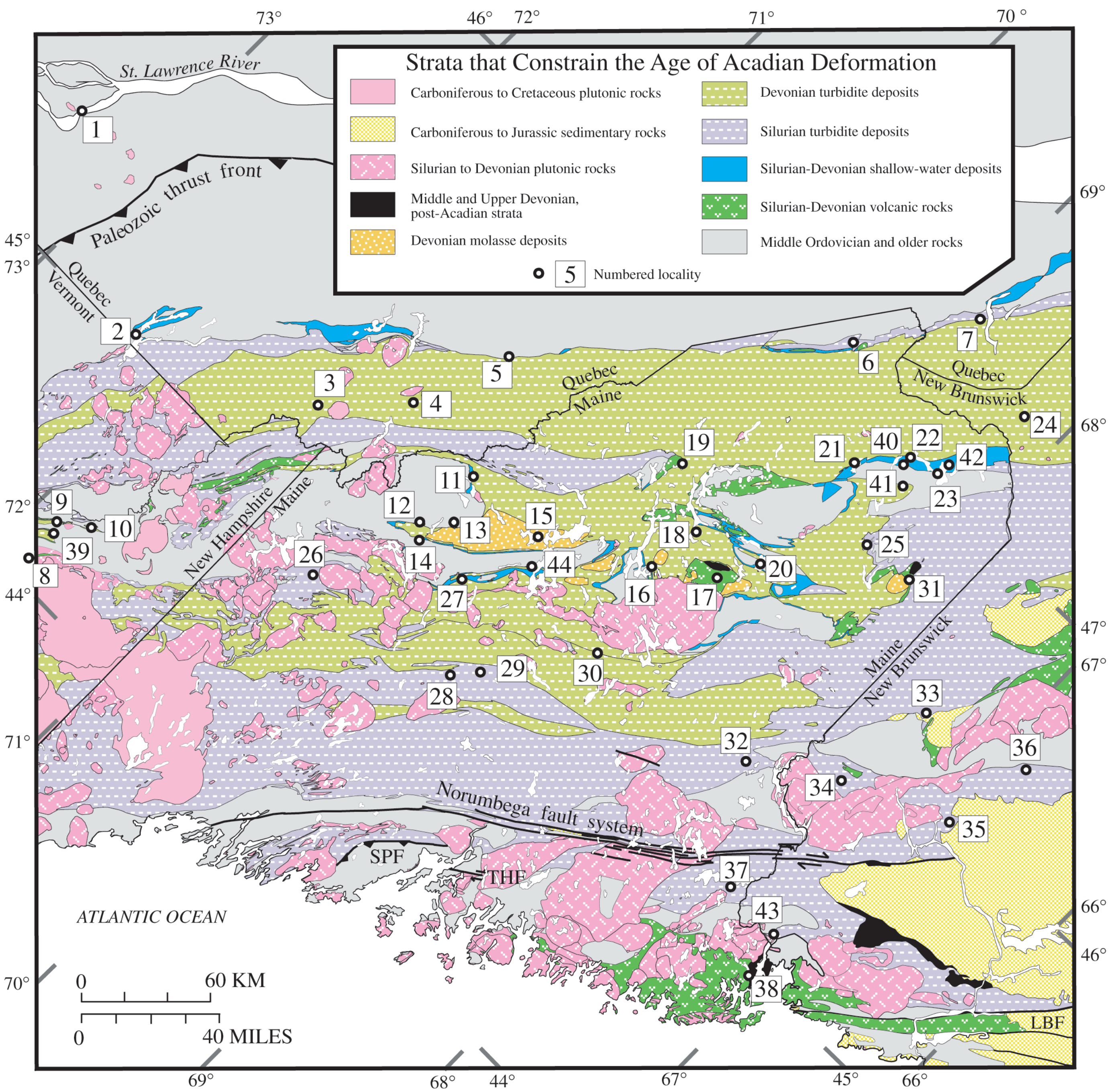


Figure 1.—Geologic map of Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing distribution of strata in study area that help constrain date of Acadian foreland-basin sedimentation and (or) Acadian deformation. Numbered localities are keyed to figure 4. LBF, Lubec-Belleisle Fault; SPF, Sennebec Pond Fault; THF, Turtle Head Fault.

England (Bradley, 1983) and the Tobique volcanic belt in Canada (Keppie and Dostal, 1994). The Piscataquis volcanic rocks were erupted in a belt that had been accreted to North America during the Ordovician.

In all but the northernmost part of the study area (fig. 1), the rocks were deformed during a Silurian to Devonian orogeny that can be termed "Acadian" in the loose sense favored here. Where deformation was polyphase, the date of the first deformation event (D1) is of greatest interest for the present purposes. The style and intensity of Acadian deformation vary considerably across the study area; the common thread is that the deformation was dominantly contractional. In the low-grade rocks of northern and central Maine, Acadian structures are mostly upright, tight to isoclinal folds. At higher metamorphic grades in central New Hampshire, east-directed fold nappes predominate (Eusden and others, 1996); in southern New Hampshire, nappes have westerly vergence (Thompson and others, 1968; Robinson and others, 1998). Some strike-slip faulting also occurred during Late Silurian and Devonian times, as discussed below.

Silurian and Devonian (Acadian) plutons, ranging in composition from gabbro to granite, are widespread in the study area (fig. 1); many are demonstrably syntectonic or posttectonic. A major goal of our study was to date key plutons across a broad swath from the Maine coast to the Maine-Québec border, using U-Pb and $^{40}Ar/^{39}Ar$ methods. As we will show, the locus of plutonism shifted over time.

Of the various post-Acadian rocks in the study area (fig. 1), only Middle and Upper Devonian sedimentary and volcanic rocks that overlie Acadian-deformed rocks are relevant to

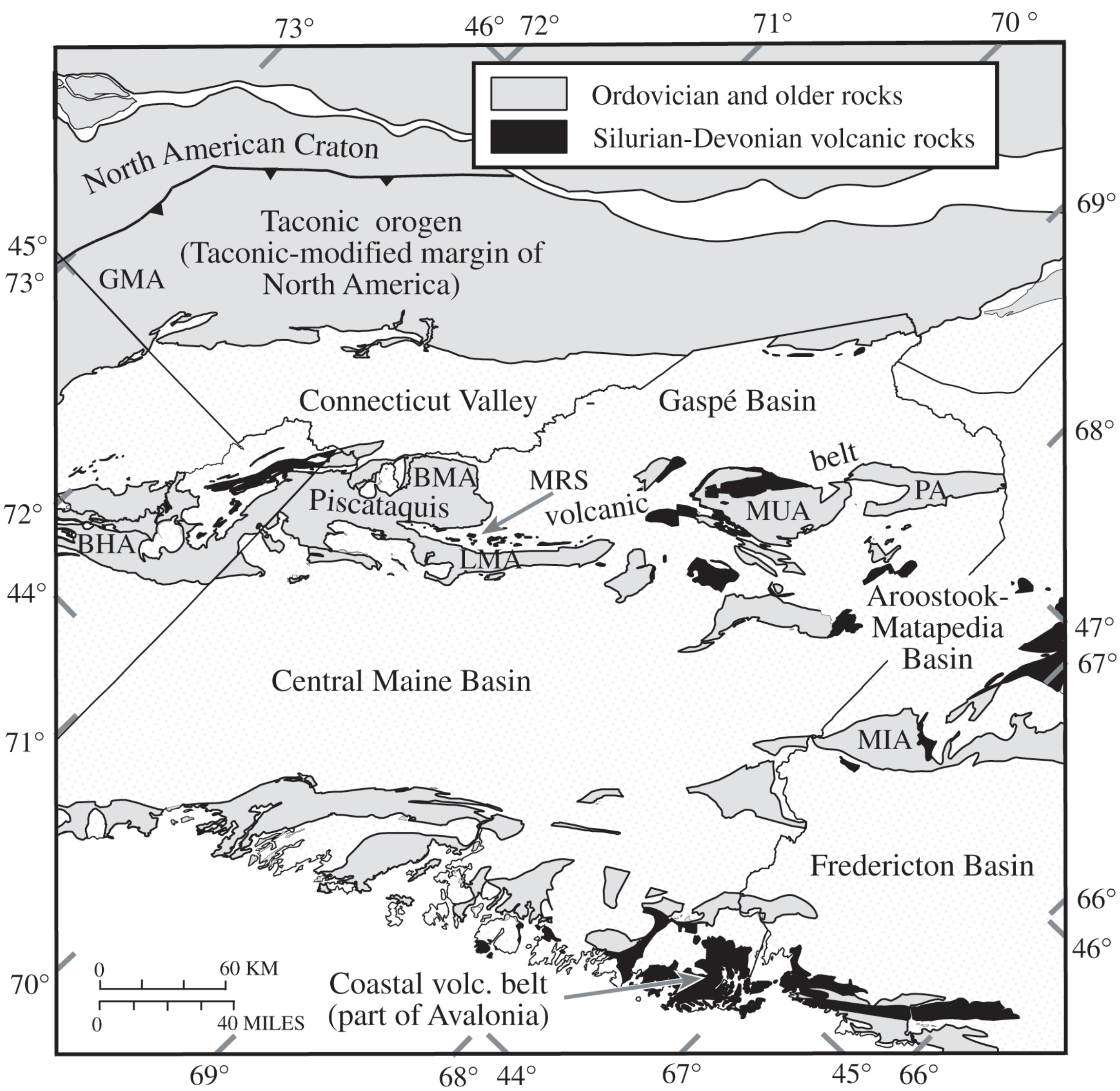


Figure 2.—Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing distribution of major Silurian-Devonian paleogeographic elements in study area. BHA, Bronson Hill anticlinorium; BMA, Boundary Mountains anticlinorium; GMA, Green Mountain-Sutton Mountain anticlinorium; LMA, Lobster Mountain anticlinorium; MIA, Miramichi anticlinorium; MRS, Moose River synclinorium; MUA, Munsungun anticlinorium; PA, Pennington anticlinorium.

orogenic timing. In this report, we present new palynologic ages for two key post-Acadian units, ages that previously had been documented only in unpublished paleontologic reports of the Geological Survey of Canada. Other post-Acadian rocks in the study area include Carboniferous to Cretaceous plutons and Carboniferous to Jurassic sedimentary rocks.

Determining the Timing of Collisional Orogeny from the Rock Record

Traditionally, orogenesis in a given place has been judged to be younger than the youngest deformed strata but older than the oldest unit that unconformably overlies the deformed rocks. For most formations in the study area (fig. 1), we know more than whether they participated in Acadian deformation: interpretations of the tectonic setting of the dated rocks can supplement more traditional evidence for orogenic timing.

Typical contractional orogenic systems have three components: (1) an *underriding plate*, (2) an *overriding plate*, and, between them, (3) an actively deforming *orogenic wedge* (fig. 3). Two broad depositional regimes exist side by side on the underriding plate. The *foreland basin* is a marine or nonmarine sedimentary basin that flanks the orogen and is typically filled with orogenically derived sediment. (The terms "foreland basin" and "foredeep" are sometimes used interchangeably, but a *foredeep*, more properly, is an underfilled foreland basin characterized by deep-marine flysch sedimentation.) The term "far foreland" is coined here for the

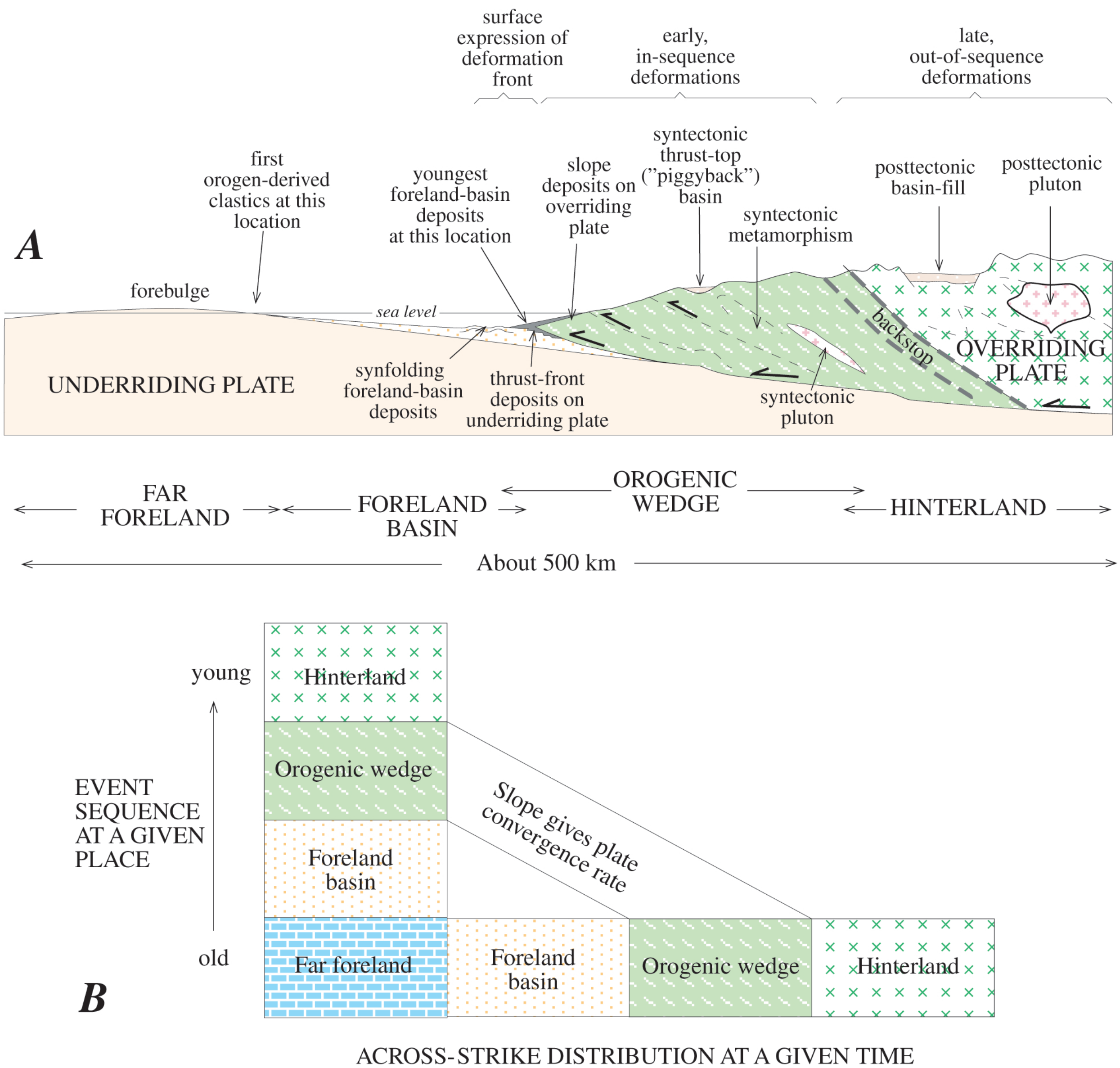


Figure 3.—Conceptual model of a two-plate collision. A, Cross section showing positions of far foreland, foreland basin, orogenic wedge, and hinterland, as discussed in text. B, Plate convergence causes each of these paleogeographic elements to migrate to left over time at a rate similar to plate-convergence rate, giving rise to succession of depositional and deformational events in a given rock sequence.

region beyond the limit of orogen-derived sedimentation. A *forebulge*—a broad, gentle uplift formed by lithospheric flexure—may or may not be present in the distal part of a foreland basin or in the far foreland.

The *deformation front* is the surface expression of the gently dipping boundary between the underriding plate and orogenic wedge; it may be a sharp boundary at an emergent thrust fault or a broad zone of folding. Much, but not all, of the convergence between the underriding and overriding plates typically occurs at and just to the rear of the deformation front and is manifested as the D1 event that is of greatest interest here. Some additional convergence may be taken up farther to the rear, within the orogenic wedge, accounting for some D2 and later events. The boundary between the actively deforming orogenic wedge and the relatively rigid—or even extensional—*hinterland* on the overriding plate is called the *backstop*.

The model just outlined is sufficiently generalized that it applies to several plate-tectonic regimes, such as arc-passive margin collision, Andean-style foreland contraction, and arc-arc collision. Figure 3 might be thought of as a single frame of a moving picture; as time goes on, new rocks of the underriding plate are fed into the orogenic system.

Most of the relevant Silurian-Devonian strata in the Acadian orogen can be placed with at least fair confidence in figure 3. In the descriptions of stratigraphic sections that follow, we focus not only on the conventional stratigraphic evidence (preorogenic versus postorogenic) but also on observations that enable us to place each section in the far foreland, foreland basin, orogenic wedge, or hinterland at a particular time. We use only those age constraints provided by fossils where they actually occur; the New England tradition of extrapolating fossil ages *across strike* hinders the recognition of diachronous facies, which are characteristic of all collisional orogens in the world.

Paleontologic and Stratigraphic Evidence for the Date of the Acadian Orogeny

Passamaquoddy Bay, Maine and New Brunswick

The Coastal volcanic belt section of southeastern Maine (loc. 38, figs. 1, 4) and southwestern New Brunswick consists of 8 km of interbedded bimodal volcanic and fossiliferous sedimentary rocks of late Llandovery to Lochkovian age. The following highlights (from Gates and Moench, 1981) are key to Acadian tectonic problems. The late Llandovery Quoddy Formation, mostly graptolitic shale, contains a few tuffaceous horizons that record the first known pulse of Silurian volcanism in the belt (loc. 38, fig. 4). Mixed volcanism and ever-shallowing marine sedimentation continued throughout Wenlock, Ludlow, and Pridoli time and the first part of the Lochkovian (Dennys, Edmunds, Leighton, Hershey, and lower Eastport Formations). In New Brunswick, the upper part of the Eastport Formation consists of fluvial sandstone, siltstone, and conglomerate (Pickerill and others, 1978). The Eastport Formation, which is the youngest pre-Acadian stratified unit, has been assigned a Lochkovian age on the basis of a restricted fauna of lingulids, gastropods, pelecypods, and ostracodes (Berdan, 1971; Pickerill and Pajari, 1976). We note that the Castine Volcanics in the Penobscot Bay region, which were once regarded as Silurian and thus part of the Coastal volcanic belt, have recently yielded a U-Pb zircon age of 505 Ma (R.D. Tucker, unpub. data., 1997) and so are irrelevant to the Acadian orogeny. All the volcanic units were involved in Acadian folding and development of a regional east-northeast-striking cleavage (Pickerill and others, 1978, p. 42). The Red Beach pluton has yielded two conflicting isotopic ages, both suspect: a Givetian Rb-Sr isochron age of 385±6 Ma (Spooner and Fairbairn, 1970; age recalculated herein) and a Lochkovian U-Pb age of 415±6 Ma (Jurinski, 1990). The nominal U-Pb age is likely too old because the pluton truncates regional-scale Acadian folds that deform the Lochkovian Eastport Formation (Abbott, 1991); however, Rb-Sr ages from coastal Maine plutons have consistently proved to be too young. Accordingly, the age of folding cannot be determined with confidence. The Perry Formation unconformably overlies both the Eastport Formation and the Red Beach pluton. The basal conglomerate of the Perry Formation contains boulders of unmistakable Red Beach granite. Plant fossils suggest a general Late Devonian age for the Perry Formation (Kasper and others, 1988, p. 127); an ash bed in the Perry Formation was processed for datable zircons but was barren.

Fredericton Basin, Maine and New Brunswick

In easternmost Maine, the thick, isoclinally folded turbidites of the Fredericton Basin (or "Trough" of some workers) are devoid of fossils (loc. 37, figs. 1, 4; Ludman and others, 1993). Nonetheless, some useful information regarding the timing of Acadian deformation can be gleaned. The youngest turbidite unit, the Flume Ridge Formation of presumed Silurian age (Ludman and others, 1993), was penetratively deformed before being intruded by the 423±2-Ma Pocomoonshine pluton (see below). In New Brunswick, however, graptolites from turbidites (loc. 35, figs. 1, 4) reveal that sedimentation was still taking place in the northwestern part of the basin at this time. Specifically, the Burtts Corners Beds have yielded graptolites ranging in age from *Cyrtograptus linnarssoni* Zone (Wenlock) to *Monograptus nilsonni* Zone (early Ludlow) (Fyffe, 1995, p. 352–353).

Canterbury Inlier, New Brunswick

The Canterbury inlier (loc. 34, figs. 1, 4) lies along the southeast flank of the Miramichi anticlinorium. The inlier is separated by the Pokiok batholith from the main part of the Fredericton Basin, and the stratigraphy differs markedly. The Silurian-Devonian section is poorly fossiliferous, but the youngest stratified unit (Hartin Formation), which is of greatest interest here, is dated reasonably well. The Hartin Formation is composed of sandstone and slate, plus minor limestone, conglomerate, and felsic volcanic rocks. A brachiopod fauna indicates a Helderbergian age (that is, Lochkovian; A.J. Boucot, in Venugopal, 1979, p. 17). The Hartin Formation has been contact-metamorphosed by the Harkshaw and Skiff Lake phases of the Pokiok batholith (Venugopal, 1979, p. 28), which have yielded U-Pb ages of 411±2 Ma (sphene) and 409±2 Ma (zircon), respectively (Bevier and Whalen, 1990a). Along strike in easternmost Maine, the Skiff Lake phase truncates Acadian structures that deform Silurian-Devonian strata (Ludman, 1990), and so appears to be broadly posttectonic. Acadian deformation in this area must therefore have occurred during the late Lochkovian or the Pragian. Given that Acadian deformation must have followed closely upon deposition of the Hartin Formation, a foreland-basin depositional setting seems likely.

Carlisle Area, New Brunswick

Acadian deformation is anomalously young in the Aroostook-Matapedia Basin in western New Brunswick (loc. 33, figs. 1, 4). The Late Ordovician and Silurian were times of deep-water sedimentation (Carys Mills and Smyrna Mills Formations; St. Peter, 1982). The Smyrna Mills Formation is overlain by the Costigan Mountain Formation of predominantly felsic volcanic rocks; it is undated but regarded as probably Lower Devonian on the basis of correlations with fossiliferous rocks to the northeast (St. Peter, 1982, p. 35). The youngest pre-Acadian unit is the Wapske Formation, estimated to be 4 km thick. It consists of slate, siltstone, sandstone, and conglomerate, with isolated mafic volcanic lenses (St. Peter, 1982). Clasts in conglomerate of the Wapske Formation can be traced to the Smyrna Mills, Costigan Mountain, and other older units (St. Peter, 1982, p. 40). The lower part of the Wapske Formation has yielded brachiopods from five localities that, according to A.J. Boucot (in St. Peter, 1982, p. 45), indicate a late Helderbergian (Becraft-Port Ewen) age. This interval corresponds approximately to that of the *delta* and *pesavis* conodont zones of the Lochkovian. Spores of late Emsian or early Eifelian age have been recovered from the upper, nonmarine part of the formation (D.C. McGregor, in St. Peter, 1982, p. 478). We suggest that the Wapske Formation was deposited in a foreland-basin setting and that it was derived, at least in part, from an Acadian orogenic source to the south, which must have existed by about 411 Ma.

Presque Isle Quadrangle, Maine

The Presque Isle-Mapleton section (loc. 31, figs. 1, 4) provides age constraints on the timing of Acadian foreland-basin sedimentation, the main Acadian folding, and postorogenic hinterland sedimentation. Deep-water sedimentation (slate, deep-water limestone, siliciclastic turbidites) was already underway by the Caradoc and continued into the Ludlow (Carys Mills and Spragueville Formations and Perham Group; Roy and Mencher, 1976). A gap in the megafossil record, spanning the late Ludlow, Pridoli, and early Lochkovian, corresponds to the “Salinic Disturbance” of Boucot and others (1964). Whatever its cause, this hiatus did not last as long as once suspected—no more than 6 m.y., judging from the new time scale (Tucker and others, 1998). The next rocks to be deposited were volcanic rocks (Dockendorff Group), deltaic sandstone (Chapman Sandstone), and prodeltaic turbidites (Swanback Formation). Boucot and others (1964) showed these three units to be lateral equivalents of Lochkovian (“New Scotland”) age. Roy (1980) interpreted the Swanback and Chapman Formations as easterly derived flysch and molasse, respectively, which we assign to the Acadian foreland basin. Regional Acadian folding took place some time after deposition of the Lochkovian strata. The folded rocks are unconformably overlain by the Mapleton Formation, a local accumulation of nonmarine conglomerate and sandstone (Boucot and others, 1964). Conglomerate clasts in the Mapleton include fossiliferous clasts from the Chapman Sandstone and various older units (Boucot and others, 1964, p. 61–62; Roy and White, 1987). On the basis of plant fossils, the Mapleton was previously assigned a Middle Devonian, possibly Givetian age (Kasper and others, 1988). We now assign the Mapleton a slightly older age, early or middle Eifelian (*Grandispora velatus-Rhabdosporites langii* Zone), on the basis of spores from material submitted by Eli Mencher to D.C. McGregor during the early 1960’s (table 1; McGregor, 1962; new slides examined, 1989). The following forms were identified: *Apiculiretusispora gaspiensis*, *Ancyrospora* sp., *Calamospora* cf. *C. atava* (Naumova) McGregor, *Calyptosporites* sp., *Emphanisporites annulatus*, *E. rotatus*, *Grandispora velata*, *Stenozonotriletes* sp., *Dibolisporites echinaceus*, and *Acinosporites macrospinosus*?. The Mapleton is folded into a gentle syncline of middle Eifelian or younger age, which is nearly coaxial with an older, tighter syncline of the main phase of Acadian deformation (Boucot and others, 1964, p. 73).

Central Maine Basin

The stratigraphy of the Central Maine Basin is best known from its northwest flank in the Rangeley and Phillips quadrangles (loc. 26, figs. 1, 4; Moench, 1971), where the Silurian-Devonian deep-water section is about 8 km thick and most of the formations have their type localities (Moench and Pankiwskyj, 1988). Key sedimentary-facies interpretations and paleocurrent data for these units, however, are from 150 km to the east in the Jo Mary Mountain area (loc. 30, fig. 1; Hanson and Bradley, 1989), where the rocks are at much lower metamorphic grade and sedimentary features are better preserved. Complicating matters still further, the fossil control that is critical for present purposes is from yet a third area, Kingsbury-Guilford (locs. 28, 29, figs. 1, 4), which before Acadian shortening lay many tens of kilometers from the basin’s northwestern margin. The following discussion draws on information from all of these places.

Along the northwest flank of the basin, the Silurian-Devonian section is divisible into lower and upper parts on the basis of provenance and paleocurrents (Hanson and Bradley, 1993). Rocks assigned to the lower sequence (Greenvale Cove, Rangeley, Perry Mountain, and Smalls Falls Formations; see fig. 4) are known or inferred to have been derived from the northwest. A northwesterly provenance is well established for the Llandovery Rangeley Formation (at loc. 26, figs. 1, 4), which contains conglomerate clasts from the Attean pluton some 50 km to the northwest (near loc. 12, fig. 1; Moench and Pankiwskyj, 1988). Meager paleocurrent data suggest that the Smalls Falls is the youngest unit of northwest derivation (Hanson and Bradley, 1993). It consists of rusty-weathering turbiditic sandstone; the type Smalls Falls Formation is assigned a Ludlow age on the basis of fossils in broadly similar strata from a more southeasterly strike belt, as discussed below. The upper sequence, which was derived from outboard sources, consists of two widespread units—the Madrid and Carrabassett Formations—with poorly understood younger strata of local distribution. The Madrid Formation, in central Maine, is a thick succession of sandstone-dominated, siliciclastic turbidites; it becomes increas-

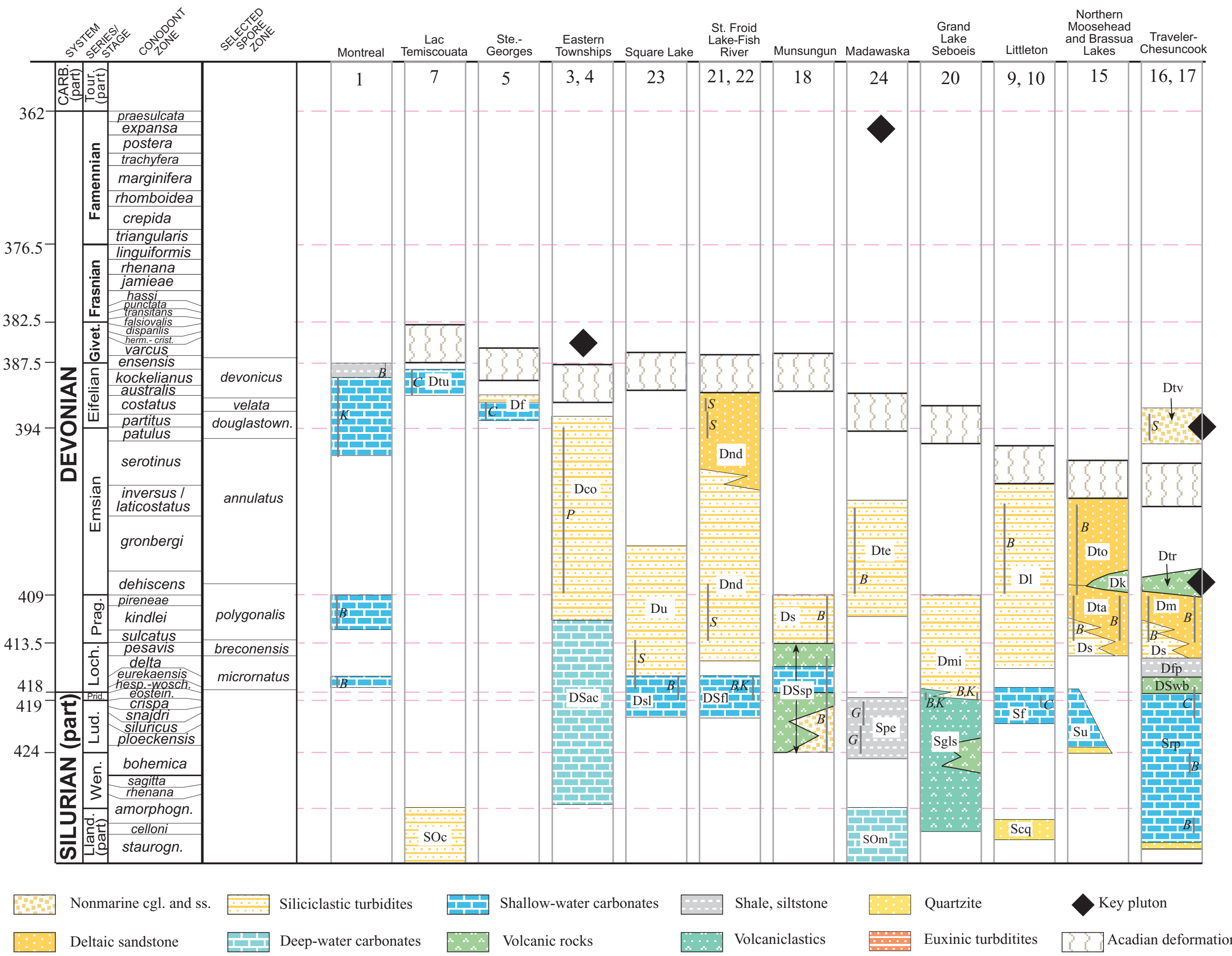
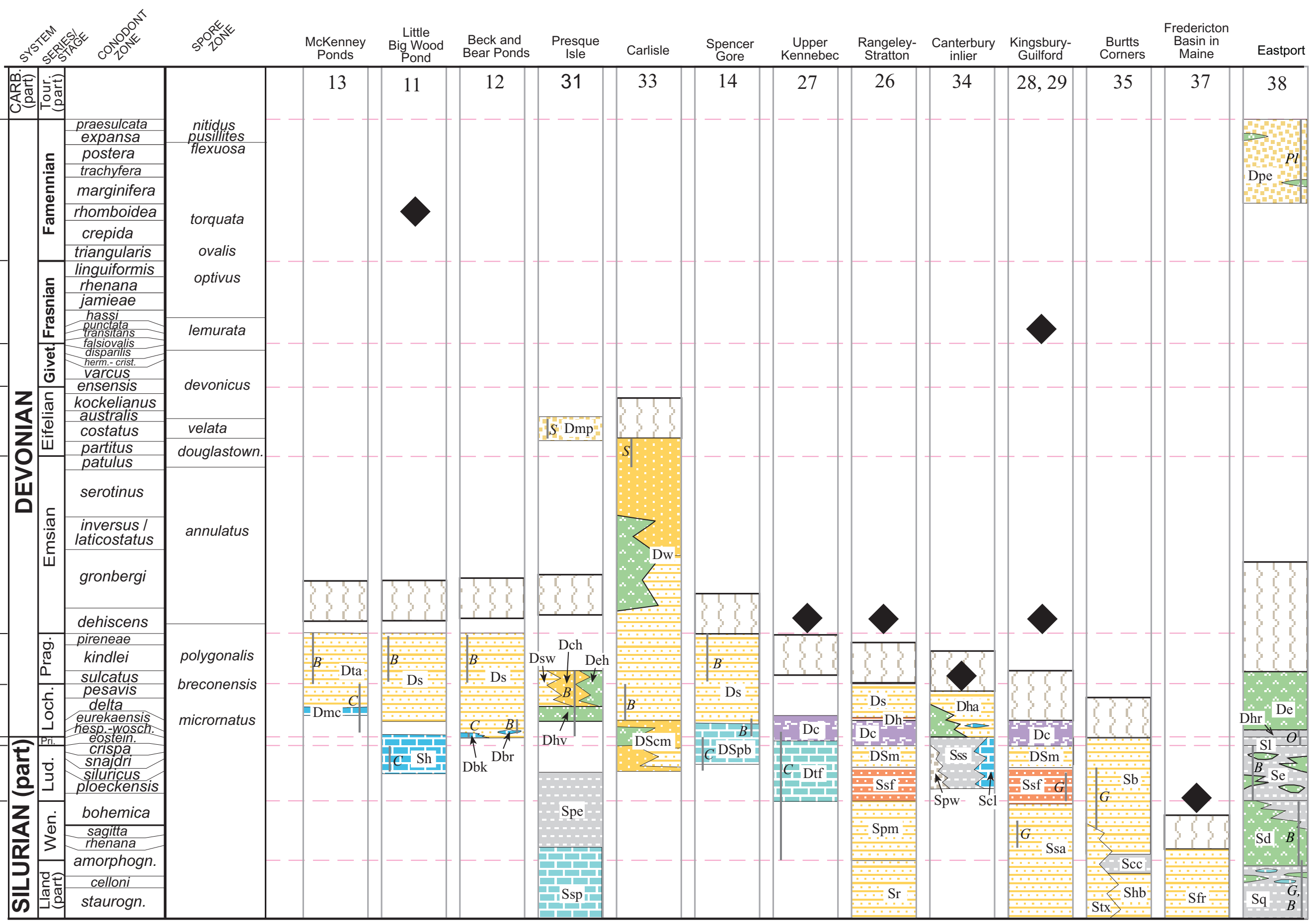


Figure 4.—Stratigraphic sections bearing on date of Acadian deformation and (or) foreland-basin sedimentation in study area (fig. 1). Numerical ages of various stage boundaries from time scale of Tucker and others (1998); spore zones from Richardson and McGregor (1986). Units: Dbk, Beck Pond Limestone; Dbr, Bear Pond Limestone Member; Dc, Carrabassett Formation; Dch, Chapman Sandstone; Dco, Compton Formation; De, Eastport Formation; Deh, Edmunds Hill Andesite; Df, Famine Limestone; Dfp, Frost Pond Formation; Dh, Hildreths Formation; Dha, Hartin Formation; Dhr, Hersey Formation; Dhv, Hedgehog Volcanics; Dk, Kineo Rhyolite; Dl, Littleton Formation; Dm, Matagamon Sandstone; Dmc, McKenney Ponds Limestone Member; Dmi, Millimagassett Lake Formation; Dmp, Mapleton Formation; Dnd, Nadeau Thoroughfare beds (informal); Dpe, Perry Formation; Ds, Seboomook Group (formerly Formation); DSac, Ayres Cliff Formation; DScm, Costigan Mountain Formation; DSfl, lower part of the Fish River Lake Formation; DShl, Hersey and Leighton Formations; Dsl, Square Lake Limestone; DSm, Madrid Formation; DSpb, Parker Bog Formation; DSsp, Spider Lake Formation; Dsw, Swanback Formation; DSwb, West Branch Volcanics; Dta, Tarratine Formation; Dte, Temiscouata Formation; Dtf, The Forks Formation; Dto, Tomhegan Formation; Dtr, Traveler Rhyolite; Dtu, Touladi Limestone; Dtv, Trout Valley Formation; Du, unnamed Devonian strata; Dw, Wapske Formation; Sb, Burtts Corners beds; Scc, Cross Creek beds; Scl, Canterbury Limestone; Scq, Clough Quartzite; Sd, Dennys Formation; Se, Edmunds Formation; Sf, Fitch Formation; Sfr,

ingly calcareous to the southwest in western Maine and New Hampshire. Paleocurrents clearly show a southwesterly direction of flow, along the basin axis (Bradley and Hanson, 1989). The gradationally overlying Carrabassett Formation consists of another kilometer or so of chaotic, pelitic olistostrome with subordinate, coherent sand and silt turbidites. Paleocurrents show overall northerly flow (Hanson and Bradley, 1993). Hanson and Bradley (1989) interpreted the Madrid and Carrabassett Formations as having been deposited along a foreland-basin axis and on a north-facing slope, respectively; however, neither the Madrid nor the Carrabassett Formation has yielded useful fossils. The Madrid Formation is early Ludlow or younger. The Carrabassett Formation has yielded nondiagnostic brachiopods and plant fragments consistent with an Early Devonian age (Espenshade and Boudette, 1967, p. F12–F13), and is cut by the earliest Emsian Sebec Lake pluton.

The depositional history of the central part of the Central Maine Basin is not as well understood as that of the northwest flank because we lack detailed sedimentologic studies and deformation and metamorphism in the basin are more intense. A few observations bear on the timing of Acadian deformation. Most of the area is underlain by the Sangerville Formation, an isoclinally folded succession of siliciclastic and calcareous turbidites many kilometers thick. The Sangerville



Bars in columns show age constraints by fossil groups:

B, brachiopod — C, conodont — G, graptolite
K, coral — O, ostracode — P, pelecypod
Pl, plant — S, palynomorph

Flume Ridge Formation; Sgls, Grand Lake Seboeis Formation; Sh, Hardwood Mountain Formation; Shb, Hayes Brook beds; Sl, Leighton Formation; SOc, Cabano Formation; SOm, Matapedia Group; Spe, Perham Formation; Spm, Perry Mountain Formation; Spw, Pocowogamis Conglomerate; Sq, Quoddy Formation; Sr, Rangeley Formation; Srp, Ripogenus Formation; Ssa, Sangerville Formation; Ssf, Smalls Falls Formation; Ssp, Spragueville Formation; Sss, Scott Siding Slate; Stx, Taxis River beds; Su, unnamed Silurian strata. References (numbers at top of columns): 1, Boucot and others (1986); 3, 4, Hueber and others (1990); 5, Uyeno and Lespérance (1997); 7, Uyeno and Lespérance (1997); 9, 10, Boucot and Arndt (1960); 11, Albee and Boudette (1972) and this report; 12, Boucot and others (1959) and this report; 13, Boucot and Heath (1969) and this report; 14, Boucot and Heath (1969) and this report; 15, Boucot and Heath (1969); 16, 17, Griscom (1976), Rankin (1968), and this report; 18, Hall (1970); 20, Hibbard (1993); 21, 22, Boone (1970) and this report; 23, this report; 24, St. Peter and Boucot (1981); 26, Moench and Pankiwksyj (1988); 27, Marvinney (1984) and this report; 28, 29, Pankiwskyj and others (1976), Ludman (1978) and this report; 31, Boucot and others (1964) and this report; 33, St. Peter (1982); 34, Venugopal (1979); 35, Fyffe (1995); 37, West and others (1992); 38, Gates and Moench (1981).

Table 1. Palynologic data from Devonian rocks, Maine

[GSC, Geological Survey of Canada. Do., ditto]

Rock unit	GSC number	Station, collector	Location (fig. 1)	Latitude, Longitude	Location	Lithology and structure	Age	Reference
Fish River Lake Formation.	0–106531	95MDw116A, Bradley	22	47°00′52″, 68°35′18″	Eagle Lake 7.5′ quadrangle. Nadeau Thoroughfare railroad cuts, 90–120 m NE. of bridge.	Medium sandstone, friable, plant bearing, 65° bedding dip, no cleavage	Most likely, and no older than, mid-late Emsian to very early Eifelian, *douglastownense-eurypterota* Zone. Spores from succeeding *velata-langii* Zone, early Eifelian, are absent.	McGregor (1996).
Do.	0–106532	95MDw117, Bradley	22	47°01′04″, 68°35′05″	Eagle Lake 7.5′ quadrangle. Nadeau Thoroughfare railroad cuts, 490–520 m NE. of bridge.	Medium sandstone, micaceous, friable, plant bearing, 51° bedding dip, no cleavage.	Most likely, and no older than, mid late Emsian to very early Eifelian, *douglastownense-eurypterota* Zone. Acritarchs suggest marine waters.	Do.
Do.	0–106533	95MDw118, Bradley	22	47°01′14″, 68°34′53″	Eagle Lake 7.5′ quadrangle. Nadeau Thoroughfare railroad cuts, 910–950 m NE. of bridge.	Medium sandstone, micaceous, friable, plant bearing, 50° bedding dip, no cleavage.	Probably assignable to, and no older than, mid late Emsian to very early Eifelian, *douglastownense-eurypterota* Zone. Acritarchs suggest marine waters.	Do.
Do.	0–106536	95MDw124, Bradley	22	47°02′30″, 68°33′48″	Eagle Lake 7.5′ quadrangle. Eagle Lake, south shore, near Camps of Acadia.	Gray micaceous siltstone, 65° bedding dip, strong cleavage.	Spores present, Late Silurian or younger	Do.
Do.	0–106537	95MDw125, Bradley	22	47°00′52″, 68°35′44″	Eagle Lake 7.5′ quadrangle. Nadeau Thoroughfare, under Route 11 bridge.	Gray siltstone, outcrop-scale folds with axial-planar cleavage.	Probably middle Pragian to earliest Emsian, *polygonalis-emsiensis* Zone. Could be next younger zone, but less likely.	Do.
Do.	6274	MK–62–8, Mencher	22	47°58′37″, 68°38′09″	Island Pond 7.5′ quadrangle. St. Froid Lake, W. shore; exact location uncertain.	Lithology unknown. Strongly cleaved, gray, micaceous siltstone observed near here in 1995.	Early Eifelian, *douglastownense-eurypterota* Zone	McGregor (1968).
Do.	6309	MK–62–9, Mencher	22	47°58′35″, 68°38′09″	Island Pond 7.5′ quadrangle. St. Froid Lake, W. shore, 30 m S. of MK–62–8; exact location uncertain.	Lithology unknown. Strongly cleaved, gray, micaceous siltstone observed near here in 1995.	Late Emsian to early Eifelian, *douglastownense-eurypterota* Zone.	McGregor (1962).
Do.	6612	CL–186, Mencher	40	46°58′26″, 68°34′32″	Winterville 7.5′ quadrangle. Route 11, 480 m N. of Winterville.	Lithology unknown; float sample.	Early to middle Eifelian, *velata-langii* Zone	McGregor (1963).
Fish River Lake Formation.	7754	EM–82, Mencher	22	47°02′58″, 68°30′02″	Eagle Lake 7.5′ quadrangle. Eagle Lake, N. shore, 80 m from east boundary of quadrangle.	Unknown	Middle Pragian or early Emsian	McGregor (1968).
Do.	7760	EM–980, Mencher	22	47°57′24″, 68°39′38″	Island Pond 7.5′ quadrangle. Red River, 640 m S. of Labbe Brook.	Unknown	Emsian	Do.
Mapleton Formation.	6275	MK–62–10, Mencher	31	46°46′51″, 68°04′57″	Presque Isle 15′ quadrangle. Same locality as Stop 2 of Roy and White (1987).	Unknown	Early to middle Eifelian, *velata-langii* Zone	McGregor (1992).
Trout Valley Formation.	0–105776	Unknown, Mencher	17	Unknown	Unknown	Unknown	Late Emsian to early Eifelian, *douglastownense-eurypterota* Zone.	Do.
Unit DSus of Osberg and others (1985).	0–106534	95MDw119, Bradley	23	47°06′36″, 68°22′57″	Square Lake 7.5′ quadrangle. 2.2 km N. of Burnt Landing on Square Lake access road.	Tan laminated siltstone, 90° bedding dip, weak cleavage.	Small spores suggest Late Silurian to Lochkovian	McGregor (1996).
Do.	0–106535	95MDw120, Bradley	23	47°07′26″, 68°22′52″	Square Lake 7.5′ quadrangle. 4.7 km N. of Burnt Landing on Square Lake access road.	Dark siltstone, 80° bedding dip, weak cleavage.	Undoubtedly Early Devonian, probably Lochkovian or early Pragian.	Do.
Do.	7758	EM–856, Mencher	41	Approx 46°57′, 68°27′	Portage 15′ quadrangle. Lumber road approx. 5.4 km S and 1.3 km E. from NW. corner of quadrangle.	Unknown	Probably Pragian	McGregor (1968).
Do.	7766	DR–1, Mencher	42	46°37′24″, 68°24′17	Ashland 15′ quadrangle. Sheridan Road opposite Ashland Cemetery in Ashland.	Unknown	Probably Pragian or Emsian	Do.

has yielded graptolites from 13 localities (Pankiwskyj and others, 1976); the most diagnostic collection (loc. 28, figs. 1, 4) is middle Wenlock. The overlying Smalls Falls Formation has also yielded graptolites from several localities; the most diagnostic collection (loc. 29, figs. 1, 4) is early Ludlow (Pankiwskyj and others, 1976). The overlying Madrid and Carrabassett Formations are the youngest units along the central part of the basin. The Carrabassett Formation is absent in the most southerly strike belt of Madrid Formation either because it has since been eroded away (Moench and Pankiwskyj, 1988) or because it was never deposited (Hanson and Bradley, 1989).

Littleton Area, New Hampshire

The classic Silurian-Devonian section near Littleton, New Hampshire (loc. 9, figs. 1, 4; Billings, 1937), constrains the timing of both Acadian foreland-basin sedimentation and deformation. The Lower Silurian Clough Quartzite unconformably overlies Ordovician rocks and is overlain by the Fitch Formation, which consists of calcareous metasiltstone, quartzite, granofels, and some limestone. Harris and others (1983, p. 731) reported an age of late Ludlow to mid-Pridoli for the Fitch Formation in the Littleton area, on the basis of a conodont fauna representing the *Ozarkodina remscheidensis eosteinhornensis* Zone; along strike in Massachusetts, strata assigned to the Fitch Formation are as young as the *woschmidti* to *eurekaensis* Zones of the Lochkovian (Elbert and others, 1988). The Clough and Fitch are relatively thin, shallow-marine deposits; the Fitch has been interpreted as a far-foreland deposit (Bradley, 1983). The overlying Littleton Formation is an approximately 1.6-km-thick flysch succession (a minimum thickness, because the top is not preserved) deposited in a foreland-basin setting (Bradley, 1983). It has yielded Emsian brachiopods at two localities in the Littleton area (loc. 9, figs. 1, 4) and at a third locality in Whitefield (loc. 10, figs. 1, 4), where the fossils occur only a few tens of meters above the base (Boucot and Arndt, 1960, p. 41–43). At Gale River (loc. 39, fig. 1) in the next strike belt to the east, a tuff from low in the Littleton Formation has yielded an Emsian U-Pb zircon age of 407±2 Ma (R.D. Tucker and D. Rankin, unpub. data, 1998). Still farther east, at Beaver Brook (loc. 8, fig. 1), a brachiopod fauna suggests a slightly older age (“Oriskany, possibly Esopus”; Boucot and Rumble, 1980, p. 192–194) than elsewhere (locs. 9, 10, fig. 1). The Littleton thus appears to be diachronous across strike at the latitude of its type area.

Southwestern Part of the Moose River Synclinorium, Maine

In the southwestern part of the Moose River synclinorium, western interior Maine, local carbonate-bearing units overlie Ordovician and older rocks, providing good age constraints on the onset of Acadian foreland-basin sedimentation. At Beck Pond (loc. 12, figs. 1, 4), the post-Ordovician section begins with the Beck Pond Limestone (Boucot and others, 1959) and includes reef limestone intercalated with boulder conglomerate derived from the nearby Attean pluton of Ordovician age. A rich brachiopod fauna initially suggested an “upper Helderberg, New Scotland” age (Boucot and Heath, 1969); however, conodonts from the upper part of the Beck Pond (member 5 of Boucot and others, 1959) include *Belodella* sp., *Decoriconus* sp., *Dvorakia* sp., *Icriodus* sp. indet., and *Ozarkodina remscheidensis eosteinhornensis*, indicating, instead, a very latest Pridoli to early Lochkovian age (pl. 1, figs. 1–13; table 2). The Beck Pond Limestone is overlain by a considerable thickness, probably 2 to 3 km, of Lower Devonian flysch assigned to the Seboomook Group (formerly Formation; Pollock, 1987). Boucot and others (1959) interpreted the contact between Member 5 of the Beck Pond and overlying slate of the Seboomook as an angular unconformity. Upon digging out this contact (at loc. 10 of Boucot and others, 1959), we observed, instead, that reefal limestone debris interfingers with black silty shale. The conodonts thus date the top of the Beck Pond Limestone, the base of the Seboomook flysch sequence, and the onset of Acadian foreland-basin sedimentation.

About 1 km to the north (loc. 12, figs. 1, 4), the Bear Pond Limestone (Boucot and others, 1959) is a carbonate lens or buildup not far above the base of the Seboomook Group. It is known from a single outcrop that exposes about 10 m of section. A rich brachiopod fauna again indicates an “upper Helderberg, New Scotland” age (Boucot and Heath, 1969) for the buildup and, by implication, for the immediately overlying and underlying siltstone and sandstone of the Seboomook Group. A large conodont sample yielded a meager conodont fauna of *Belodella devonica*, *Ozarkodina remscheidensis eosteinhornensis*, and fragments of *Icriodus* sp. indet. or *Pedavis* sp. indet., allowing a possible age range from late Ludlow to earliest Pragian (table 2). Nearby, Boucot and Heath (1969, p. 36) reported several brachiopod occurrences that suggest a Pragian age for all but the lowest beds of the Seboomook Group in the Moose River synclinorium. The Seboomook Group is the youngest pre-Acadian stratified unit in the area; it is not known to extend upward beyond the Pragian.

About 15 km to the northeast, the post-Ordovician section (loc. 13, figs. 1, 4) begins with the McKenney Ponds Limestone Member of the Tarratine Formation. Boucot and Heath (1969) reported a “Becraft-Oriskany” (late Lochkovian to Pragian) age based on brachiopods. This assignment can be somewhat refined on the basis of our conodont collections (table 2). The most diagnostic of three samples (USGS colln. 12516–SD, table 2), located 2 m stratigraphically below siltstone of the Tarratine Formation, yielded *Belodella devonica*, *Icriodus* sp. indet., *Ozarkodina remscheidensis*, and *Pseudooneotodus beckmanni* (pl. 1, figs. 19–26). A Lochkovian, but not earliest Lochkovian, age is indicated by the icriodids, which are of a post-*I. woschmidtii* morphotype. The limestone member is gradationally overlain by sandstone of the main body of the Tarratine Formation, which both interfingers with and overlies the Seboomook Group and likewise is interpreted as part of the foreland-basin fill. The Tarratine Formation has yielded Oriskany-age (Pragian) brachiopods from several nearby localities (Boucot and Heath, 1969).

Northeastern Part of the Moose River Synclinorium, Maine

The northeastern part of the Moose River synclinorium in Maine provides crucial stratigraphic constraints on the age of the youngest foreland-basin deposits. In the area of northern Moosehead and Brassua Lakes (loc. 15, figs. 1, 4), Ordovician and older basement rocks are overlain by poorly dated Silurian calcareous rocks and Devonian(?) red shale (Boucot and Heath, 1969). The Acadian foreland-basin succession begins with the Tarratine Formation, which is about 2 km thick in this area (Boucot and Heath, 1969, pl. 19). The Tarratine Formation has yielded abundant Oriskany-age—that is, Pragian—brachiopods (Boucot and Heath, 1969, p. 27). The Kineo Rhyolite, sandwiched between the Tarratine Formation, below, and the Tomhegan Formation, above, represents a pulse of silicic magmatism in the foreland basin. The Kineo Rhyolite has yielded only discordant zircon ages, marked by inheritance. It is comparable in stratigraphic position, composition, and eruptive environment to the 407- to 406-Ma Traveler Rhyolite (Rankin and Tucker, 1995) 65 km to the northeast (loc. 17, figs. 1, 4); the Kineo and Traveler are presumably coeval. The Kineo ranges in thickness from 0 m at its north limit to about 1,200 m just a few kilometers to the south. It is overlain by the main body of the Tomhegan Formation, consisting of crossbedded, probably deltaic sandstone (L. Hanson and D. Bradley, unpub. data). Boucot and Heath (1969, p. 17) estimated the thickness at about 1,800 m—a minimum, because the top is not preserved. Brachiopods in the Tomhegan indicate a "Schoharie" (late Emsian; Boucot and Heath, 1969, p. 20) age. The Tomhegan is the youngest pre-Acadian unit in this part of Maine, indicating that the Acadian deformation here was no earlier than the late Emsian.

Traveler-Chesuncook Area, Maine

Along the west side of the Katahdin batholith (loc. 16, figs. 1, 4), Silurian sedimentation began with deposition of shallow marine conglomerate, limestone, and siltstone assigned by Griscom (1976) to the Ripogenus Formation. Various localities within the Ripogenus Formation have yielded brachiopods of Llandovery (C3–C5), late Wenlock, and possible Ludlow age (Boucot and Heath, 1969, p. 53). A conodont sample (95MDW130) from Ripogenus Dam (loc. 16, figs. 1, 4; table 2) yielded *Decoriconus* sp. indet., *Dvorakia* sp. indet., *Oulodus* sp. indet., *Ozarkodina excavata*, and *Panderodus* sp. indet. (pl. 2, figs. 1–5); this collection permits an age range of late Ludlow to early Lochkovian. The Ripoge-

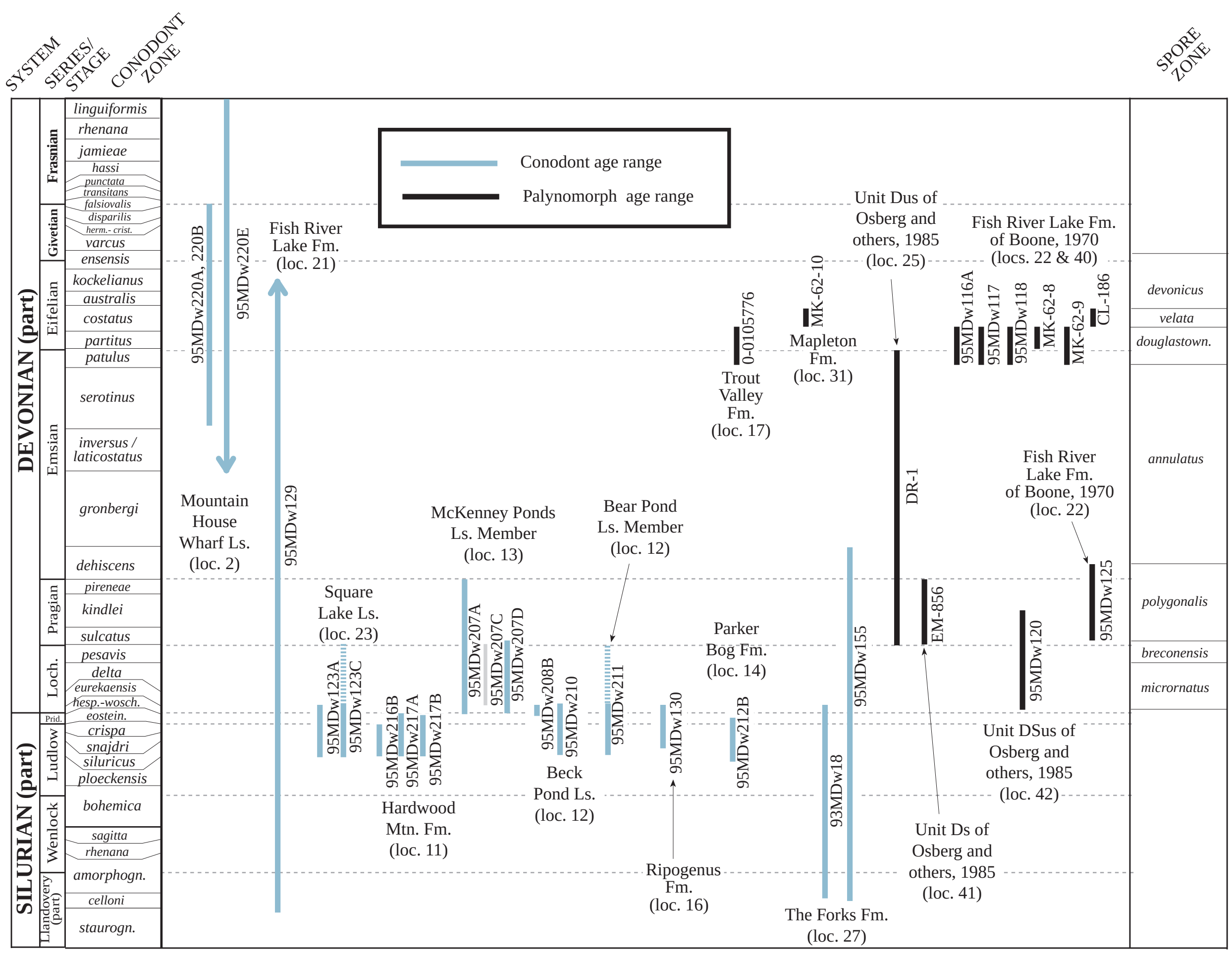


Figure 5.—Correlation chart showing conodont age ranges for samples dated in this report, and palynomorph age ranges for samples dated in this report and for samples from northern Maine that were previously discussed in internal paleontologic reports of the Geological Survey of Canada (McGregor, 1962, 1963, 1968, 1992).

Table 2. Conodont collections from Silurian and Devonian rocks, Maine and Québec.

[See plates 1 and 2 for illustrations of key conodonts; see figure 5 for the Silurian and Devonian conodont zonation and a graphical display of the age range of each sample. CAI, conodont-alteration index; USGS, U.S. Geological Survey]

Loc. no. in fig. 1	Field no. (USGS colln. no.)	County quadrangle lat./long.	Formation, lithology, and locality description	Conodonts	Age	Biofacies	CAI	Remarks
23	95MDw123A (12511–SD)	Aroostook Square Lake West 47°03'52"/ 68°22'57"	Square Lake Limestone—prominent lakeshore outcrop of biostromal limestone containing abundant brachiopods, corals, crinoids, and bryozoans.	145 *Belodella* spp. including *B.* cf. *B. resima* (Philip) (pl. 2, figs. 27–32) 21 *Decoriconus fragilis* (Branson and Mehl) (pl. 1, figs. 22–26) 4 *Dvorakia* sp. (pl. 2, figs. 36, 37) *Ozarkodina excavata* (Branson and Mehl)? 1 Sa and 1 Sb element fragments *Ozarkodina remscheidensis remscheidensis* (Ziegler) 10 Pa, 5 Pb, 2 M, and 2 Sc elements (chiefly fragments) 9 *Pseudooneotodus beckmanni* (Bischoff and Sannemann) (pl. 2, fig. 38) 42 indet. bar, blade, platform and coniform fragments	late Ludlow–early Lochkovian (*siluricus* Zone to at least *woschmidti* Zone).	Belodellid biofacies: this collection, with an abundance of coniform elements, is typical of reefoid deposits in the Silurian and earliest Devonian.	1.5	8.7 kg processed; 4.02 kg +20 mesh and 220 g 20–200 mesh insoluble residue.
23	95MDw123C (12512–SD)	Aroostook Square Lake West 47°03'52"/ 68°22'57"	Square Lake Limestone—grainstone, from low outcrops at water line about 50 m west and stratigraphically above 95MDw123A.	1 Pb *Oulodus* sp. indet. *Ozarkodina remscheidensis* (Ziegler) 8 Pa (*O. r. remscheidensis*), 1 Pa (*O. r. eosteinhornensis*), 1 Pb, 1 Sa, and 1 Sc elements (chiefly fragments) 15 indet. bar, blade, and platform fragments	late Ludlow through earliest Pragian (*siluricus* Zone into lowermost *sulcatus* Zone	Indeterminate—too few conodonts. Conodonts indicate high-energy, normal-marine depositional setting.	1.5	3.8 kg processed; 400 g +20 mesh and 62 g 20–200 mesh insoluble residue.
21	95MDw129	Aroostook Fish River Lake 46°50'52"/ 68°46'02"	Fish River Lake Formation—silty limestone containing crinoid and stromatoporoid debris.	2 Sc element fragments of post-Ordovician Paleozoic morphotype several phosphatic bryozoan pearls	Middle Ordovician–Mississippian on the basis of the bryozoan pearls.	Indeterminate—too few conodonts.	3	5.2 kg processed; 3.3 kg +20 mesh and 466 g 20–200 mesh insoluble residue.
16	95MDw130 (12513-SD)	Piscataquis Harrington Lake 45°52'55"/ 69°10'38"	Ripogenus Formation—limestone. Sample is from 2-m-thick limestone with shaly partings separated from an overlying 5-m-thick conglomerate by 1 m of calcareous sandstone. At 90° bend in Frost Pond Rd.	2 *Decoriconus* sp. indet. 6 *Dvorakia* sp. indet. (pl. 2, figs. 1, 2) 1 Pb element *Oulodus* sp. indet. *Ozarkodina excavata* (Branson and Mehl) 1 Pa, 1 M, 1 Sa, 6 Sb, and 4 Sc elements (pl. 2, figs. 3, 4) 59 *Panderodus* sp. indet. 1 Sb element of a digyrate apparatus (pl. 2, fig. 5) 34 indet. bar, blade, and platform fragments	late Ludlow–early Lochkovian (*siluricus* Zone–earliest Early Devonian)	Panderodid biofacies: normal-marine, probably relatively shallow-water depositional environment.	chiefly 5.5; minor 5 and 6	11.5 kg processed; 4.7 kg +20 mesh and 195 g 20–200 mesh insoluble residue.
27	95MDw155 (12514–SD)	Somerset The Forks 45°19'58"/ 69°58'05"	The Forks Formation—calcareous sandstone to sandy limestone. Roadcut on E side of US 201, 0.3 mi S of Kennebec River bridge; sample from prominent 20-cm-thick, white calcareous sandstone. Stratigraphically higher than 95MDw180.	*Ozarkodina excavata* (Branson and Mehl) 5 Pa, 1 Pb, 1 M, and 1 Sc elements 46 indet. bar, blade, and platform fragments	late Llandovery–early Emsian.	Indeterminate—too few conodonts. Normal-marine, relatively high-energy depositional setting.	6.5	12.1 kg processed; 4.9 kg +20 mesh and 1.2 kg 20–200 mesh insoluble residue.
27	93MDw18 (12334–SD)	Somerset The Forks 45°18.3'/ 69°59.2'	The Forks Formation from roadcut on E side of US 201, 4.9 mi N of Caratunk. Recessive calcareous horizon 75 ft. N of telephone pole no. 234,445,131.	3 *Decoriconus* sp. indet. *Ozarkodina excavata* (Branson and Mehl) 5 Pa, 2 M, and 1 Sb elements 2 *Panderodus* sp. indet. 33 indet. bar, blade, and platform fragments Phosphatized steinkerns: 1 gastropod and 1 pelmatozoan ossicle	late Llandovery–early Lochkovian.	Indeterminate—too few conodonts. Preservation and composition of fauna suggest a high-energy, normal-marine setting.	5.5–6.5	9.9 kg processed; 4.02 kg +20 mesh and 737 g 20–200 mesh insoluble residue.

Table 2. Conodont collections from Silurian and Devonian rocks, Maine and Québec—Continued.

Loc. no. in fig. 1	Field no. (USGS colln. no.)	County quadrangle lat./long.	Formation, lithology, and locality description	Conodonts	Age	Biofacies	CAI	Remarks
13	95MDw207A (12515–SD)	Somerset Enchanted Pond 45°27'15"/ 70°12'23"	McKenney Ponds Limestone Member of Tarratine Formation. Sample from middle of 3 caves below outlet of pond and ~5 m below lowest exposure of Tarratine Formation siltstone.	14 *Belodella devonica* (Stauffer) 8 *Dvorakia* sp. indet. 34 I element fragments of *Icriodus* sp. indet. of Lochkovian–Pragian morphotype (pl. 2, figs. 16–18) 3 *Pseudooneotodus beckmanni* (Bischoff and Sannemann)	Lochkovian–Pragian.	Post mortem transport within or from the icriodid biofacies; relatively high-energy normal-marine depositional setting.	5	11.4 kg processed; 200 g +20 mesh and 96 g 20–200 mesh insoluble residue. Heavy-mineral concentrate: chiefly phosphatic brachiopod fragments and euhedral pink zircons.
13	95MDw207C (12516–SD)	Somerset Enchanted Pond 45°27'15"/ 70°12'23"	McKenney Ponds Limestone Member of Tarratine Formation. Sample is from the southwesterly of the 3 caves and from an irregularly bedded bioclastic limestone, 3.5 m stratigraphically above 95MDw207D.	37 *Belodella devonica* (Stauffer) 67 I (all fragments) and 5 coniform elements of *Icriodus* sp. indet. of Lochkovian, but not early Lochkovian morphotype (not *I. woschmidti*) (pl. 1, figs. 19–24) 2 Pa elements *Ozarkodina remscheidensis* (Ziegler) (pl. 1, figs. 25, 26) 2 *Pseudooneotodus beckmanni* (Bischoff and Sannemann)	Lochkovian, but not early Lochkovian.		5	19.5 kg processed; 20 g +20 mesh and 77 g 20–200 mesh insoluble residue. Heavy-mineral concentrate includes phosphatic brachiopod fragments.
13	95MDw207D (12517–SD)	Somerset Enchanted Pond 45°27'15"/ 70°12'23"	McKenney Ponds Limestone Member of Tarratine Formation. Sample is from the southwesterly of the 3 caves and from an irregularly bedded bioclastic limestone, about 3.5 m below 95MDw207C, and from approximately the same level as 95MDw207A.	14 *Belodella devonica* (Stauffer) coniform elements 1 small P element fragment *Icriodus* sp. indet. *Ozarkodina remscheidensis* (Ziegler) 23 Pa and 1 Pb elements 1 *Pseudooneotodus beckmanni* (Bischoff and Sannemann) 32 indet. bar, blade, and platform fragments	Lochkovian.	Ozarkodinid biofacies: normal-marine, shelf depositional environment	5	10.7 kg processed; 180 g +20 mesh and 29 g 20–200 mesh insoluble residue. Heavy-mineral concentrate includes phosphatic brachiopod fragments.
12	95MDw208B (12518–SD)	Somerset King and Bartlett Lake 45°22'05"/ 70°20'18"	Beck Pond Limestone. Carbonate matrix conglomerate with clasts of granite and reef debris. This is outcrop 10 of Boucot and others (1959). This sample is from the youngest member (Boucot and others' Member 5) of the Beck Pond.	110 *Belodella* spp. (Stauffer) (pl. 1, fig. 13) 14 *Decoriconus* sp. 2 *Dvorakia* sp. (pl. 1, figs. 7, 8) 6 coniform elements *Icriodus* sp. indet. (pl. 1, figs. 9, 10) *Ozarkodina remscheidensis eosteinhornensis* (Walliser) 75 Pa, 17 Pb, 16 M, 1 Sa, 3 Sb, and 6 Sc elements (pl. 1, figs. 1–6) 75 indet. bar, blade, and platform fragments	Very latest Pridoli–early Lochkovian (based on overlap of icriodids and *Decoriconus*).	Ozarkodinid biofacies: normal-marine, shallow-water shelf depositional environment.	4.5–5	41.2 kg processed; 7.6 kg +20 mesh and 1.7 kg 20–200 mesh insoluble residue. Heavy-mineral concentrate includes phosphatic zooecial linings and pearls of bryozoans, detrital muscovite, biotite, chlorite, and euhedral zircon.
12	95MDw210 (12519–SD)	Somerset King and Bartlett Lake 45°22'11"/ 70°20'41"	Beck Pond Limestone. Sandy limestone containing abundant bryozoans and corals. This is the same locale as outcrop 32 of Boucot and others (1959) and is from the oldest member (Boucot and others' Member 1) of the Beck Pond.	5 Pa elements *Ozarkodina remscheidensis eosteinhornensis* (Walliser)	late Ludlow–early Lochkovian (*siluricus* Zone to at least *woschmidti* Zone).	Indeterminate—too few conodonts.	4.5–5	8.3 kg processed; 340 g +20 mesh and 208 g 20–200 mesh insoluble residue.

Table 2. Conodont collections from Silurian and Devonian rocks, Maine and Québec—Continued.

Loc. no. in fig. 1	Field no. (USGS colln. no.)	County quadrangle lat./long.	Formation, lithology, and locality description	Conodonts	Age	Biofacies	CAI	Remarks
12	95MDw211 (12520–SD)	Somerset Spencer Lake 45°22'40" 70°20'34"	Bear Pond Limestone Member of Seboomook Group (formerly Seboomook Formation). This is outcrop 42 of Boucot and others (1959).	1 *Belodella devonica* (Stauffer) 1 Pa element *Ozarkodina remscheidensis eosteinhornensis* (Walliser) 3 P element fragments *Icriodus* sp. indet. or *Pedavis* sp. indet.	late Ludlow–earliest Pragian (no older than *siluricus* Zone)	Indeterminate—too few conodonts.	5	11.1 kg processed; 1.43 kg +20 mesh and 2.2 kg 20–200 mesh insoluble residue. Heavy-mineral concentrate includes phosphatic zooecial linings of bryozoans and phosphatic tubes, and rare euhedral zircon.
14	95MDw212B (12521–SD)	Somerset King and Bartlett Lake ~45°18.7'/ ~70°19.6'	Parker Bog Formation—limestone interbedded with siliceous siltstone or tuff. From south bank of Spencer Stream; detailed outcrop map in Boucot and others (1958). From the thickest carbonate bed ~7m downstream from a prominent point.	*Oulodus elegans* (Walliser) (pl. 2, figs. 6–9) 1 Pa, 1 Pb, 1 M, 1 Sb, and 1 Sc *Ozarkodina remscheidensis* (Ziegler) subsp. indet. 29 Pa, 5 Pb, and 2 M element fragments 2 *Panderodus* sp. indet. 4 *Pedavis* sp. indet. coniform elements (pl. 2, figs. 10, 11) 85 indet. bar, blade, and platform fragments	late Ludlow into late Pridoli (from at least *snajdri* Zone into late Pridolian).	Ozarkodinid biofacies; normal-marine, shelf depositional environment. Condition of conodonts indicates relatively high-energy depositional regime.	Chiefly 5.5 and minor 5 and 6	8.5 kg processed; 1.56 kg +20 mesh and 973 g 20–200 mesh insoluble residue. Heavy-mineral concentrate: chiefly ferruginous siltstone and detrital mica.
11	95MDw216B (12523–SD)	Somerset Jackman 45°38'32"/ 70°20'50"	Hardwood Mountain Formation from outlet stream of Sugar Bertg Pond, near Little Big Wood Pond. Rubble crop of coral-bearing limestone.	*Corryssognathus dubius* (Rhodes) (pl. 2, figs. 13–16) 1 Sb and 3 coniform elements *Oulodus* sp. indet. 1 Pa, 4 Pb, 3 Sb, and 1 Sc elements *Ozarkodina excavata excavata* (Branson and Mehl) (pl. 2, fig. 12) 5 Pa and 1 M elements *Ozarkodina remscheidensis remscheidensis* (Ziegler) (pl. 2, fig. 17–20) 8 Pa, 2 M, 1 Sb, and 3 Sc elements 1 *Panderodus* sp. indet. UNASSIGNED ELEMENTS: 1 Pb, 1 M, and 1 Sa 87 indet. bar, blade, and platform fragments	late Ludlow (late Gorstian through Ludfordian). *Corryssognathus dubius* is the most biostratigraphically diagnostic species and is restricted to the late Ludlow.	Ozarkodinid-oulodid; relatively shallow-water, moderate- to high-energy shelf depositional setting.	5	16.6 kg processed; 100 g +20 mesh and 126 g 20–200 mesh insoluble residue. Heavy-mineral concentrate includes rare subrounded to euhedral zircon.
11	95MDw217A (12524–SD)	Somerset Jackman 45°38'35"/ 70°20'58"	Hardwood Mountain Formation from Sugar Berth Pond. Shaly limestone, 5 m above 95MDw216B and youngest sample collected at 95MDw216 or 95MDw217.	2 Sb elements *Oulodus* sp. indet. 2 Pa elements *Ozarkodina confluens* Branson and Mehl (pl. 2, fig. 21) *Ozarkodina remscheidensis remscheidensis* (Ziegler) 8 Pa, 1 Pb, and 1 Sb robust elements 2 *Panderodus* sp. indet. 27 indet. bar, blade, and platform fragments	late Ludlow–Pridoli.	Indeterminate—too few conodonts. Normal-marine, relatively high energy depositional environment.	5	24.0 kg processed; 3.97 kg +20 mesh and 220 g 20–200 mesh insoluble residue. Heavy-mineral concentrate: chiefly ferruginous micaceous and argillaceous flakes and common phosphatic bryozoan pearls.
11	95MDw217B (12525–SD)	Somerset Jackman 45°38'35"/ 70°20'58"	Hardwood Mountain Formation near Little Big Wood Pond. Shaly limestone 10 m below 95MDw216B.	*Oulodus* sp. indet. 1 Pa, 1 Pb, 1 M, 1 Sb, and 1 Sc elements *Ozarkodina remscheidensis remscheidensis* (Ziegler) 12 Pa, 5 Pb, and 1 M elements (mostly fragments) UNASSIGNED ELEMENTS: 1 Pb, 1 Sb, and 1 Sc 21 indet. bar, blade, and platform fragments	late Ludlow–Pridoli; if 95MDW217B is indeed stratigraph-ically below 216B, then the age of 217B is late Ludlow.	Ozarkodinid; normal-marine shelf depositional setting.	5	16.9 kg processed; 2.36 kg +20 mesh and 202 g 20–200 mesh insoluble residue.

Table 2. Conodont collections from Silurian and Devonian rocks, Maine and Québec—Continued.

Loc. no. in fig. 1	Field no. (USGS colln. no.)	County quadrangle lat./long.	Formation, lithology, and locality description	Conodonts	Age	Biofacies	CAI	Remarks
2	95MDw220A (12526–SD)	Quebec, Canada Lake Memphremagog 45°03'40"/ 72°16'50"	Mountain House Wharf Limestone. Calcareous slate and limestone. At N end of Chemin Mountain. 95MDw220A is from lakeshore cliffs, lowest horizon at extreme L end of outcrop.	20 deformed I elements *Icriodus* sp. indet. of late Emsian–Givetian morphotype (short-platform icriodids) (pl. 1, figs. 30, 31) Icriodids of this type are speciated on the basis of platform-margin outline.	late Emsian–Givetian.	Post mortem transport within or from the icriodid biofacies. Specimens originated in a relatively high energy reefal or biohermal environment.	upper 5	11.1 kg processed; 8.72 kg +20 mesh and 570 g 20–200 mesh insoluble residue. Heavy-mineral concentrate includes semischist fragments.
2	95MDw220B (12527–SD)	Quebec, Canada Lake Memphremagog 45°03'40"/ 72°16'50"	Mountain House Wharf Limestone. From 30 m below old dam and about 120 m up from lake along stream that flows past tennis courts—from farthest downstream horizon exposed.	4 deformed I elements *Icriodus* sp. indet. of late Emsian–Givetian morphotype (short-platform icriodids) (pl. 1, fig. 29)	late Emsian–Givetian (latest Early–Middle Devonian)	Indeterminate—too few conodonts. Icriodids indicate derivation from relatively high energy reefal or biohermal environment.		10.2 kg processed; 2.12 kg +20 mesh and 320 g 20–200 mesh insoluble residue. Heavy-mineral concentrate includes semischist fragments.
2	95MDw220E (12528–SD)	Quebec, Canada Lake Memphremagog 45°03'40"/ 72°16'50"	Mountain House Wharf Limestone. About 8 m structurally above and 30 m right of 95MDw220A.	2 I element fragments *Icriodus* sp. indet.	Devonian	Indeterminate—too few conodonts.		7.1 kg processed; 4.58 kg +20 mesh and 124 g 20–200 mesh insoluble residue.

nus is gradationally overlain by siltstone and andesite (West Branch Formation), followed by red shale (Frost Pond Shale); these units are barren of fossils but are probably Lochkovian. The overlying Seboomook Group, which begins the Acadian-derived clastic succession, consists of thin-bedded turbidites that have yielded Oriskany-age (Pragian) brachiopods from a locality at Chesuncook Lake (R.B. Neumann, in Griscom, 1976, p. 147–148). The Seboomook Group is overlain progradationally by the Matamagon Sandstone, which has also yielded Pragian brachiopods (for example, R.B. Neumann, in Griscom, 1976, p. 151–152). North of the Katahdin pluton (loc. 17, figs. 1, 4), the Seboomook and Matagamon together compose a prodelta and delta deposited by west-directed paleocurrents (Hall and others, 1976; Pollock and others, 1988). The Matagamon Sandstone is overlain conformably by the Traveler Rhyolite. There appears to have been no hiatus between the two units because clastic dikes of Matagamon Sandstone intrude upward into the basal rhyolite (Rankin, 1968). Concordant U-Pb zircon ages of 406 Ma from the Black Cat Member and 407 Ma from the Pogy Member (Rankin and Tucker, 1995) indicate an early Emsian age for the Traveler Rhyolite and, most likely as well, for the uppermost beds of the Matagamon Sandstone. The Silurian through upper Emsian strata were involved in the regional Acadian folding, which produced a regional northeast-striking cleavage.

The Traveler Rhyolite is overlain by essentially undeformed nonmarine sandstone and conglomerate of the Trout Valley Formation. Although an angular unconformity cannot be demonstrated at outcrop, 1:62,500-scale mapping shows that the Trout Valley Formation truncates major structures in the Traveler (Rankin, 1968). Kasper and others (1988) reviewed the history of age assignments for the Trout Valley on the basis of plant fossils. Palynomorphs from the Trout Valley (McGregor, 1992) indicate a latest Emsian or earliest Eifelian age, corresponding to the *douglastownense-eurypterota* spore assemblage zone or the *patulus* and *partitus* conodont zones (table 1). The Trout Valley is, therefore, significantly younger than the Traveler Rhyolite.

Munsungun Anticlinorium, Maine

The Munsungun anticlinorium is cored by Ordovician and older rocks, which are unconformably overlain by a complex assortment of locally derived clastic, volcanic, and carbonate rocks (loc. 18, figs. 1, 4). Hall (1970) assigned these rocks to the East Branch Group and to the laterally equivalent Spider Lake Formation. Corals and brachiopods from several fossil occurrences give ages of Ludlow, Pridoli, and "middle or upper Helderberg" (Hall, 1970, p. 35). Next follows the Seboomook Group, the youngest unit in the area, which has yielded brachiopods of Oriskany-age (Pragian) age from three places (loc. 18, figs. 1, 4; Hall, 1970, p. 41). Thus, Acadian deformation must be Emsian or younger along the Munsungun anticlinorium.

A similar succession has been reported from the Millimagassett Lake area, at the south end of the Munsungun anticlinorium (loc. 20, figs. 1, 4). Here, the Grand Lake Seboeis

Group consists of slate, sandstone, conglomerate, calcareous slate, and volcanic rocks. Hibbard (1994, p. 216) cited nine fossil collections that give a range of "Late Silurian to earliest Devonian." (No details were published). The Grand Lake Seboeis Group grades up into, but may interfinger with, the Millimagassett Lake Formation, which has yielded Late Silurian, most likely Pridoli, brachiopods and corals near its base (Hibbard, 1993); presumably, its upper part extends at least into the Lochkovian, if not the Pragian. According to Hibbard (1994), the Millimagassett Lake Formation is a flysch succession resembling the Seboomook Group. It appears to be the oldest such flysch yet identified along the anticlinorial belt, possibly 2 to 3 m.y. older than the Seboomook Group at Beck Pond (loc. 12, figs. 1, 4) and 5 to 10 m.y. older than along the Munsungun anticlinorium (loc. 18, figs. 1, 4). In the Millimagassett Lake area, Hibbard (1994) reported evidence for an early Acadian sinistral shear deformation that was followed by the main, regional Acadian folding on northeast-trending axes. Both phases of deformation are Lochkovian or younger.

Pennington Anticlinorium, Maine

To determine the timing of Acadian foreland-basin sedimentation and deformation along the northwestern limb of the Pennington anticlinorium (locs. 21–23, figs. 1, 4), we must first unravel some stratigraphic problems. The Silurian to Devonian Fish River Lake Formation of Boone (1970) consists of sandstone, siltstone, conglomerate, mafic and silicic volcanic rocks, and impure limestone. Unnamed unit DSus, as shown on the bedrock geologic map of Maine (Osberg and others, 1985), occurs along strike from the Fish River Lake Formation at the same broad stratigraphic level. New information reported here suggests that both the Fish River Lake and unit DSus are made up of two unrelated successions: an older one that includes volcanic and calcareous rocks, and a younger one composed of plant-bearing siltstone, sandstone, and conglomerate.

Carbonate rocks from the Fish River Lake Formation have yielded brachiopods of Late Silurian ("Cobleskill," Pridoli) and Early Devonian ("New Scotland," Lochkovian) age (A.J. Boucot, in Boone, 1970, p. 33–34). We processed a shaly limestone from one of Boucot's Cobleskill-age locations (loc. 21, figs. 1, 4) for conodonts, but the sample yielded only nondiagnostic Silurian-Devonian forms (table 2). Within unit DSus, the informally named Square Lake Limestone occurs in a single lakeshore exposure (loc. 23, figs. 1, 4). A.J. Boucot (in Boone, 1970) placed the brachiopod and coral fauna "definitely within the lower Helderberg series" (that is, *woschmidti* Zone to at least *eurekaensis* Zone of the Lochkovian). Conodonts from the most diagnostic of two samples of reef limestone give an age range of late Ludlow to early Lochkovian (pl. 2, figs. 22–32, 36–38; loc. 23, figs. 1, 4; table 2).

Plant-bearing sandstone and siltstone from the Fish River Lake Formation of Boone (1970) and unit DSus have yielded spores, some of which are considerably younger than the brachiopods (table 1; McGregor, 1962, 1963, 1968, 1996) This conclusion is based on material submitted (without lithologic descriptions) to D.C. McGregor by Eli Mencher in the 1960's and by D.C. Bradley in 1995. In unit DSus, the most closely dated of two siltstone samples is clearly Early Devonian, probably Lochkovian or early Pragian (table 1). Bedding and a strong cleavage (presumably Acadian) in these rocks both are subvertical. In the Fish River Lake Formation, two samples of dark, thin-bedded siltstone are of late Pragian-early Emsian age (*polygonalis-emsiensis* Zone). These rocks, which resemble the Seboomook Group, are also folded and cleaved. Three sandstone samples from 1995 and two samples of unknown lithology from the 1960's are of late Emsian or early Eifelian age (*douglastownense-eurypterota* Zone). The 1995 samples of this age are from a succession of fluvial or marginal-marine plant-bearing sandstone and conglomerate at Nadeau Thoroughfare (loc. 22, figs. 1, 4), which dip moderately (50°) but lack cleavage. Finally, one sample of unknown lithology submitted to McGregor by Mencher is early or middle Eifelian (*velata-langii* Zone).

Piecing together these various facts, we suggest that far-foreland carbonate sedimentation was underway during the Late Silurian and at least part of the Lochkovian. An influx of orogenically derived siliciclastic materials began during late Lochkovian or early Pragian time. Deposition of the plant-bearing sandstone probably lasted from about late Emsian into the early or middle Eifelian. Taken together, the palyniferous strata can be interpreted as components of a prograding foreland-basin sequence, akin to the somewhat older Seboomook-Matagamon succession farther south.

Madawaska Area, New Brunswick

Two useful constraints on the date of the Acadian orogeny are available from northwesternmost New Brunswick, where the Aroostook-Matapedia and Connecticut Valley-Gaspé Basins come together (loc. 24, figs. 1, 4). The Upper Ordovician and Silurian are represented by various deep-water facies (St. Peter, 1977). The youngest Silurian rocks are mostly green slate and calcareous siltstone of the Perham Formation, which has yielded graptolites as young as middle to late Ludlow (St. Peter, 1977, p. 30). There follows a gap in the fossil record spanning the Pridoli, Lochkovian, and Pragian. The overlying Temiscouata Formation consists of several kilometers of siliciclastic turbidites characterized by abundant slump folds; Hesse and Dalton (1995) interpreted the Temiscouata as a submarine foreland-basin succession. It has yielded brachiopods now assigned an Emsian age (St. Peter and Boucot, 1981, p. 91). The Temiscouata Formation has been penetratively deformed; Acadian deformation must be Emsian or younger in age.

LaPatrie and Ste.-Cécile de Whitton, Eastern Townships, Québec

In the Eastern Townships of Québec, the axial region of the Connecticut Valley-Gaspé Basin is occupied by a thick succession of Acadian-deformed sandstone and slate assigned to the Compton Formation (Slivitski and St.-Julien, 1987). Sandstone from the Compton Formation has yielded Emsian plant fossils (from locs. 3, 4, figs. 1, 4; Hueber and others, 1990). Although the age of onset of foreland-basin sedimentation in this area is unknown, the presence of Emsian foreland-basin deposits suggests that the orogen was not far off at the time. Acadian deformation in the Eastern Townships must have taken place between 409–394 Ma (possible age range of the Compton Formation) and the 384±2-Ma Scotstown pluton, which is the oldest postfolding pluton that intrudes the Compton Formation (Simonetti and Doig, 1990).

Northwestern Margin of the Connecticut Valley-Gaspé Basin, Québec

Three localities along the northwestern margin of the Connecticut Valley-Gaspé Basin in Québec provide important constraints on the age of Acadian foreland-basin sedimentation and deformation. The most informative strata are at St.-Georges (loc. 5, figs. 1, 4), where the Eifelian Famine Formation unconformably overlies the Ordovician Magog Group (Uyeno and Lespérance, 1997). The Famine begins with a basal conglomerate, overlain by limestone, which is coral bearing at the bottom but becomes progressively shalier upsection. The maximum preserved thickness is 215 m. Uyeno and Lesperance (1997) documented a conodont fauna of *costatus* Zone age, a refinement on the previously quoted Eifelian age (Oliver, 1971), which was based on corals and brachiopods. The Famine Formation is the youngest unit at St.-Georges; it was involved in regional Acadian deformation, which thus must postdate the *costatus* Zone.

In the Lac Temiscouata area (loc. 7, figs. 1, 4), the situation is similar to that at St.-Georges. The key unit, comparable to the Famine, is the Touladi Limestone, which rests with slight angular unconformity on various older units ranging in age from possibly Ordovician to Late Silurian (Uyeno and Lespérance, 1997). The Touladi consists of mostly bioclastic limestone, with a maximum preserved thickness of 105 m. Recent conodont studies by Uyeno and Lespérance (1997) indicate a position in either the *australis* or *kockelianus* Zone of the Eifelian, an assignment that refines but does not significantly alter the Eifelian age quoted by Boucot and Drapeau (1968).

Devonian limestone at Lac Memphremagog (loc. 2, fig. 1) suggests a similar interpretation, although biostratigraphic control is not so tight. There, the Mountain House Wharf Limestone is a dark, slaty limestone surrounded by, and probably faulted against, Ordovician rocks (Boucot and Drapeau, 1968). It has yielded a brachiopod and coral fauna of Eifelian age (Boucot and Drapeau, 1968, p. 9). Three of four large conodont samples yielded only poorly preserved forms that could be assigned only a middle Emsian to Givetian age (table 2).

The coral-bearing shallow-water limestone at St.-Georges, Lac Temiscouata, and Lac Memphremagog was probably deposited in a far-foreland setting. The upward-deepening trend at St.-Georges most likely records the beginnings of foreland-basin subsidence. Reckoning that an interval of foreland-basin sedimentation would have preceded deformation, the Acadian orogeny most likely took place at St.-Georges during the Givetian or, conceivably, the late Eifelian. Deformation at Lac Temiscouata probably took place a few million years later than at St.-Georges.

Montreal, Québec

Devonian strata do not crop out anywhere north of the Mohawk River in central New York but were once present in Montreal, 330 km to the north (loc. 1, figs. 1, 4). The Ste. Hélène Island breccia (Boucot and others, 1986) is a Cretaceous diatreme that was emplaced into a region of gentle foreland deformation about 20 km cratonward of the Paleozoic thrust front. The diatreme contains an assortment of xenoliths that reveal the former existence of a Devonian succession similar to that in eastern and central New York State. This diatreme, in the words of J.B. Thompson, is a “drill hole in the sky.” The inferred sequence is Lochkovian (Helderbergian) limestone, Pragian (Oriskany-age) sandstone and limestone, Emsian (Bois Blanc or Schoharie age) limestone, Eifelian (early and late Onondagan) limestone, and Givetian (Hamilton-age) micritic limestone, siltstone, and sandstone. The latter siltstone and sandstone resemble the Hamilton, which in the Catskills of eastern New York State constitutes the base of the main foreland-basin sequence. By implication, the leading edge of the foreland basin was near Montreal during Givetian time.

Isotopic Ages of Acadian Plutons

When the present study was first conceived, only a few high-precision U-Pb and $^{40}Ar/^{39}Ar$ ages were available for Silurian and Devonian plutons in the study area (fig. 1)—not enough to reveal a clear regional age pattern for the plutons. Accordingly, we set out to date all the significant Acadian plutons in a swath from the Maine coast to the Maine-Québec border. New U-Pb and (or) $^{40}Ar/^{39}Ar$ ages are reported here for 17 plutons. Other new ages from the geochronology laboratory of Washington University, St. Louis, were reported by Tucker and others (in press) for plutons near and on the Maine coast, by Rankin and Tucker (1995) for the Katahdin area, by Solar and others (1998) for western interior Maine, and by Eusden and others (in press) for northern New Hampshire. Figure 8 and table 3, which summarize the best available emplacement ages, clearly reveal plutonic belts of about 420 Ma near the coast and 407–406 Ma farther inland, which are essential in positioning the deformation front. Other plutons are significantly younger than about 406 Ma, some of which also occur in discrete belts, but they are not so useful in tracking the Acadian front.

Acadian plutons fall into three broad and somewhat overlapping categories, pretectonic, syntectonic, and posttectonic, with respect to the local first phase (D1) of Acadian deformation. A pluton might be classified as posttectonic, even though the Acadian orogeny was still underway elsewhere during the pluton’s emplacement. We considered three types of anecdotal evidence in assessing the age of plutonism relative to Acadian deformation: (1) map-scale relations to Acadian folds and faults, (2) plutonic-rock fabrics, and (3) metamorphic textures in the contact aureole.

U-Pb Analytical Methods

U-Pb analyses (fig. 7; table 4) were performed at the geochronology laboratory of Washington University, St. Louis. Zircon was extracted from 5- to 10-kg samples, using standard techniques of density and magnetic separation. Zircon grains were selected for analysis on the basis of size, color, clarity, and morphology. Sample sizes varied, depending on U content, grain size, and age, but most analyses were performed on fractions of 5 to 20 crystals, generally with less than 200 ng of radiogenic Pb. All analyses were air abraded (Krogh, 1982) and cleaned sequentially in warm 4*N* HNO_3, water, and distilled acetone to remove components carrying common Pb. Washed and weighed zircon fractions were then loaded in Teflon bombs, spiked with a mixed ^{205}Pb-^{235}U tracer solution, and digested in 48 percent HF and 7*N* HNO_3 for 72 hours at 210°C. After digestion and conversion to chloride form, Pb and U were purified by using ion-exchange techniques, as described in Krogh (1973). Total-procedure blanks for Pb and U, measured during the period of analysis, were 1–8 and less than 1 pg, respectively; total common-Pb abundances are reported for each analysis in table 4. Initial-Pb compositions result in insignificant changes to the calculated ages.

Isotopic-ratio measurements of Pb and U were made in a VG Sector 54 thermal-ionization mass spectrometer with enhanced pumping capacity, seven movable collectors, and a Daly-type detector with ion-counting capability. Lead and uranium were loaded together on outgassed single-Re filaments with silica gel and phosphoric acid, and all measurements were made by the method of peak hopping in ion-counting mode. Ion-beam intensities ranged from 0.5×10^{-13} to 1.5×10^{-13} A for $^{206}Pb^+$ and from 0.5×10^{-13} to 2.5×10^{-13} A for ^{235}U (measured as UO_2^+). Daly bias and nonlinearity were periodically monitored with U.S. National Institute of Standards & Technology and Central Bureau for Nuclear Measurements isotopic reference materials, and correction factors and errors for Daly gain were used in data reduction.

Errors for the $^{238}U/^{206}Pb$, $^{235}U/^{207}Pb$, and $^{207}Pb/^{206}Pb$ ages were estimated by using the method of Ludwig (1980); all age uncertainties are quoted at the 95-percent-confidence level. Cited ages are the mean $^{207}Pb/^{206}Pb$ age of concordant or slightly discordant analyses weighted according to the inverse variance of each analysis (Ludwig, 1992); the quoted age error is the standard error of the average value calculated by using the assigned error for each analysis. The reliability of all cited ages may be evaluated by the mean square of weighted deviates (MSWD), which is a measure of the observed scatter to that predicted by the assigned errors to each analysis. For all samples, the MSWD is much less than 1, indicating that assigned errors may be overestimated.

$^{40}Ar/^{39}Ar$ Analytical Methods

$^{40}Ar/^{39}Ar$ analyses (fig. 8; table 5) were performed at the geochronology laboratory at the University of Maine, Orono. Samples were separated by using standard magnetic and density separation techniques. The purity of the samples was estimated to be greater than 99.5 percent. Samples, flux monitors (interlaboratory standard SBG–7), and K and Ca salts were encapsulated in Al foil and sealed in silica glass vials. These vials were irradiated in the L67 position of the Ford Nuclear Reactor at the University of Michigan. Micas and flux monitors weighed approximately 35 mg. Samples were heated in a molybdenum crucible within the ultra-high-vacuum system on line to the mass spectrometer using radio-frequency induction. Temperatures have an estimated uncertainty of ±50°C. Inert gases were purified by using standard gettering techniques. The Ar-isotopic composition was measured digitally, using a Nuclide 6–60–SGA 1.25 mass spectrometer. All data were corrected for mass discrimination and interfering Ar isotopes produced during irradiation (Dalrymple and others, 1981). The decay constants recommended by Steiger and Jäger (1977) were used to calculate the ages. Error calculations, which included both the uncertainty in the analytical measurement and the uncertainty in the *J* value, are reported at the 2σ level. Each sample was analyzed in a minimum of 8 and no more than 15 increments. A plateau age represents the mean of ages in consecutive increments that do not differ on the basis of 2σ analytical uncertainties.

Ludlow, Pridoli, and Lochkovian Plutons

Until a few years ago, the numerous gabbroic to granitic plutons along the Maine coast were all assigned to the Devonian, but in light of modern $^{40}Ar/^{39}Ar$ and U-Pb dating and the new time scale, the oldest group of plutons now is known to range from Ludlow to Lochkovian. The first pluton in Maine to yield an undoubted Silurian age was the Pocomoonshine gabbrodiorite (PO, fig. 6), which yielded an $^{40}Ar/^{39}Ar$ amphibole plateau age of 422.7±3 Ma (West and others, 1992). The pluton is posttectonic: it intruded already-deformed turbidites of the presumably Lower Silurian Flume Ridge Formation and “stitches” the fault contact between the Fredericton Basin and the St. Croix belt to the south (West and others, 1992). New zircon ages from the Utopia (423.0±1.0 Ma; UT, fig. 6) and Welsford (422.0±1.0 Ma; WL, fig. 6) plutons (M.L. Bevier, written commun., 1996) extend the belt of known Ludlow plutons into southern New Brunswick. Several plutons in the Penobscot Bay area have also yielded concordant zircon ages of 424 to 417 Ma—that is, Late Silurian to earliest Devonian—including the Somesville (424.0±2.0 Ma; SV, fig. 6), Blinn Hill (424.0±2.0 Ma; BH, fig. 6), Lake St. George (422.0±2.0 Ma; LS, fig. 6), North Union (422.0±2.0 Ma; NU, fig. 6), Spruce Head (421.0±1.0 Ma; SH, fig. 6), Youngtown (420.0±2.0 Ma; YT, fig. 6), Sedgwick (419.5±1.0 Ma; SE, fig. 6), South Penobscot (419.2±2.2 Ma; SP, fig. 6), Cadillac (419.0±2.0 Ma; CD, fig. 6), Northport (419.0±2.0 Ma; NP, fig. 6), and Lincoln (417.7±1.0 Ma; LI, fig. 6) (table 3; Seaman and others, 1995; Stewart and others, 1995b; Tucker and others, in press) plutons. The North Union and Lake St. George plutons are composed of granitic gneiss inferred to have been emplaced syntectonically (Tucker and others, in press). West and others (1995) reported widespread Late Silurian to earliest Devonian $^{40}Ar/^{39}Ar$ ages from metamorphic hornblende outboard of the Sennebec Pond Fault, essentially coeval with the plutons listed above.

Early Emsian Plutons

Previous Results

Nearly two dozen closely dated plutons of Emsian age define a northeast-southwest belt that can be traced across the entire strike length of figure 6 and beyond, into Massachusetts. Previously, there was some indication of plutonic activity at about 400 to 410 Ma along this trend, but little

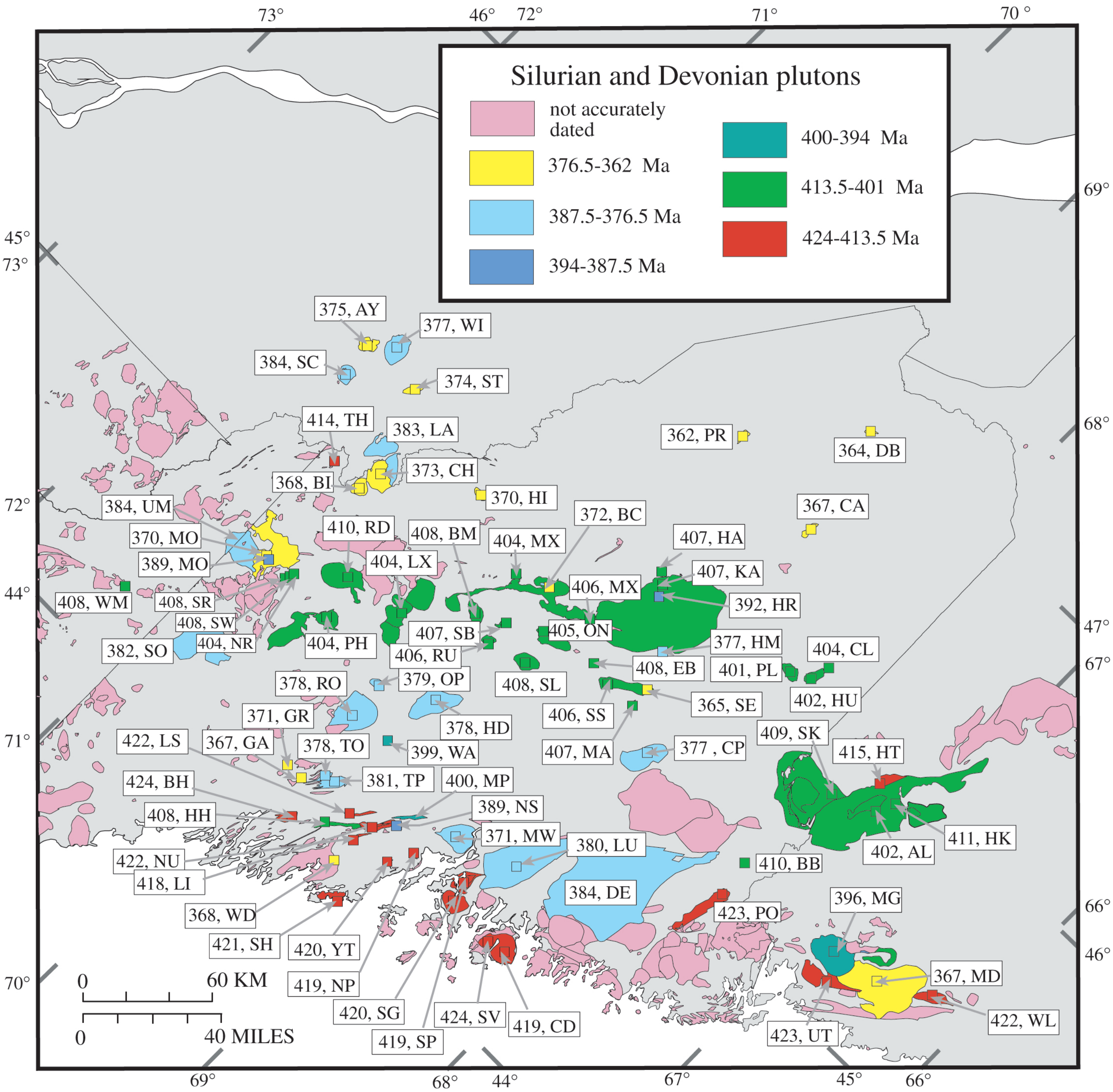


Figure 6.—Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing distribution of Late Silurian, Devonian, and earliest Carboniferous plutons. Boxes show best estimate of intrusive age (in Ma; concordant U-Pb or $^{40}Ar/^{39}Ar$ plateau) and two-letter abbreviation, keyed to table 3.

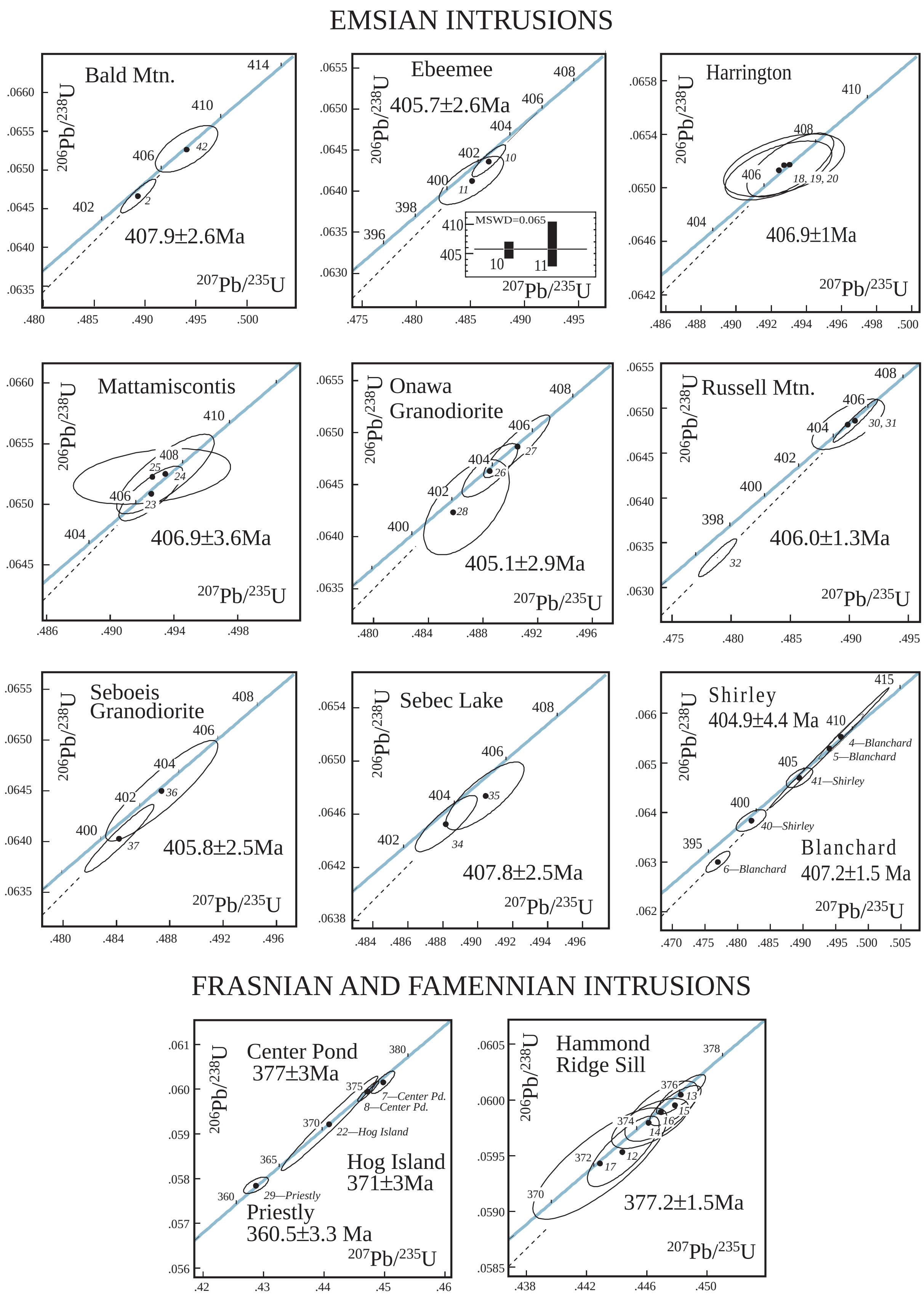


Figure 7.—U-Pb concordia diagrams of intrusive rocks dated in this report.

evidence to suggest that most plutons along the belt would turn out to fall within the same narrow age range. In western New Brunswick, Bevier and Whalen (1990a) reported U-Pb ages of 409±2 Ma for the Skiff Lake phase of the Pokiok batholith and 402±1 Ma for the Allandale phase (SK and AL, respectively, fig. 6). In eastern Maine, Hubacher and Lux (1987) obtained $^{40}Ar/^{39}Ar$ ages, which are now known to be Emsian, from three plutons that truncate regional Acadian structures: the Pleasant Lake (400.5±4.5 Ma; PL, fig. 6), Hunt Ridge (401.5±4 Ma; HU, fig. 6), and Cochrane Lake (404.3±3.7 Ma; CL, fig. 6) plutons. In western Maine, Solar and others (1998) recently reported Emsian U-Pb ages from the South Roxbury granite (408.2±2.5 Ma; SR, fig. 6), Swift River granite (407.9±1.9 Ma; SW, fig. 6), Redington pluton (407.6±4.6 Ma; RD, fig. 6), North Roxbury granite (404.3±1.9 Ma; NR, fig. 6), Lexington pluton (404.3±1.8 Ma; LX, fig. 6), and various phases of the Phillips pluton (405.3±1.8, 403.6±2.2, and 403.5±1.6 Ma; PH, fig. 6). In the Presidential Range of northern New Hampshire, the Wamsutta pluton (WM, fig. 6), which was emplaced during F1 Acadian folding, has yielded a U-Pb zircon age of 408.4±1.9 Ma (Eusden and others, in press). In New Hampshire, outside the area of figure 6, Emsian U-Pb ages have been obtained for the Spaulding pluton (408.4±1.9 Ma on zircon; Robinson and Tucker, 1996), the Ashuelot pluton (403±3 Ma on monazite; Robinson and Tucker, 1996), and the Cardigan pluton (404±2 Ma on zircon; R.D. Tucker, unpub. data). Finally, the Prescott gabbro, which postdates the nappe phase of regional Acadian

Figure 8.—$^{40}Ar/^{39}Ar$ age spectra of intrusive rocks dated in this report. Shaded rectangles correspond to data used to calculate plateau ages.

deformation in Massachusetts, has a U-Pb age of 407±3/2 Ma (Tucker and Robinson, 1990). Far outboard of the main belt of Emsian magmatism are two outliers: the Haskell Hill pluton (U-Pb age, 408±5 Ma; HH, fig. 6; Tucker and others, in press) and the Berry Brook gabbro-diorite (^{40}Ar/^{39}Ar age, 410 Ma; BB, fig. 6; Ludman and Idleman, 1998).

Russell Mountain Pluton

The Russell Mountain pluton (RU, fig. 6) is a unfoliated two-mica granodiorite. Mapping by Ludman (1978) showed that the pluton truncates regional folds in the Madrid and Carrabassett Formations, and so, on this basis, it must postdate at least some Acadian deformation. Contact-metamorphic rocks on the northwestern margin, however, have a schistose foliation and aligned chiastolite with biotite strain shadows (fig. 9*A*). Because metamorphic biotite is incipient or absent outside the aureole (Ludman, 1978), some deformation must have taken place while the pluton was being emplaced. Whether deformation in the contact aureole was caused by pluton emplacement or was a regional event is unknown. Two concordant zircon fractions yield a mean ^{207}Pb/^{206}Pb age of 406.0±1.3 Ma (fig. 7). Two biotite fractions yielded ^{40}Ar/^{39}Ar plateau ages of 393.6±2.7 and 393.4±3.0 Ma (fig. 8).

Sebec Lake Pluton

The Sebec Lake pluton (SL, fig. 6) is a unfoliated biotite granodiorite that, like the Russell Mountain pluton, truncates regional folds involving the Madrid and Carrabassett Formations (Griffin, 1971). On the basis of its map-scale relations, it must postdate at least some Acadian deformation. Contact-metamorphic textures, however, are synkinematic. Metamorphosed Carrabassett Formation in the aureole contains large andalusite porphyroblasts with asymmetric strain shadows made up of biotite and chlorite aligned in the foliation (fig. 9*B*). Outside the contact aureole, the rocks are at subbiotite grade; thus, the biotite and andalusite are both related to plutonism. Whether deformation in the contact aureole was caused by pluton emplacement or was a regional event is unknown. We obtained two concordant zircon fractions that yielded a mean ^{207}Pb/^{206}Pb age of 407.8±2.5 Ma (fig. 7) and ^{40}Ar/^{39}Ar plateau ages of 392.5±2.5 Ma from biotite and 377.2±4.4 Ma from muscovite (fig. 8).

Bald Mountain Pluton

The Bald Mountain pluton (BM, fig. 6) is a unfoliated two-mica granodiorite; in map view, it has the shape of an ellipse with its long axis normal to regional strike. Mapping by Espenshade and Boudette (1967) and Ludman (1978) showed that the pluton truncates regional folds in the Madrid and Carrabassett Formations and on this basis would appear to postdate at least some Acadian deformation. Unoriented chiastolite and biotite porphyroblasts in steeply dipping contact-metamorphosed strata also suggest posttectonic emplacement (fig. 9*C*). A single concordant zircon fraction indicates an emplacement age near 408 Ma (table 4). ^{40}Ar/^{39}Ar ages were obtained from three biotite separates, two of which yielded plateaus of 390.9±3.0 and 394.3±3.7 Ma, giving a mean age of 392.6±2 Ma for cooling through the closure temperature of argon in biotite (fig. 8).

Onawa Pluton

The Onawa pluton (ON, fig. 6) is a zoned body with a granodiorite core and a gabbroic rim. It truncates and deflects the regional structural grain defined by steeply dipping bedding in the Carrabassett Formation. Thin-section observations reported by Van Heteren and Kusky (1994) suggest that andalusite in the contact aureole overgrew a preexisting cleavage; thus, the pluton is posttectonic, at least in part. We obtained three concordant zircon fractions that yielded a mean ^{207}Pb/^{206}Pb age of 405.1±2.9 Ma for the granodiorite (fig. 9), and an ^{40}Ar/^{39}Ar plateau age of 400.1±3.7 Ma on biotite from the gabbro (fig. 8).

Shirley-Blanchard Pluton

The Shirley-Blanchard pluton (SB, fig. 6) is a composite pluton that includes a more mafic phase of pyroxene-hornblende granodiorite (Blanchard body) and a more felsic phase of granodiorite (Shirley body; Espenshade and Boudette, 1967). The pluton truncates regional-scale tight to isoclinal folds involving the Madrid and Carrabassett Formations. From the Shirley body, we obtained two concordant zircon analyses (fig. 7) with a mean ^{207}Pb/^{206}Pb age of 404.9±4.4 Ma and an ^{40}Ar/^{39}Ar plateau age of 390.3±3.0 Ma on biotite (fig. 10). From the Blanchard body, we obtained two concordant zircon fractions (fig. 7) that yielded a mean ^{207}Pb/^{206}Pb age of 407.2±1.5 Ma. The weighted mean ^{207}Pb/^{206}Pb age of all zircon analyses from the Shirley-Blanchard body is 406.9±1.4 Ma (fig. 7), our preferred age for this composite pluton.

Mattamiscontis Pluton

The Mattamiscontis pluton (MA, fig. 6) is an unfoliated biotite granite; fieldwork for this study was limited to geochronologic sampling of the pluton and brief examination of outcrops in the contact aureole. Reconnaissance mapping compiled by Osberg and others (1985) showed the pluton cutting a syncline cored by Carrabassett Formation; similarly, at one outcrop in the contact aureole, tightly folded metapelite displays a weak axial-planar cleavage that is overgrown by cordierite. A thin section from another outcrop, however, shows oriented cordierite surrounded by biotite strain shadows. (The regional metamorphic grade is chlorite.) These observations together suggest that the pluton postdates Acadian folding, but that some deformation—either regional or pluton related—accompanied pluton emplacement. A granite sample yielded three concordant zircon analyses with a mean ^{207}Pb/^{206}Pb age of 406.9±3.6 Ma (fig. 7) and an ^{40}Ar/^{39}Ar biotite plateau age of 402.1±2.8 Ma (fig. 8).

Table 3. Emplacement ages of plutons mapped in figure 6.

[State or province: ME, Maine; NB, New Brunswick; NH, New Hampshire; QU, Québec;]

Pluton	State or province	Abbrev. (fig. 7)	Age (Ma)	Error (Ma)	Method	Mineral	Comments	Reference
Allandale pluton, Pokiok batholith	NB	AL	402.0	1	U-Pb	Monazite	Mean $^{235}U/^{207}Pb$ age; 2 concordant fractions.	Bevier and Whalen (1990a).
Aylmer pluton	QU	AY	375.0	3	U-Pb	Monazite	$^{238}U/^{206}Pb$ age of 1 nearly concordant fraction.	Simonetti and Doig (1990).
Bald Mountain pluton	ME	BM	408.0	– –	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	This report.
Beaver Cove pluton	ME	BC	372.4	2.7	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	Do.
Berry Brook gabbro-diorite	ME	BB	410.0	– –	$^{40}Ar/^{39}Ar$	Hornblende	– – –	Ludman and Idleman (1998).
Big Island Pond pluton	ME	BI	367.7	1.3	$^{40}Ar/^{39}Ar$	Hornblende	Average of 2 plateau ages: 367.7±2.1 and 367.6±2.1 Ma.	Heizler and others (1988).
Blinn Hill granite gneiss	ME	BH	424.0	2	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Tucker and others (in press).
Cadillac pluton	ME	CD	419.0	2	U-Pb	Zircon	No details given.	Seaman and others (1995).
Center Pond pluton	ME	CP	377.0	3	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant fractions.	This report.
Chain of Ponds pluton	ME, QU	CH	373.3	2	$^{40}Ar/^{39}Ar$	Hornblende	Average of 6 plateau ages: 371.6, 371.1, 375.4, 373.2, 372.5, and 375.9 Ma (all ±2.1 Ma).	Heitzler and others (1988).
Chandler pluton	ME	CA	366.5	3.7	$^{40}Ar/^{39}Ar$	Biotite	Total-gas age; almost a plateau age.	This report.
Cochrane Lake pluton	ME	CL	404.3	3.7	$^{40}Ar/^{39}Ar$	Hornblende	Average of 3 plateau ages: 401±4, 405±5, and 407±7 Ma.	Hubacher and Lux (1987).
Deblois pluton, northeastern arm	ME	DE	384.0	5	U-Pb	Zircon	No details given.	A. Ludman (written commun., 1998); date by T. Lanzirotti.
Deboullie pluton, granodiorite	ME	DB	364.2	3.2	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	This report.
Deboullie pluton, monzonite	ME	DB	363.9	– –	$^{40}Ar/^{39}Ar$	Biotite	Average of two plateau ages: 362.3±3 and 365.4±3.2 Ma.	Do.
Ebeemee pluton	ME	EB	407.8	2.4	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	Do.
Ebeemee pluton	ME	EB	405.7	2.6	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant zircon fractions.	Do.
Hammond Ridge sill	ME	HM	377.2	2.5	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 6 concordant fractions.	Do.
Harkshaw pluton, Pokiok batholith	NB	HK	411.0	1	U-Pb	Sphene	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Bevier and Whalen (1990a).
Harrington pluton	ME	HA	406.9	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant zircon fractions.	This report.
Hartfield pluton, Pokiok batholith	NB	HT	415.0	1	U-Pb	Sphene	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Bevier and Whalen (1990a).
Hartland pluton	ME	HD	378.9	1.3	U-Pb	Zircon and sphene	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	This report.
Haskell Hill pluton	ME	HH	408.0	+5/–4	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Tucker and others (in press).
Hog Island pluton	ME	HI	370.0	– –	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	This report.
Horserace quartz diorite	ME	HR	392.0	– –	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Tucker (unpub. data).
Hunt Ridge pluton	ME	HU	401.5	4	$^{40}Ar/^{39}Ar$	Clinopyroxene and hornblende	Average of 2 plateau ages: 401±4 and 402±5 Ma.	Hubacher and Lux (1987).
Katahdin pluton	ME	KA	407.0	.4	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 5 concordant zircon fractions.	Rankin and Tucker (1995).
Lac aux Araignées (Spider Lake) pluton	QU	LA	383.0	3	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Simonetti and Doig (1990).
Lake Saint George granite gneiss	ME	LS	422.0	2	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Tucker and others (in press).
Lexington pluton	ME	LX	404.2	1.8	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Solar and others (1998).
Lincoln shonkinite	ME	LI	417.7	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	Tucker and others (in press).
Lucerne pluton	ME	LU	380.0	4	U-Pb	Zircon	No details given.	Zartman and Gallego (1979).
Magaguadavic pluton	NB	MG	396.0	1	U-Pb	Zircon	No details given.	M.L. Bevier (written commun., 1996).
Mattamiscontis pluton	ME	MA	406.9	3.6	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant zircon fractions.	This report.
Mixer Pond granite gneiss	ME	MP	400.0	3	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant fractions.	Tucker and others (in press).
Mooselook-Meguntic two-mica leucogranite	ME	MO	370.3	1.1	U-Pb	Monazite	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant fractions.	Solar and others (1998).
Mooselook-Meguntic pluton, biotite granite	ME	MO	388.9	1.6	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Do.
Mount Waldo pluton	ME	MW	371.0	1.9	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	Stewart and others (1995b).
Moxie pluton, central part	ME	MX	404.4	3.4	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	This report.
Moxie pluton, eastern part	ME	MX	406.3	3.8	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	Do.
Mount Douglas pluton	NB	MD	366.5	1	U-Pb	Monazite	Mean of two U-Pb ages.	M.L. Bevier (written commun., 1996).
North Roxbury two-mica granite	ME	NR	404.3	1.9	U-Pb	Zircon and monazite	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Solar and others (1998).
North Searsmont granite gneiss	ME	NS	389.0	2	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	Tucker and others (in press).
North Union granitic gneiss	ME	NU	422.0	2	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$; 2 concordant zircon fractions.	Do.

Table 3. Emplacement ages of plutons mapped in figure 6—Continued.

Pluton	State or province	Abbrev. (fig. 7)	Age (Ma)	Error (Ma)	Method	Mineral	Comments	Reference
Northport pluton	ME	NP	419.0	2	U-Pb	Zircon	No details given	D.B. Stewart (written commun., 1996); date by R.D. Tucker.
Onawa pluton, gabbro	ME	ON	401.1	3.7	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	This report.
Onawa pluton, granodiorite	ME	ON	405.1	2.9	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant zircon fractions.	Do.
Pegmatite, Gardiner	ME	GA	367.0	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Tucker and others (in press).
Pegmatite, Greeley Corner	ME	GR	371.0	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant fractions.	Do.
Phillips pluton	ME	PH	403.6	2.2	U-Pb	Zircon and monazite	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Solar and others (1998).
Phillips pluton schlieritic granite	ME	PH	405.3	1.8	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Do.
Phillips pluton, leucogranite	ME	PH	403.5	1.6	U-Pb	Zircon and monazite	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Do.
Pleasant Lake pluton	ME	PL	401.3	4.7	$^{40}Ar/^{39}Ar$	Biotite	Average of 3 plateau ages: 404±5, 397±9, and 403±7 Ma.	Hubacher and Lux (1987).
Pleasant Lake pluton	ME	PL	400.5	4.5	$^{40}Ar/^{39}Ar$	Hornblende	Average of 2 plateau ages: 400±5 and 401±4 Ma.	Do.
Pocomoonshine gabbro-diorite	ME	PO	422.7	3.0	$^{40}Ar/^{39}Ar$	Hornblende	Average of two plateau ages: 423.5±2.7 and 421.5±3.0 Ma. Dates intrusion of more northerly mafic phase.	West and others (1992).
Priestly pluton	ME	PR	361.7	– –	$^{40}Ar/^{39}Ar$	Biotite	Average of two plateau ages: 360.3±2.78 and 363.2±2.3 Ma.	This report.
Priestly pluton	ME	PR	361.0	3	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Do.
Redington biotite granite	ME	RD	407.6	4.7	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Solar and others (1998).
Rome pluton	ME	RO	378.0	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Tucker and others (in press).
Russell Mountain pluton	ME	RU	406.0	1.3	U-Pb	Zircon	No details given.	This report.
Scotstown pluton	QU	SC	384.0	2	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age of 2 slightly discordant fractions.	Simonetti and Doig (1990).
Sebec Lake pluton	ME	SL	407.8	2.5	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant zircon fractions.	This report.
Seboeis granodiorite at East Branch Lake	ME	SE	364.7	4.3	$^{40}Ar/^{39}Ar$	Biotite	Plateau age.	Do.
Seboeis pluton, granodiorite	ME	SS	405.8	2.5	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$; 1 concordant zircon fraction.	Do.
Sedgwick pluton	ME	SG	419.5	1.4	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Stewart and others (1995b).
Shirley-Blanchard composite pluton	ME	SB	406.9	1.4	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$; 4 concordant zircon fractions.	This report.
Skiff Lake pluton, Pokiok batholith	NB	SK	409.0	2	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age of most concordant fraction.	Bevier and Whalen (1990a).
Somesville pluton	ME	SV	424.0	2	U-Pb	Zircon	No details given.	Seaman and others (1995).
Songo pluton	ME	SO	382.0	3	U-Pb	Zircon and sphene	Mean $^{207}Pb/^{206}Pb$ age; 2 slightly discordant fractions.	Lux and Aleinikoff (1985), Lux and others (1989).
South Penobscot pluton	ME	SP	419.2	2.2	U-Pb	Zircon	Supersedes date of Stewart and others (1995).	D.B. Stewart (written commun., 1996); date by R.D. Tucker.
South Roxbury two-mica granite	ME	SR	408.2	2.5	U-Pb	Zircon and monazite	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	Solar and others (1998).
Spruce Head granite	ME	SH	421.0	1	U-Pb	Zircon and sphene	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	Tucker and others (in press).
Ste.-Cécile pluton	QU	ST	374.0	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant or nearly concordant fractions.	Simonetti and Doig (1990).
Swift River two-mica leucogranite	ME	SW	407.9	1.9	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Solar and others (1998).
Thrasher Peaks intrusions	ME	TH	414.0	– –	U-Pb	Zircon	$^{207}Pb/^{206}Pb$ age; 1 concordant fraction.	Moench and others (1995).
Threemile Pond pluton	ME	TP	381.0	1	U-Pb	Zircon and sphene	Mean $^{207}Pb/^{206}Pb$ age; 4 concordant fractions.	Tucker and others (in press).
Togus pluton	ME	TO	378.0	1	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant fractions.	Do.
Umbagog pluton	ME	UM	384.0	6	U-Pb	Zircon	No details given.	Aleinikoff and Moench (1987).
Utopia pluton	NB	UT	422.7	1	U-Pb	Zircon	No details given.	M.L. Bevier (written commun., 1996).
Waldoboro binary granite	ME	WD	367.8	1.6	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Tucker and others (in press).
Wamsutta pluton	NH	WM	408.0	1.9	U-Pb	Zircon	No details given.	J.D. Eusden, Jr. (written commun., 1998).
Waterville granitic dike	ME	WA	399.4	.7	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 2 concordant fractions. Intrudes first-generation Acadian folds but is involved in second-generation folds.	Tucker and others (in press).
Welsford pluton	NB	WL	422.0	1	U-Pb	Zircon	No details given.	M.L. Bevier (written commun., 1996).
Winslow pluton	QU	WI	377.0	6	U-Pb	Sphene	Mean $^{207}Pb/^{206}Pb$ age; 2 nearly concordant fractions.	Simonetti and Doig (1990).
Youngtown pluton	ME	YT	420.0	2	U-Pb	Zircon	Mean $^{207}Pb/^{206}Pb$ age; 3 concordant fractions.	Tucker and others (in press).

Table 4. U-Pb isotope-dilution analyses of zircon from plutonic rocks, Maine.

[Brackets, field No. (U.S. Geological Survey colln. No.)]

Fractions			Concentrations			Atomic Rations					Age
No. in fig. 9	Properties	Wt. [µg]	Pb rad [ppm]	U [ppm]	Pb com [pg]	Th/U	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$
	(1)	(2)	(2)	(2)	(3)	(4)	(5)	(6)	(6)	(6)	(6)
Bald Mountain [95MDw140]											
†1	20 gr,-200,cl,c,p	32	24.9	401	2.9	0.249	18,028	0.05555±3	0.4878±6	0.06369±7	434.3±1.1
†2	6 gr,-200,cl,c,t-p	18	37.9	385	3.5	0.223	11,979	0.07231±8	0.9540±22	0.09570±24	994.6±2.3
3	3 gr,-200,cl,c,n	11	30.1	482	5.0	0.213	4,497	0.05489±4	0.4893±7	0.06466±9	407.9±1.4
42	2 gr, -200, cl, cl, n	9	9.15	129	1.9	0.634	2,555	0.05490±11	0.4941±12	0.06527±12	408.2±4.3
Blanchard [94MDw98]											
4	4 gr,-200,cl,pb,s-p	14	431	6630	7.12	0.306	57,749	0.05488±4	0.4958±14	0.06553±19	407.2±1.4
5	10 gr,-200,cl,cr,sk	29	62.3	879	2.15	0.641	48,679	0.05487±3	0.4940±37	0.06530±49	406.9±1.0
6	7 gr,-200,cl,c,n	11	32.1	509	2.81	0.331	7,653	0.05490±5	0.4769±7	0.06301±9	408.3±2.0
Center Pond [95MDw157]											
7	2 gr,-200,cl,c,t-p	5	31.6	511	1.4	0.431	7,219	0.05423±5	0.4497±8	0.06015±10	380.7±2.2
8	21 gr,-200,cl,c,t-p	40	27.1	434	5.7	0.485	11,804	0.05411±3	0.4473±7	0.05995±9	375.8±1.2
Debouille [95MDw115]											
†9	10 gr,-200,cl,c,s-p	28	33.2	451	3.7	1.146	13,092	0.05463±4	0.4515±8	0.05993±11	397.3±1.5
Ebeemee [95MDw167]											
10	15 gr,-200,cl,c,n	40	24.5	392	17.5	0.232	3,736	0.05484±3	0.4866±6	0.06437±8	405.6±1.4
11	5 gr,-200,cl,c,n	13	28.5	452	11.0	0.270	2,294	0.05486±9	0.4850±12	0.06413±12	406.6±3.8
Hammond Ridge Sill [95MDw184]											
12	2 gr,-200,cl,pb,p	3	157	2710	10.8	0.243	2,918	0.05414±8	0.4444±9	0.05954±13	376.8±3.2
13	2 gr,-200,cl,pb,p	3	238	4065	21.7	0.257	2,200	0.05414±4	0.4483±6	0.06005±7	377.0±1.8
14	4 gr,-200,cl,pb,fp	6	139	2406	57.1	0.236	913.3	0.05413±10	0.4462±10	0.05979±9	376.4±4.0
15	2 gr,-200,cl,pb,p	3	138	2340	12.2	0.277	2,239	0.05419±5	0.4479±7	0.05995±7	378.9±2.3
16	2 gr,-200,cl,pb,p	2	224	3800	17.9	0.284	1,657	0.05412±9	0.4469±10	0.05990±11	376.1±3.7
17	2 gr,-200,cl,pb,p	3	141	2375	19.6	0.333	1,416	0.05405±13	0.4429±18	0.05943±20	373.0±5.5
Harrington [94MDw48]											
18	7 gr, -200,cl,c,l-p	21	8.90	125	2.05	0.661	5,461	0.05488±8	0.4931±10	0.06517±10	407.3±3.4
19	4 gr, -200,cl,c,l-p	14	7.86	111	2.85	0.646	2,324	0.05484±14	0.4927±14	0.06517±9	405.6±5.5
20	5 gr, -200,cl,c,l-p	18	6.94	98.5	1.85	0.626	1,302	0.05483±12	0.4924±12	0.06513±9	405.3±4.7
†21	5 gr, -200,cl,c,l-p	16	10.0	141	2.98	0.615	3,312	0.05525±10	0.4990±12	0.06550±11	422.4±4.1
Hog Island [95MDw205A]											
22	5 gr,-200,cl,c,lp	14	36.2	567	13.7	0.619	2,142	0.05399±7	0.4409±32	0.5923±43	370.6±3.0
Mattamiscontis [95MDw164A]											
23	10 gr,-200,cl,c,t-p	17	15.4	236	2.2	0.355	7,530	0.05488±6	0.4925±8	0.06509±9	407.4±2.3
24	8 gr,-200,cl,c,t-p	10	18.8	285	3.4	0.366	3,677	0.05485±8	0.4935±13	0.06525±13	406.3±3.3
25	2 gr,-200,cl,c,t-p	6	19.2	292	13.1	0.367	557.6	0.05478±21	0.4926±20	0.06523±9	403.1±8.6
Onawa [95MDw68]											
26	4 gr,-200,cl,c,cr,fr	13	28.7	400	2.2	0.723	9,468	0.05481±6	0.4885±8	0.06464±10	404.4±2.4
27	10 gr,-200,cl,cr,c,fr	21	28.7	396	5.3	0.708	6,433	0.05484±5	0.4905±9	0.06487±12	405.6±2.0
28	1 gr,-200,cl,cr,fr	3	26.2	364	2.3	0.771	2,180	0.05485±4	0.4858±13	0.06424±19	405.9±5.7
Priestly [95MDw108]											
29	6 gr,-200,cl,c,p	14	18.2	289	5.6	0.646	2,716	0.05375±8	0.4287±8	0.05785±7	360.5±3.3

Table 4. U-Pb isotope-dilution analyses of zircon from plutonic rocks, Maine—Continued.

Fractions			Concentrations			Atomic Rations					Age
No. in fig. 9	Properties	Wt. [µg]	Pb rad [ppm]	U [ppm]	Pb com [pg]	Th/U	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{235}U$	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$
	(1)	(2)	(2)	(2)	(3)	(4)	(5)	(6)	(6)	(6)	(6)
						Russell Mountain [95MDw98]					
30	20 gr,-200,cl,c,cr,n	43	36.4	578	6.9	0.228	15,074	0.05485±2	0.4905±7	0.06486±10	406.2±0.9
31	3 gr,-200,cl,c,cr,n	5	30.3	469	3.0	0.319	3,013	0.05481±10	0.4899±13	0.06482±11	404.6±4.2
32	15 gr,-200,cl,cr,p	30	54.1	770	4.9	0.723	19,181	0.05484±3	0.4788±6	0.06333±9	405.7±1.1
†33	22 gr,-200,cl,cr,c,n	34	35.7	572	10.7	0.242	7,417	0.05510±9	0.4873±18	0.06414±25	416.2±3.5
						Sebec Lake [95MDw71]					
34	20 gr,-200,cl,c,n	26	29.5	473	12.2	0.219	4,143	0.05487±4	0.4882±7	0.06453±9	406.9±1.7
35	8 gr,-200,cl,c,n	12	23.0	367	6.5	0.220	2,854	0.05494±6	0.4904±9	0.06474±10	409.7±2.6
						Seboeis [95MDw132]					
36	8 gr,-200,cl,cr,pb,n	13	246	3800	71.1	0.340	2,802	0.05481±9	0.4874±17	0.06450±20	404.3±3.5
37	2 gr,-200,cl,cr,pb,n	5	177	2800	6.8	0.289	8,458	0.05485±3	0.4842±11	0.06403±14	406.0±1.4
						Shirley [95MDw90B]					
38	2 gr, +200,cl,c,n	5	34.5	561	4.39	0.205	2,731	0.05476±8	0.4820±9	0.06384±9	402.6±3.4
39	4 gr,+200,cl,c,n	9	7.63	117	2.47	0.368	1,817	0.05486±7	0.4894±8	0.06470±8	406.6±3.0
†40	6 gr,-200 cl,c,n	11	36.4	562	4.0	0.326	6,511	0.05513±4	0.4941±7	0.06499±8	417.6±1.7
†41	4 gr,+200,cl,c,a,fr	42	32.6	473	4.7	0.258	18,777	0.05776±4	0.5581±9	0.07008±12	520.6±1.4

Notes:

(1) Cardinal number indicates the number of zircon grains analyzed (e.g., 35 grains); all grains were selected from nonparamagnetic separates at 0 degree tilt at full magnetic field in Frantz Magnetic Separator; +200 = size in mesh (> 75µm); c = colorless; cl= clear; el = elongate; e = equant; eu = euhedral; f = faceted; lp = long prismatic; n = 5:1 prismatic needles; p = prismatic; pb = pale brown; s = stubby; s-p = short-prismatic; t = tips from prisms. NA = nonabraded fraction; all other grains were air-abraded following Krogh (1982). † = analyses rejected from regression analysis because of inheritance.

(2) Concentrations are known to ±30% for sample weights of about 20 µg and ±50% for samples <5 µg.

(3) Corrected for 0.0215 mole fraction common-Pb in the ^{205}Pb-^{235}U spike.

(4) Calculated Th/U ratio assuming that all ^{208}Pb in excess of blank, common-Pb, and spike is radiogenic ($\lambda\ ^{232}Th = 4.9475 \times 10^{-11} y^{-1}$).

(5) Measured, uncorrected ratio.

(6) Ratio corrected for fractionation, spike, blank, and initial common-Pb (at the determined age from Stacey and Kramers (1975)). Pb fractionation correction = 0.094%/amu (±0.025% 1σ); U fractionation correction = 0.111%/amu (± 0.02% 1σ). U blank = 0.2 pg; Pb blank <= 10 pg. Absolute uncertainties (1σ) in the Pb/U and $^{207}Pb/^{206}Pb$ ratios calculated following Ludwig (1980). U and Pb half-lives and isotopic abundance ratios from Jaffey and others (1971).

Ebeemee Pluton

The Ebeemee pluton (EB, fig. 6) is a small, unfoliated stock of biotite granite; fieldwork for this study was limited to sampling in the pluton and a fruitless search for informative outcrops in the contact aureole. Reconnaissance mapping compiled by Osberg and others (1985) showed the pluton cutting a doubly plunging anticline cored by the Madrid Formation and flanked by the Carrabassett Formation; on this basis, the pluton would appear to postdate Acadian folding. Two concordant zircon analyses yielded a $^{207}Pb/^{206}Pb$ age of 405.7±2.6 Ma, and a biotite separate yielded an $^{40}Ar/^{39}Ar$ plateau age of 407.8±2.4 Ma (fig. 10).

Seboeis Pluton

The Seboeis pluton (SS, fig. 6) is shown on the bedrock geologic map of Maine (Osberg and others, 1985) as a large (5 by 30 km) body that cuts the Madrid and Smalls Falls Formations and intervening faults. Because the pluton has not been mapped in detail and was sampled only in reconnaissance fashion, the following observations are necessarily preliminary. We dated four phases. Weakly foliated granodiorite at Schoodic Lake yielded two concordant zircon fractions that give a mean $^{207}Pb/^{206}Pb$ age of 405.8±2.5 Ma (fig. 7). Near Dudley Rips, an unfoliated granite that yielded an $^{40}Ar/^{39}Ar$ muscovite plateau age of 401.3±2.8 Ma (fig. 8) intrudes a tec-

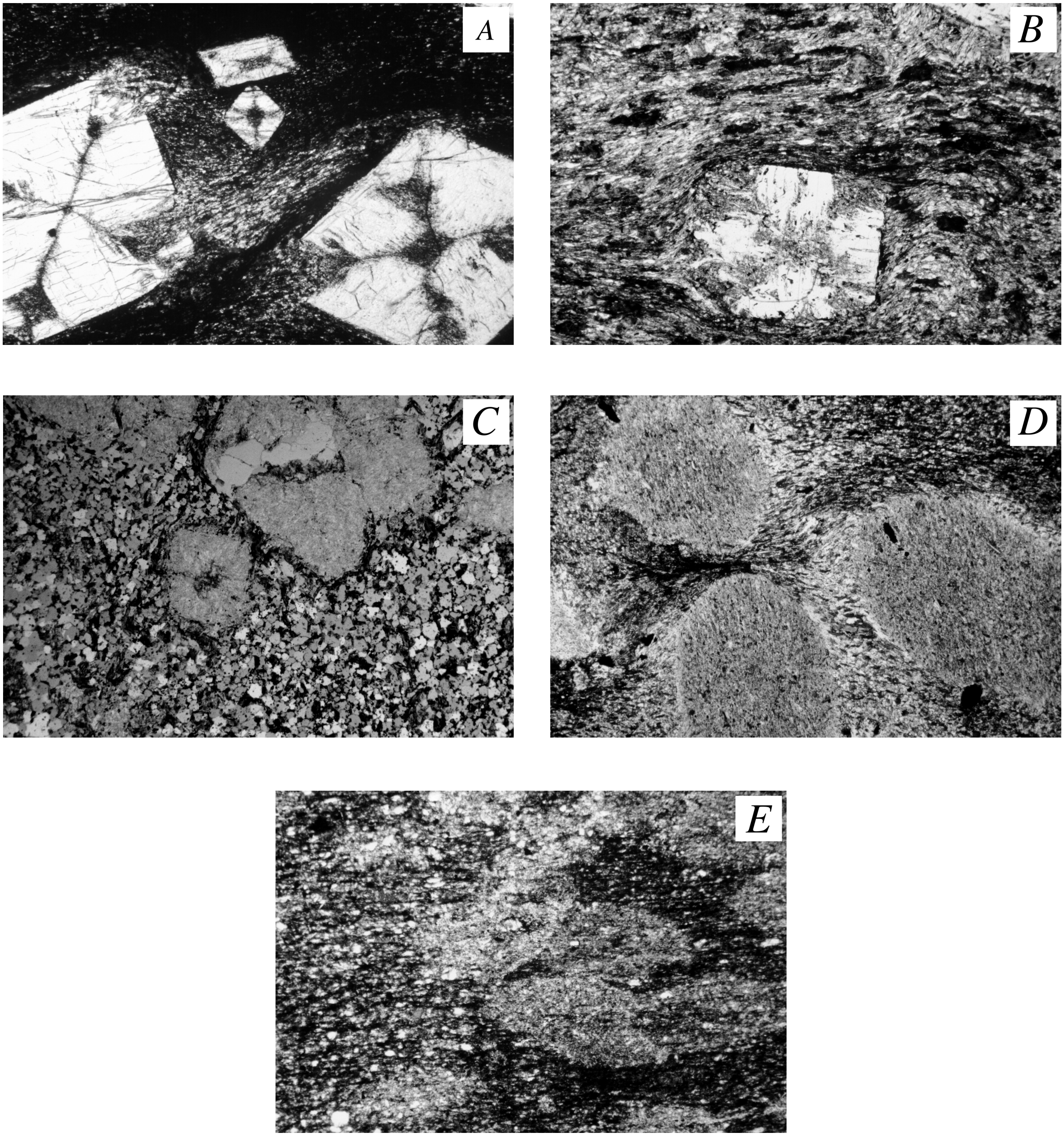


Figure 9.—Thin-section photomicrographs of contact-metamorphosed rocks adjacent to Acadian plutons. All sections are cut normal to foliation and are viewed in plane-polarized light. Field of view in each photomicrograph is about 4 mm across. *A*, Syntectonic metamorphic textures in aureole of the Russell Mountain pluton. Large, randomly oriented chiastolite phenocrysts are flanked by strain shadows containing aligned flakes of biotite, implying some deformation during contact metamorphism. *B*, Syntectonic metamorphic textures in aureole of the Sebec Lake pluton. Chiastolite phenocrysts are flanked by strain shadows containing aligned chlorite and biotite, implying some deformation during contact metamorphism. *C*, Static metamorphic textures in aureole of the Bald Mountain pluton, showing randomly oriented, retrograded chiastolite phenocrysts and randomly oriented biotite. Rock shows no vestige of pervasive regional foliation found outside contact aureole and so may have been metamorphosed before regional deformation. Alternatively, contact metamorphism might have been intense enough to have obliterated all preexisting fabrics. *D*, Syntectonic metamorphic textures in aureole of the Hog Island pluton. Large cordierite phenocrysts overgrow a preexisting foliation defined by aligned biotite and white mica; this first foliation now survives only within cordierite grains. A second, postcordierite foliation is defined by biotite in pressure shadows. Although metamorphic textures in this rock are broadly syntectonic, on the basis of regional relations, the first Acadian deformation probably preceded emplacement of the Hog Island pluton by 15 to 20 m.y. *E*, Posttectonic metamorphic textures in aureole of the Priestly pluton. Cordierite phenocrysts overgrow a weak foliation that lies at a high angle to bedding; biotite in groundmass is randomly oriented.

tonically foliated diorite that yielded a slightly hump shaped $^{40}Ar/^{39}Ar$ age spectrum with a total-gas age of 396.3±3.8 Ma (fig. 8). Evidently, the diorite was intruded before 401 Ma and before at least some of the regional deformation. A much younger biotite age from rocks that had been previously mapped as part of the Seboeis pluton at East Branch Lake is discussed below in the subsection entitled "Givetian to Early Tournaisian Plutons."

Moxie Pluton

The northwestern (inboard) part of the Emsian plutonic belt is dominated by two large intrusions: the Moxie gabbro and the Katahdin quartz monzonite. The intrusive relations of the Moxie pluton (MX, fig. 6) differ considerably from one end to the other. In the Jo Mary Mountain area at the pluton's east end (loc. 30, fig. 1), the contact aureole contains remarkably pristine, gently dipping strata that display perfectly preserved crosslaminae (Hanson and Bradley, 1989). We observed no evidence of contact-metamorphic minerals overgrowing a regional slaty cleavage, which is visible everywhere outside the aureole. Midlength along the pluton (at loc. 44, fig. 1), however, contact-metamorphic andalusite overgrows an axial-planar cleavage associated with nearly recumbent mesoscale folds. At the pluton's west end, rocks of the contact aureole show a strong foliation defined by biotite and white mica and by the preferred orientation of andalusite with asymmetric strain shadows of biotite. Thus, at three places where we have anecdotal observations, the pluton shows local evidence of predating regional Acadian folding, of postdating Acadian folding, and of synkinematic emplacement. We dated the Moxie pluton in its eastern and central sections and obtained $^{40}Ar/^{39}Ar$ biotite plateau ages of 404.4±3.4 and 406.3±3.8 Ma, respectively (fig. 8).

Katahdin Pluton

The Katahdin pluton (KA, fig. 6) is a large (60 by 35 km) body of unfoliated quartz monzonite. It has long had an enigmatic age relationship to the Acadian deformation, but this problem can now be resolved—or, at least, narrowed—with a better time scale and precise U-Pb ages (Bradley and others, 1996; Rankin and Tucker, 1995). The Katahdin pluton intrudes rocks as young as the Seboomook Group (Pragian), the Matagamon Sandstone (Pragian and earliest Emsian), and the pluton's own volcanic carapace, the Traveler Rhyolite (406 Ma; Rankin and Tucker, 1995). Near Harrington Lake (HA, fig. 6), the Seboomook Group near the pluton consists of gently dipping hornfelsed siltstone and sandstone that preserve pristine crossbedding. These rocks show no sign of ever having acquired the regional, northeast-striking, axial-planar cleavage that is pervasive just outside the contact aureole and that typically obscures or obliterates delicate sedimentary structures. In this area, the rocks in the contact aureole were baked and never cleaved, suggesting that the pluton is locally pretectonic. Along the pluton's northeastern border, in contrast, Neumann (1967) mapped a breccia zone consisting of plastically deformed metasedimentary-rock fragments, set in a gneissic granitic matrix. These rocks appear to have formed along a syntectonic intrusive contact. The Katahdin pluton resisted many early attempts to date it accurately, but Denning and Lux (1989) reported an $^{40}Ar/^{39}Ar$ biotite plateau age of 400.1±1 Ma. More recently, Rankin and Tucker (1995) reported a $^{207}Pb/^{206}Pb$ age of 406.9±0.4 Ma that is the weighted average of five concordant zircon fractions.

Harrington Pluton

The Harrington pluton (HA, fig. 6) is a laccolith-like satellite to the Katahdin pluton (Griscom, 1976). Its map pattern (Griscom, 1976) implies that it may have intruded a folding sedimentary section because its basal contact is planar and concordant with homoclinally dipping beds in the subjacent Seboomook Group, whereas its upper contact is concordant with regionally folded bedding in the overlying part of the Seboomook Group. The Harrington pluton yielded three concordant zircon analyses, with a mean $^{207}Pb/^{206}Pb$ age of 406.9±1 Ma (fig. 7), which suggests an Emsian age for the folding.

Late Emsian to Eifelian Plutons

A few plutons in the age range 400–387.5 Ma are scattered across the study area (fig. 6) but form no obvious pattern. The Magagauadavic pluton in southern New Brunswick has a U-Pb age of 396±1 Ma (MG, fig. 6; M.L. Bevier, written commun., 1996). A biotite granite phase of the Mooselookmeguntic pluton (MO, fig. 6) yielded a U-Pb age of 388.9±1.6 Ma (Solar and others, 1998). The North Searsmont granitic gneiss (NS, fig. 6) yielded a U-Pb age of 389.0±2.0 Ma (Tucker and others, in press). The Horserace quartz diorite (HR, fig. 6), a small body within the Katahdin batholith, yielded a concordant U-Pb zircon age of 392±3.1 Ma (R.D. Tucker, unpub. data); an $^{40}Ar/^{39}Ar$ hornblende age of 374.9±3.2 Ma was previously reported for this pluton (Denning and Lux, 1989).

Givetian to Early Tournasian Plutons[2]

Previous Results

The best-defined group of intrusions of this age consists of Frasnian plutons in central Maine. Tucker and others (in press) recently obtained U-Pb ages for four of these plutons: the Hartland (379±1 Ma; HD, fig. 6), Old Point (379±3 Ma; OP, fig. 6), Rome (378±1 Ma; RO, fig. 6), and Togus (378±1 Ma; TO, fig. 6). New ages are reported below for the Center Pond pluton and a sill at Hammond Ridge, which clearly belong in this group. Two other plutons that may belong in the same group crop out near the Maine coast: the Lucerne (380±4 Ma; LU, fig. 6; R.E. Zartman and M.D. Gallego, in Marvin and Dobson, 1979, p. 18) and the Deblois (384±5 Ma;

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[2]Plutons in this age range are widespread in the area of figure 6. In hindsight, they are not particularly relevant to the main point of this report, but they are important to late Acadian events, and their place in the scheme of things could only have been learned by obtaining reliable ages.

Table 5. $^{40}Ar/^{39}Ar$ data for plutonic rocks, Maine.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles $^{39}Ar^{*}$	% total ^{39}Ar	% ^{40}Ar rad	K/Ca	Apparent age (Ma)	Error (Ma)
Bald Mtn. pluton, granodiorite; 95MDw143; biotite; J=.007489									
760	32.92	0.017	0.0068	745.4	7.2	93.8	28.3	375.4	3.7
900	32.78	0.009	0.0019	1746.9	16.8	98.3	54.0	389.8	3.6
1010	32.98	0.013	0.0013	1303.1	12.6	98.8	38.7	393.8	3.9
1090	32.99	0.018	0.0014	1899.3	18.3	98.7	28.0	393.6	3.6
1130	32.94	0.024	0.0006	1967.7	19.0	99.4	20.0	395.5	4.6
1230	32.98	0.026	0.0012	1338.0	12.9	98.9	18.9	394.3	3.6
1310	33.09	0.045	0.0008	803.1	7.7	99.2	10.9	396.6	3.9
FUSE	33.26	0.129	0.0021	564.9	5.4	98.1	3.8	394.5	4.1
TOTAL			10368.3	100.0				392.4	3.9
PLATEAU AGE								394.3	3.7
Bald Mtn. pluton, granodiorite; 95MDw146; biotite; J=.007697									
760	33.32	0.028	0.0295	220.8	1.5	73.8	17.6	312.8	9.7
900	31.72	0.006	0.0025	3063.0	21.4	97.6	86.4	385.7	3.5
1010	31.91	0.008	0.0022	2536.4	17.7	97.9	58.1	388.8	3.7
1090	31.90	0.009	0.0015	1948.3	13.6	98.5	55.8	390.9	3.6
1130	31.72	0.009	0.0009	2259.6	15.8	99.1	56.9	390.8	3.5
1230	31.61	0.013	0.0007	2345.5	16.4	99.3	38.5	390.3	3.5
1310	31.70	0.019	0.0006	1299.2	9.1	99.3	25.4	391.6	3.6
FUSE	31.85	2.337	0.0012	668.7	4.7	99.4	0.2	394.0	3.7
TOTAL			14341.5	100.0				388.3	3.7
PLATEAU AGE								390.9	3.0
Beaver Cove pluton, granodiorite; 95MDw195; biotite; J=.007539									
760	59.90	0.122	0.1505	67.5	0.7	25.8	4.0	198.6	24.8
900	31.77	0.018	0.0086	907.6	9.2	91.9	27.0	359.0	3.6
1010	31.21	0.015	0.0043	1486.4	15.0	95.9	33.3	367.1	3.4
1090	31.26	0.018	0.0036	1894.4	19.1	96.5	27.2	369.7	3.4
1130	31.22	0.035	0.0027	1502.4	15.2	97.4	14.2	372.5	3.5
1230	30.90	0.074	0.0016	1793.3	18.1	98.4	6.7	372.5	3.4
1210	30.82	0.080	0.0012	1599.4	16.1	98.8	6.2	372.8	3.5
FUSE	30.89	0.137	0.0017	664.3	6.7	98.3	3.6	371.9	3.5
TOTAL			9915.1	100.0				368.8	3.6
PLATEAU AGE								372.4	2.7
Center Pond pluton, granodiorite; 95MDw158A; biotite; J=.00787									
770	34.084	0.01791	0.04073	10.0	2.9	64.6	27.4	288.5	13.1
870	30.933	0.00298	0.01092	26.2	7.6	89.5	164.3	355.7	3.5
950	29.676	0.00541	0.00304	67.3	19.5	96.9	90.6	368.1	3.3
1010	29.466	0.00603	0.00209	70.0	20.3	97.9	81.3	368.9	3.3
1085	29.484	0.01491	0.00208	46.4	13.4	97.9	32.9	369.2	3.4
1145	29.548	0.02216	0.00209	37.0	10.7	97.9	22.1	369.9	3.4
1230	29.429	0.04037	0.00115	61.3	17.7	98.8	12.1	371.8	3.4
FUSE	29.385	0.03921	0.00139	27.4	7.9	98.6	12.5	370.4	3.3
TOTAL			345.6	100.0				366.2	3.6
PLATEAU AGE								368.7	2.9
Chandler pluton, granodiorite; 95MDw106; biotite; J=.007527									
760	29.22	0.106	0.0524	67.0	0.9	47.0	4.6	177.3	20.8
900	31.35	0.019	0.0085	404.1	5.3	91.9	25.2	354.2	3.9
1010	31.24	0.019	0.0052	547.3	7.2	95.0	25.4	363.7	3.8
1090	31.33	0.023	0.0030	677.5	8.9	97.1	21.0	371.9	3.6
1130	30.94	0.024	0.0017	1034.6	13.7	98.4	20.2	372.1	3.5
1230	30.50	0.018	0.0009	1659.7	21.9	99.0	26.9	369.6	3.5
1310	30.26	0.013	0.0006	2066.3	27.3	99.3	37.8	367.9	3.4
FUSE	30.45	0.019	0.0013	1117.5	14.8	98.7	25.2	368.0	3.8
TOTAL			7573.9	100.0		366.5		3.7	
Deboullie pluton, monzonite; 95MDw112; biotite; J=.007521									
760	34.57	0.021	0.0201	216.1	2.2	82.7	22.9	351.5	6.3
900	30.90	0.006	0.0041	677.8	6.9	96.0	77.5	363.3	3.5
1010	30.50	0.004	0.0018	677.5	6.9	98.2	132.6	366.5	3.7
1090	30.45	0.006	0.0018	784.5	8.0	98.2	82.5	366.0	3.5
1130	30.51	0.005	0.0012	686.7	7.0	98.8	106.7	368.6	3.7
1230	30.18	0.004	0.0008	1650.2	16.8	99.1	127.8	366.2	3.4
1310	30.05	0.005	0.0007	2874.0	29.3	99.3	100.7	365.1	3.4
FUSE	30.16	0.008	0.0011	2235.7	22.8	98.9	62.9	365.0	3.4
TOTAL			9802.3	100.0				365.3	3.5
PLATEAU AGE								365.4	3.2

Table 5. $^{40}Ar/^{39}Ar$ data for plutonic rocks, Maine—Continued.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles $^{39}Ar^*$	% total ^{39}Ar	% ^{40}Ar rad	K/Ca	Apparent age (Ma)	Error (Ma)
			Deboullie pluton, monzonite; 95MDw114; biotite; J=.007917						
760	94.97	0.015	0.2690	33.3	0.3	16.3	32.7	208.5	28.3
900	32.73	0.004	0.0174	564.7	4.9	84.2	133.8	356.1	3.7
1010	29.57	0.001	0.0045	1027.9	8.9	95.4	692.6	363.9	3.4
1090	29.06	0.004	0.0019	915.2	8.0	98.0	129.2	367.0	3.6
1130	28.74	0.004	0.0010	1170.6	10.2	98.9	112.5	366.3	3.4
1230	28.38	0.003	0.0005	2513.7	21.9	99.4	145.9	363.6	3.4
1310	28.25	0.004	0.0004	3387.2	29.5	99.5	126.6	362.5	3.3
FUSE	28.37	0.008	0.0005	1883.9	16.4	99.4	61.6	363.5	3.3
TOTAL			11496.4	100.0				363.0	3.5
PLATEAU AGE								363.2	3.0
			Deboullie pluton, granodiorite; 95MDw115; biotite; J=.007839						
760	35.11	0.086	0.0312	131.0	3.2	73.7	5.7	333.2	12.9
900	30.86	0.045	0.0094	368.3	9.1	91.0	11.0	359.0	5.6
1010	30.18	0.052	0.0050	437.5	10.8	95.1	9.4	366.1	5.7
1090	29.93	0.061	0.0032	352.8	8.7	96.8	8.0	369.1	6.4
1130	29.47	0.068	0.0019	454.2	11.2	98.1	7.2	368.4	3.8
1230	28.80	0.052	0.0007	1076.7	26.6	99.3	9.5	364.8	3.4
1310	28.75	0.047	0.0007	927.4	22.9	99.3	10.5	364.2	3.4
FUSE	29.02	0.071	0.0018	301.5	7.4	98.1	6.9	363.5	5.0
TOTAL			4049.4	100.0				363.9	4.6
PLATEAU AGE								364.2	3.2
			Ebeemee pluton, granodiorite; 95MDw169; biotite; J=.007458						
760	31.52	0.055	0.0131	35.7	6.7	87.7	9.0	338.0	6.3
900	33.59	0.020	0.0046	74.7	14.1	95.9	24.4	388.4	4.0
1010	34.40	0.016	0.0033	82.5	15.5	97.1	30.7	401.4	4.1
1090	34.65	0.015	0.0022	89.4	16.8	98.1	32.7	407.6	3.7
1130	34.49	0.011	0.0015	93.3	17.6	98.6	43.7	408.0	3.7
1230	34.47	0.031	0.0015	96.3	18.1	98.7	15.9	407.8	4.1
1310	35.10	0.069	0.0049	31.2	5.9	95.9	7.1	404.1	4.1
FUSE	35.98	0.053	0.0074	27.9	5.2	93.9	9.3	405.3	4.2
TOTAL			531.0	100.0				399.0	4.1
PLATEAU AGE								407.8	2.4
			Hog Island pluton, granodiorite; 95MDw205; biotite; J=.00792						
760	30.04	0.027	0.00947	27.6	4.2	90.6	17.9	352.2	3.2
900	28.98	0.018	0.00292	75.9	11.6	97.0	28.0	362.6	3.3
1010	29.13	0.014	0.00251	70.2	10.7	97.4	34.5	365.7	3.3
1090	29.19	0.015	0.00186	76.3	11.6	98.1	31.7	368.7	3.3
1130	29.06	0.021	0.00114	96.6	14.7	98.8	23.2	369.6	3.3
1230	28.83	0.027	0.00068	144.7	22.1	99.3	18.3	368.5	3.3
1310	28.86	0.046	0.00101	105.6	16.1	98.9	10.7	367.8	3.4
FUSE	28.94	0.059	0.00103	58.8	9.0	98.9	8.3	368.6	3.3
TOTAL			655.7	100.0				366.9	6.7
PLATEAU AGE								368.6	3.1
			Mattamiscontis pluton, granodiorite; 95MDw164B; biotite; J=.007498						
880	34.89	0.029	0.0172	374.6	8.5	85.4	16.6	363.6	3.9
1020	34.12	0.028	0.0046	793.9	18.0	96.0	17.5	396.2	3.6
1110	34.05	0.038	0.0023	1187.4	27.0	97.9	13.1	402.5	3.9
1180	33.96	0.043	0.0023	915.5	20.8	98.0	11.5	401.8	3.7
1310	33.85	0.098	0.0019	741.9	16.9	98.3	5.0	402.0	3.8
FUSE	33.85	0.124	0.0035	388.4	8.8	96.9	3.9	396.7	4.4
TOTAL			4401.8	100.0				397.3	3.8
PLATEAU AGE								402.1	2.8
			Moxie pluton, gabbro; 95MDw187; biotite; J=.007449						
760	40.39	0.033	0.0269	168.3	1.8	80.3	15.0	390.3	10.9
900	35.91	0.023	0.0076	533.9	5.6	93.7	21.4	403.4	4.2
1010	35.14	0.024	0.0038	2118.6	22.3	96.7	20.7	407.2	3.8
1090	34.82	0.023	0.0018	960.6	10.1	98.4	21.7	410.2	4.0
1130	34.43	0.018	0.0013	1410.1	14.9	98.8	26.6	407.6	3.7
1230	34.08	0.011	0.0007	1942.7	20.5	99.3	43.6	405.8	3.7
1310	34.10	0.019	0.0007	1658.5	17.5	99.3	26.1	405.8	3.7
FUSE	34.30	0.027	0.0014	694.2	7.3	98.8	17.9	405.9	4.0
TOTAL			9487.0	100.0				406.4	3.9
PLATEAU AGE								406.3	3.8

Table 5. $^{40}Ar/^{39}Ar$ data for plutonic rocks, Maine—Continued.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles $^{39}Ar^*$	% total ^{39}Ar	% ^{40}Ar rad	K/Ca	Apparent age (Ma)	Error (Ma)
			Moxie pluton, gabbro; 95MDw198; biotite; J=.007962						
760	88.47	0.229	0.23360	2.4	0.6	22.0	2.1	259.7	41.4
900	37.51	0.085	0.02270	20.4	5.3	82.1	5.8	395.6	3.9
1010	34.38	0.086	0.00941	34.6	9.0	91.9	5.7	404.8	3.6
1090	33.25	0.083	0.00474	47.2	12.3	95.8	5.9	407.7	3.7
1130	32.47	0.059	0.00284	67.1	17.4	97.4	8.4	405.2	3.6
1230	32.00	0.038	0.00140	104.1	27.1	98.7	13.0	404.5	3.6
1310	31.88	0.038	0.00103	76.4	19.8	99.0	12.9	404.5	3.6
FUSE	32.07	0.048	0.00194	32.6	8.5	98.2	10.2	403.5	3.7
TOTAL			384.7	100.0				403.6	3.9
PLATEAU AGE								404.4	3.4
			Onawa pluton, gabbro phase; 95MDw68B; biotite; J=.007633						
760	40.59	0.048	0.0586	19.9	2.3	57.3	10.1	294.7	6.4
900	33.78	0.020	0.0073	126.0	14.5	93.5	24.6	389.9	4.0
1010	33.43	0.016	0.0041	176.3	20.2	96.4	31.3	396.6	3.6
1090	33.57	0.020	0.0026	104.0	11.9	97.7	24.9	402.9	3.7
1130	33.15	0.026	0.0014	144.1	16.5	98.7	19.0	402.3	3.7
1230	32.69	0.015	0.0005	97.2	11.1	99.5	33.6	400.2	3.6
1310	32.80	0.013	0.0006	115.2	13.2	99.4	37.7	401.0	3.9
FUSE	32.87	0.038	0.0014	89.4	10.2	98.7	12.8	399.2	3.9
TOTAL			872.1	100.0				396.2	3.8
PLATEAU AGE								400.1	3.7
			Priestly pluton, granodiorite; 95MDw107; biotite; J=.00786						
760	31.01	0.010	0.0159	356.9	3.4	84.8	49.2	338.9	4.5
900	29.47	0.013	0.0061	756.1	7.2	93.8	37.7	354.7	3.5
1010	29.33	0.022	0.0043	938.9	9.0	95.6	22.7	359.3	3.5
1090	29.18	0.051	0.0033	1102.0	10.5	96.6	9.6	361.2	3.6
1130	28.88	0.060	0.0020	1497.7	14.3	97.9	8.1	362.0	3.4
1230	28.47	0.038	0.0012	2528.4	24.2	98.7	12.9	360.1	3.3
1310	28.42	0.029	0.0010	2166.4	20.7	98.9	16.9	360.0	3.3
FUSE	28.59	0.032	0.0014	1118.9	10.7	98.5	15.4	360.8	3.4
TOTAL			10465.4	100.0				359.3	3.4
PLATEAU AGE								360.3	2.7
			Priestly pluton, granodiorite; 95MDw108; biotite; J=.007509						
760	50.26	0.111	0.0911	4.2	1.1	46.4	4.4	291.2	26.3
900	31.69	0.017	0.0087	26.7	7.0	91.8	28.9	356.4	4.1
1010	30.84	0.018	0.0044	38.0	9.9	95.8	28.0	361.3	3.8
1090	30.75	0.029	0.0034	38.6	10.0	96.7	17.1	363.5	3.9
1130	30.70	0.083	0.0024	43.4	11.3	97.7	5.9	366.4	3.6
1230	30.18	0.073	0.0013	74.1	19.3	98.7	6.7	364.2	3.4
1310	29.92	0.052	0.0008	124.6	32.5	99.2	9.3	363.0	3.3
FUSE	30.45	0.059	0.0025	34.4	9.0	97.6	8.3	363.3	5.8
TOTAL			384.0	100.0				362.3	4.0
PLATEAU AGE								363.2	2.3
			Russell Mtn. pluton; 95MDw80; biotite; J=.00755						
760	31.44	0.014	0.0079	639.2	5.0	92.6	34.7	358.3	3.6
900	32.11	0.005	0.0021	1722.9	13.5	98.0	98.7	384.4	3.5
1010	32.40	0.006	0.0019	1649.9	13.0	98.2	85.6	388.3	3.5
1090	32.71	0.004	0.0016	1721.9	13.5	98.5	116.0	392.8	3.6
1130	32.62	0.005	0.0010	2327.8	18.3	99.1	97.3	393.8	3.5
1230	32.44	0.006	0.0006	2508.1	19.7	99.3	81.1	392.9	3.6
1310	32.53	0.008	0.0007	1530.2	12.0	99.3	65.1	393.9	3.6
FUSE	32.66	0.011	0.0012	619.5	4.9	98.8	44.0	393.4	3.8
TOTAL			12719.5	100.0				389.7	3.6
PLATEAU AGE								393.4	3.0
			Russell Mtn. pluton; 95MDw98; muscovite; J=.007668						
770	65.509	0.04333	0.12239	3.7	1.1	44.8	11.3	366.0	22.8
870	38.853	0.01887	0.02433	9.3	2.7	81.5	26.0	392.0	6.3
950	33.449	0.00837	0.00506	19.2	5.5	95.5	58.6	395.2	3.8
1010	32.466	0.01020	0.00212	50.8	14.5	98.0	48.0	393.9	3.6
1085	32.277	0.00489	0.00162	70.3	20.1	98.5	100.2	393.5	3.6
1150	32.107	0.00234	0.00096	53.5	15.3	99.1	209.2	393.7	3.5
1230	32.111	0.00589	0.00096	83.8	24.0	99.1	83.2	393.8	3.5
FUSE	32.472	0.03211	0.00241	58.6	16.8	97.8	15.3	393.1	3.8
TOTAL			349.3	100.0				393.4	3.9
PLATEAU AGE								393.6	2.7

Table 5. $^{40}Ar/^{39}Ar$ data for plutonic rocks, Maine—Continued.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles $^{39}Ar^*$	% total ^{39}Ar	% ^{40}Ar rad	K/Ca	Apparent age (Ma)	Error (Ma)
Sebec Lake pluton, granodiorite; 95MDw71; muscovite; J=.007477									
850	35.78	0.020	0.02633	14.4	5.1	78.2	24.1	342.8	4.5
950	32.19	0.031	0.00872	20.6	7.2	92.0	16.0	360.7	11.5
1010	32.41	0.017	0.00436	34.4	12.0	96.0	28.3	377.2	3.6
1090	32.31	0.021	0.00356	42.6	14.9	96.7	24.4	378.8	3.4
1130	32.04	0.016	0.00317	39.4	13.8	97.0	30.9	377.1	3.4
1210	31.77	0.017	0.00266	43.6	15.3	97.5	29.5	375.8	3.4
1270	31.89	0.013	0.00167	45.2	15.9	98.4	38.9	380.3	3.5
FUSE	31.42	0.019	0.00181	45.1	15.8	98.2	25.7	374.7	3.4
TOTAL			285.4	100.0				374.4	4.1
PLATEAU AGE								377.2	4.4
Sebec Lake pluton, granodiorite; 95MDw72; biotite; J=.007885									
750	15.616	0.00261	0.00658	88.8	14.4	87.4	187.4	184.5	1.9
900	31.175	0.00810	0.00360	77.1	12.5	96.5	60.5	384.2	17.5
1090	31.219	0.00277	0.00122	147.4	23.8	98.8	176.6	392.7	3.5
1130	31.093	0.00313	0.00090	154.2	24.9	99.1	156.3	392.4	3.5
1230	31.269	0.00634	0.00046	79.5	12.9	99.5	77.3	395.9	3.6
1300	31.153	0.01663	0.00077	69.3	11.2	99.2	29.5	393.5	3.5
FUSE	31.031	0.08028	0.01556	2.1	0.3	85.1	6.1	341.4	17.5
TOTAL			618.4	100.0				362.0	5.1
PLATEAU AGE								392.5	2.5
Seboeis pluton, diorite; 95MDw135A; biotite; J=.007937									
770	39.935	0.07719	0.07671	6.4	1.4	43.2	6.3	231.6	11.5
870	33.444	0.02203	0.01627	26.3	5.8	85.6	22.2	369.3	4.2
950	32.429	0.01091	0.00642	47.2	10.5	94.1	44.9	391.3	3.5
1010	32.378	0.01232	0.00461	56.9	12.7	95.7	39.8	396.9	3.6
1085	32.314	0.01106	0.00221	73.2	16.3	97.9	44.3	404.3	3.6
1145	32.100	0.00865	0.00150	79.1	17.6	98.6	56.6	404.2	3.6
1230	31.710	0.00868	0.00105	101.8	22.6	99.0	56.5	401.3	3.6
FUSE	31.956	0.00836	0.00204	58.5	13.0	98.1	58.6	400.7	4.3
TOTAL			449.5	100.0				396.3	3.8
Seboeis pluton, granite; 95MDw135B; muscovite; J=.007465									
760	46.10	0.030	0.04967	7.7	1.5	68.1	16.4	380.0	9.3
900	35.11	0.008	0.00680	39.4	7.7	94.2	61.0	398.2	4.2
1010	34.19	0.003	0.00263	95.5	18.8	97.7	183.5	401.6	3.6
1090	34.02	0.007	0.00220	88.3	17.3	98.0	69.4	401.1	3.6
1130	33.87	0.003	0.00143	95.0	18.7	98.7	162.4	401.9	3.6
1230	33.80	0.004	0.00156	94.8	18.6	98.6	135.0	400.8	3.6
1310	33.84	0.006	0.00153	57.8	11.3	98.6	137.9	401.3	3.6
FUSE	33.44	0.013	0.00315	31.0	6.1	97.2	36.9	391.8	4.5
TOTAL			509.4	100.0				400.2	6.7
PLATEAU AGE								401.3	2.8
Seboeis pluton, granodiorite; 95MDw139B; biotite; J=.007437									
760	47.15	0.068	0.0896	3.5	0.6	43.8	7.2	258.0	26.8
900	32.15	0.015	0.0086	45.0	8.0	92.0	33.6	358.8	4.5
1010	31.29	0.015	0.0044	73.8	13.2	95.8	33.2	363.0	3.7
1090	31.03	0.012	0.0026	76.5	13.7	97.5	40.7	366.2	3.4
1130	30.77	0.010	0.0019	94.4	16.9	98.2	50.0	365.6	3.7
1230	30.49	0.008	0.0013	115.4	20.6	98.7	58.4	364.2	3.4
1310	30.40	0.009	0.0013	106.3	19.0	98.7	56.2	363.2	4.0
FUSE	30.86	0.015	0.0021	44.8	8.0	98.0	33.4	365.9	3.7
TOTAL			559.8	100.0				363.4	3.9
PLATEAU AGE								364.7	4.3
Shirley pluton; 95MDw90B; biotite; J=7.888001x10^{-3}									
720	40.59	0.020	0.0529	7.5	0.9	61.4	24.8	323.9	15.9
900	31.36	0.005	0.0038	107.4	12.5	96.4	103.0	385.7	3.6
1010	31.15	0.007	0.0019	135.5	15.8	98.2	74.9	389.9	3.6
1090	31.13	0.008	0.0015	129.7	15.1	98.5	63.0	391.0	3.5
1130	30.86	0.007	0.0009	178.2	20.8	99.1	75.2	389.8	3.5
1230	30.91	0.011	0.0008	167.2	19.5	99.2	45.3	390.6	3.6
1310	31.11	0.028	0.0009	83.1	9.7	99.1	17.3	392.8	3.7
FUSE	31.38	0.036	0.0018	49.0	5.7	98.3	13.8	392.8	4.1
TOTAL			857.6	100.0				389.5	3.7
PLATEAU AGE;								390.3	3.0

moles $^{39}Ar^*$=4.5x10^{-14}

DE, fig. 6; A. Ludman, written commun., 1998). Both of these plutons have been dated by the U-Pb zircon method, but both ages have fairly large errors.

A cluster of seven previously dated plutons in the Eastern Townships of Québec and adjacent Maine fall within the Givetian to Famennian age range. Those plutons dated by the U-Pb method include the Scotstown (384±2 Ma; SC, fig. 6), Lac aux Araignées (in Québec, or Spider Lake, in Maine; 383±3 Ma; LA, fig. 6), Winslow (377±6 Ma; WI, fig. 6), Aylmer (375±3 Ma; AY, fig. 6), and Ste.-Cécile (374±1 Ma; ST, fig. 6; Simonetti and Doig, 1990). Plutons in this group dated by the $^{40}Ar/^{39}Ar$ method include the Chain of Ponds (373.3±2 Ma; CH, fig. 6) and Big Island Pond plutons (367.7±1.3 Ma; BI, fig. 6; Heitzler and others, 1988). The newly dated Hog Island and Beaver Cove, discussed below, appear to belong in this group. Three other new ages from the Priestly, Deboullie, and Chandler plutons, also discussed below, are slightly younger but occur within the same general strike belt. Five previously dated plutons of Famennian age form a belt near the coast. In Maine, these plutons include the Waldoboro pluton (368±2 Ma; WD, fig. 6), the Mount Waldo pluton (371±2 Ma; MW, fig. 6), and two small pegmatite bodies dated at 367±1 and 371±1 Ma (GA, GR, fig. 6; Tucker and others, in press). (All ages are U-Pb zircon ages.) Similarly, M.L. Bevier (written commun., 1996) reported a U-Pb age of 366.5±1 Ma from the Mount Douglas pluton in southern New Brunswick (MD, fig. 6).

Center Pond Pluton

The Center Pond pluton (CP, fig. 6) is a hornblende-bearing granodiorite that intrudes steeply dipping, calcareous siltstone of Silurian age (Scambos, 1980). Two concordant zircon analyses yielded a mean $^{207}Pb/^{206}Pb$ age of 377±3 Ma (fig. 7). The pluton is cut in half and offset dextrally about 2 km by the Center Pond fault (Scambos, 1980). A biotite separate from a penetratively dextrally sheared granitoid near the fault yielded an $^{40}Ar/^{39}Ar$ age of 368.7±2.9 Ma (fig. 8), which may date the strike-slip event.

Hammond Ridge Sill

At Hammond Ridge in the aureole of the Katahdin pluton (HM, fig. 6), a 50-cm-thick, tabular, concordant body of felsite interlayered with hornfels of the Carrabassett Formation yielded five concordant zircon fractions with a mean $^{207}Pb/^{206}Pb$ age of 377.2±1.5 Ma (fig. 7). This sill was at first mistaken for an ash bed that would have dated the Carrabassett Formation (the reason why we analyzed so many fractions), but its young age requires that it be a sill.

Hog Island Pluton

The Hog Island pluton (HI, fig. 6) intrudes Ordovician granodiorite, Silurian calcareous rocks, and Devonian flysch of the Seboomook Group at the north end of the Boundary Mountains anticlinorium. The pluton was mapped by Albee and Boudette (1972), who interpreted the contact metamorphism as either synregional or postregional metamorphism. A metapelite that contains biotite and cordierite, both formed during the contact-metamorphic event (the regional metamorphic grade is chlorite), is shown in figure 9*D*. Cordierite porphyroblasts overgrew a weak foliation, which, we suggest, formed during an earlier Acadian deformation; a second foliation is defined by aligned biotite, some of which occurs in strain shadows around cordierite. Although the Hog Island pluton is probably much younger than the first Acadian deformation, its emplacement was nonetheless synkinematic. A single zircon fraction yielded a concordant $^{207}Pb/^{206}Pb$ age of 371±3 (fig. 7); biotite yielded an $^{40}Ar/^{39}Ar$ plateau age of 368.6±3.1 Ma (fig. 8).

Beaver Cove Pluton

The Beaver Cove pluton (BC, fig. 6) is a small, unfoliated stock of leucocratic biotite granodiorite surrounded by the Moxie pluton. We obtained an $^{40}Ar/^{39}Ar$ plateau age of 372.4±2.7 Ma from a biotite separate (fig. 8), suggesting that the Beaver Cove pluton is as much as 34 m.y. younger than the Moxie pluton and unrelated to it (see Gabis and others, 1994).

Deboullie Pluton

The Deboullie pluton (DB, fig. 6) is a composite body that includes syenite, granodiorite, and monzonite phases (Boone, 1958). Boone's mapping showed that the pluton truncates steeply dipping bedding and cleavage in the Seboomook Group and so postdates the main regional cleavage. Textures from the contact aureole, however, indicate that plutonism was followed by some additional shortening: Contact-metamorphic biotite is cut by a strong pressure-solution cleavage that parallels the regional fabric. Two samples of pink monzonite yielded $^{40}Ar/^{39}Ar$ plateau ages of 365.4±3.2 and 363.2±3.0 Ma on biotite. Similarly, biotite from granodiorite yielded an $^{40}Ar/^{39}Ar$ age of 364.2±3.2 (fig. 8).

Priestly Pluton

The Priestly pluton (PR, fig. 6), a stock of unfoliated biotite-hornblende granodiorite, intrudes flysch of the Seboomook Group (Boudette and others, 1976). Textures in the contact aureole suggest that the pluton was emplaced into already-foliated slate. At the north end of the pluton, cordierite porphyroblasts contain inclusion trails of an older slaty cleavage (fig. 9*E*). Biotite is randomly oriented and indicates static conditions of contact metamorphism; cordierite porphyroblasts grew preferentially in the direction of the older foliation, apparently also under static metamorphic conditions. A single zircon fraction yielded a concordant $^{207}Pb/^{206}Pb$ age of 360.5±3.3 (fig. 7). $^{40}Ar/^{39}Ar$ plateau ages of 363.2±2.3 and 360.3±2.7 Ma were obtained on biotite (fig. 8).

Chandler Pluton

The Chandler pluton (CH, fig. 6) intrudes flysch of the Seboomook Group and Ordovician mafic igneous rocks along the eastern limb of the Munsungun anticlinorium. As far as we are aware, the pluton and surrounding rocks have never been mapped in detail, and outcrop is not plentiful. A total-gas $^{40}Ar/^{39}Ar$ age of 366.5±3.7 Ma was obtained on biotite (fig. 8); the age spectrum is slightly hump shaped.

Seboeis Pluton

As mentioned above, although the Seboeis pluton is shown on the "Geologic Map of Maine" (Osberg and others, 1985) as a single body, it appears to include granodiorite, granite, and diorite of Emsian age, as well as a much younger phase. At East Branch Lake (SE, fig. 6), an unfoliated biotite granodiorite yielded an $^{40}Ar/^{39}Ar$ plateau age of 364.7±4.3Ma (fig. 8).

The Silurian-Devonian Time Scale: Basis for Correlations Between Paleontologically and Isotopically Dated Events

We are able to make precise—and, in some cases, unexpected—correlations between isotopically and paleontologically dated events because of a parallel effort to improve the Silurian-Devonian time scale (Tucker and others, 1998). New tiepoints require significant changes to previous time scales; each tiepoint is based on replicate, concordant, isotope-dilution U-Pb zircon analyses of ashes or volcanic strata, dated to within one or two conodont zones. The key dates are as follows: (1) a concordant U-Pb zircon age of 417.6±1.0 Ma on the Bald Hill ash from the early Lochkovian Kalkberg Formation of eastern New York; (2) a concordant U-Pb zircon age of 408.3±1.9 Ma on the Sprout Brook ash from the early Emsian Esopus Formation of eastern York; (3) a concordant U-Pb zircon age of 391.4±1.8 Ma on the Eifelian Tioga ash of Wytheville, Va., confirming the concordant U-Pb monazite age of 390.0±0.5 Ma reported for the Tioga ash of Pennsylvania by Roden and others (1990); (4) a concordant U-Pb zircon age of 381.1±1.3 Ma on the Little War Gap ash from the Frasnian part of the Chattanooga Shale of eastern Tennessee; and (5 and 6) concordant U-Pb zircon ages of 363.8±2.2 and 363.4±1.8 Ma on rhyolite that bracket a closely dated palyniferous horizon in the Famennian Piskahegan Group of New Brunswick. The vertical axis in figure 4 is numerically calibrated according to the new time scale of Tucker and others (1998), which is contrasted with that of Harland and others (1990) in table 6.

These revisions are critical to the present study. Using the new time scale, most of the closely dated plutons listed in table 3 are now correlated with a younger stage than they would be on the time scale of Harland and others (1990). Among the plutons that are critical to positioning the deformation front, the new time scale shows that (1) the 424- to 417-Ma plutons in southern Maine are Ludlow to Lochkovian rather than Ludlow only, (2) the 404-408 plutons of interior Maine are Emsian rather than Lochkovian, and (3) the Scotstown pluton of the Eastern Townships is Givetian rather than Eifelian. Looking at it another way, the new time scale implies that the Sebec Lake, Onawa, Russell Mountain, and Blanchard plutons are the same age as the Tomhegan Formation, and not the Beck Pond Limestone, as implied by the Harland and others (1990) time scale. Paleogeographic maps based on the old time scale would have been fraught with such errors.

Table 6.—Comparison of time scales.

[All values in millions of years before present (Ma)]

Stage or series boundary	Harland and others (1990)	Tucker and others (1998)
Base of Tournaisian (= base of Carboniferous).	362.5	362
Base of Famennian	367.0	376.5
Base of Frasnian	377.4	382.5
Base of Givetian	380.8	387.5
Base of Eifelian	386.0	394
Base of Emsian	390.4	409.5
Base of Pragian	396.3	413.5
Base of Lochkovian (= base of Devonian).	408.5	418
Base of Pridoli	410.7	419
Base of Ludlow	424.0	424

Location Through Time of the Acadian Deformation Front and Foreland Basin

In this section, we synthesize the foregoing results with the aid of two maps (figs. 10, 11) that track the migration of the leading edge of the foreland basin and of the deformation front. Seven time slices are depicted, each spanning a few million years. The gross paleogeography and key stratigraphic localities are shown in figure 12 for the four best-constrained time slices.

In constructing these figures, we interpolated the leading edge of the foreland basin at a given time between the positions of foreland-basin and far-foreland deposits of that age. Likewise, the deformation front was interpolated between fossiliferous foreland-basin deposits laid down before regional deformation and dated plutons that cut Acadian structures. In some places, as discussed below, figures 10 and 11 were used interdependently to help fill in gaps in knowledge; mostly, however, these two maps are based on independent data that nonetheless lead to the same general conclusions. For the purposes of figure 10, the details of locally complex Acadian deformation histories are not crucial because the point is merely to locate a boundary between rocks that had already undergone at least some contractional deformation from those that still had not. Likewise, unresolved aspects of the pretectonic, syntectonic, or posttectonic nature of various plutons are not crucial for present purposes, so long as evidence exists that some regional deformation occurred in a particular plutonic belt before pluton emplacement.

Palinspastic Considerations

The key localities are plotted on a present-day base map; no attempt has been made to remove the effects of Acadian folding and thrust faulting, post-Acadian shortening, or dextral displacement across the Norumbega or other strike-slip faults. We originally had intended to use a palinspastic base, but we found it impossible to construct such a map on the basis of published maps and cross sections. There are simply too few good marker units to unravel the structure, and the exposure in Maine is too poor.

We can, however, make three comments in support of our use of the present-day base. First, none of the rocks in the study area (fig. 1) has been displaced enormous, immeasurable distances across terrane-bounding faults since the beginning of the Ludlow. By that time, the Pocomoonshine pluton had linked the Fredericton Basin with the St. Croix belt (West

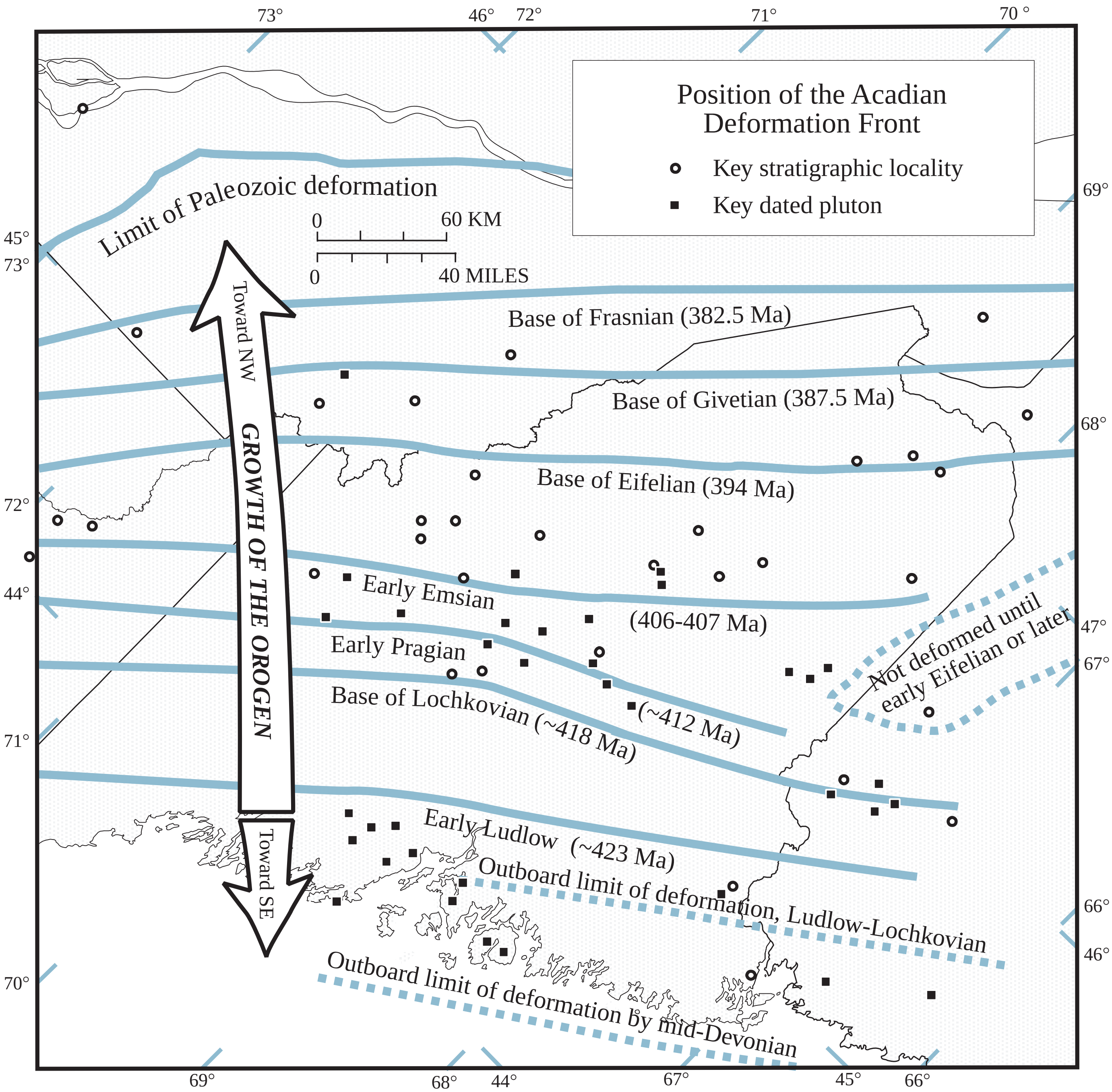

Figure 10.—Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing successive positions of Acadian deformation front as it migrated across the northern Appalachians.

and others, 1992), which, in turn, was linked by clasts in the Silurian Oak Bay conglomerate to the Coastal volcanic belt (loc. 43, fig. 1; Ludman, 1981). The Fredericton Basin and Miramichi anticlinorium were linked by conglomerate clasts in the Taxis River beds by the Early Silurian (loc. 36, fig. 1). The Aroostook-Matapedia Basin was two sided because it has provenance linkages to anticlinoria to both the west and east (Berry and Osberg, 1989; Hopeck, 1991). Specifically, the Llandovery Frenchville Formation near the basin's western margin was derived from the Pennington anticlinorium just to the west (loc. 25, fig. 1; Roy and Mencher, 1976), whereas the conglomeratic Daggett Ridge Formation (probably Late Ordovician to middle Silurian; Ludman, 1990) on the eastern margin was derived from the Miramichi anticlinorium (loc. 36, fig. 1). The Central Maine Basin is linked stratigraphically to older rocks on the northwest; specifically, conglomerate in the Llandovery Rangeley Formation (loc. 26, fig. 1) includes clasts traced to the Attean pluton along the Boundary Mountains anticlinorium (Moench and Pankiwskyj, 1988). Finally, the Connecticut Valley-Gaspé Basin can be

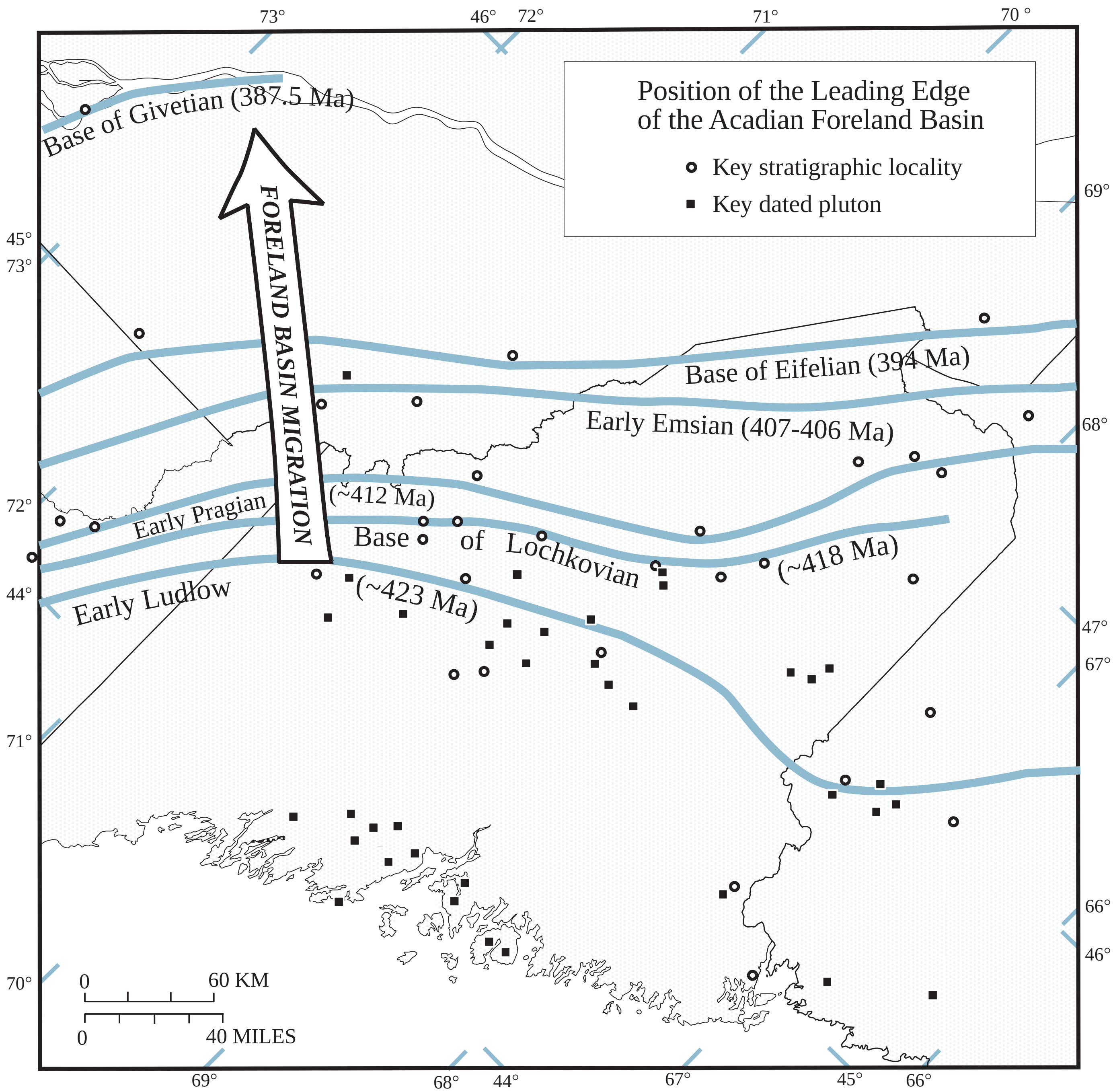

Figure 11.—Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing successive positions of leading edge of the Acadian foreland basin, constructed by following approach used by Bradley (1989) for the Taconic foreland basin.

seen as a foreshortened, two-sided basin because (1) Late Ordovician to Early Silurian conglomerate of the Cabano and Depot Mountain Formations along its northwestern margin in Québec and northernmost Maine was derived from Taconic highlands that lay to the northwest (loc. 6, fig. 1; Roy, 1989) and (2) along the opposite basin margin in northern Maine, the Fish River Lake Formation includes sandstone and conglomerate that can be traced to sources in the Pennington anticlinorium, just to the east (loc. 21, fig. 1; Boone, 1970). Although the contact between the Central Maine and Aroostook-Matapedia Basins is probably a major thrust fault (A. Ludman, written commun., 1998), the two basins shared a common northwesterly source region during the Silurian and so are not exotic relative to each other. Similarly, the Central Maine and Fredericton Basins are in fault contact and may not have been continuous with each other, as some workers

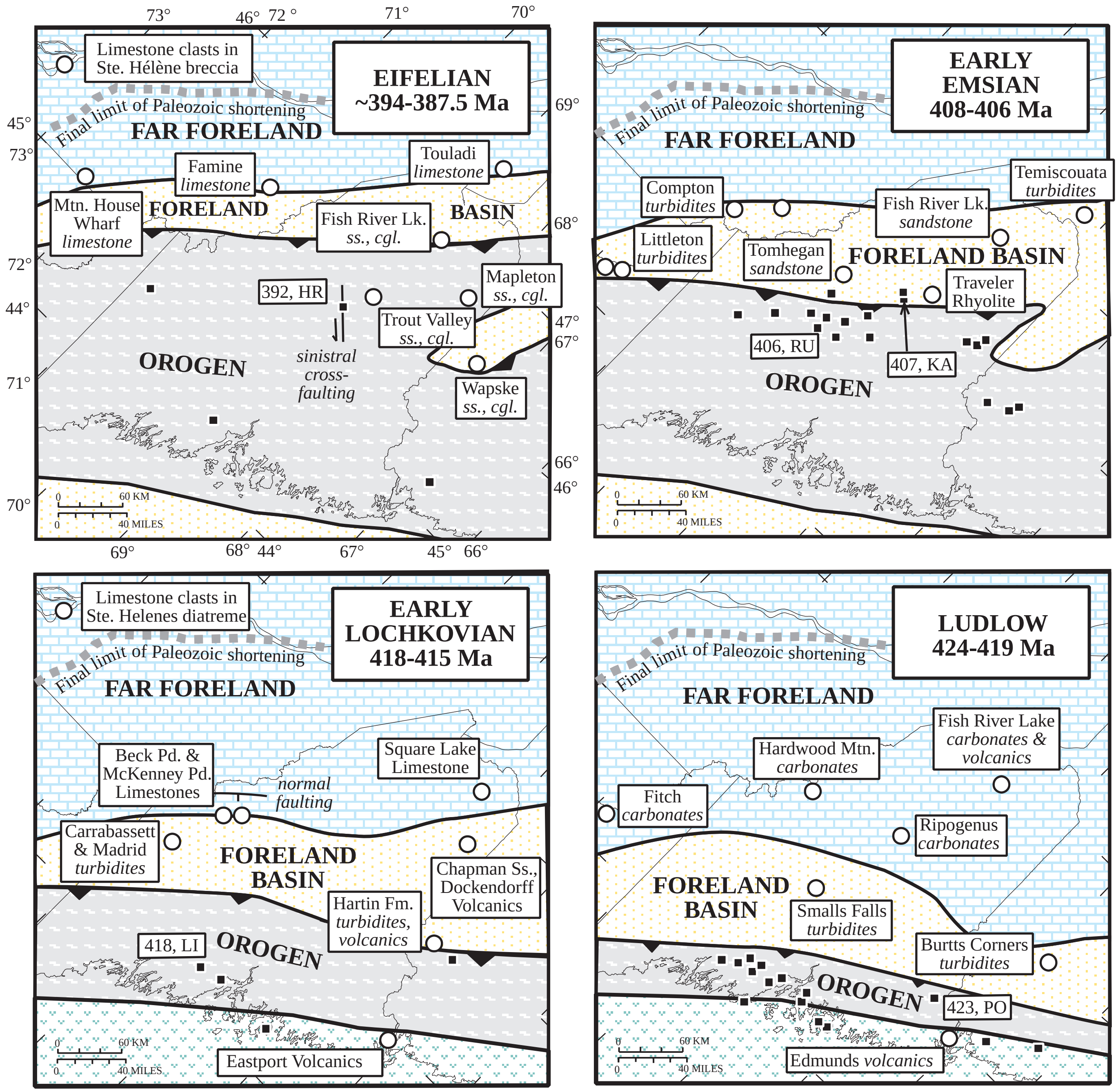

Figure 12.—Schematic, nonpalinspastic paleogeographic maps of study area (see fig. 1), showing migration of orogen and foreland basin during four time slices. Circles, locations of fossils in stratified rocks of appropriate age range, except for the Traveler Rhyolite, which is dated isotopically; black squares, locations of plutons in appropriate age range. Boxes show age and two-letter abbreviation for a few selected plutons, keyed to figure 6 and table 3. Toothed line denotes Acadian deformation front, which probably was a blind thrust rather than an emergent thrust during much of Late Silurian to Middle Devonian time. Paleogeography during the Eifelian and Emsian is unknown for area south of orogen (present coordinates).

Table 7.—Data used to estimate Acadian and younger shortening in figure 13.

Map letter (fig. 15)	Azimuth of cross section	Location	Stretch	1/stretch (plotted in fig. 13)	Final width (km)	Initial width (km)	Notes	Reference
A	130°	Spencer Stream 15′ quadrangle	0.75	1.33	6.5	8.6	Folded contact at base of the Seboomook Group.	Boucot and Heath (1969, pl. 14), part of section U–U′.
B	133°	Long Pond and Brassua Lake 15′ quadrangles	.84	1.19	18.4	21.9	Folded contact between the Seboomook Group and the overlying Tarratine Formation.	Boucot and Heath (1969, pl. 14), part of section L–L′.
C	143°	Brassua Lake 15′ quadrangle	.68	1.47	6.8	10	Folded contact between the Tarratine Formation and the overlying Tomhegan Formation (including the Kineo Rhyolite).	Boucot and Heath (1969, pl. 14), part of section F–F′.
D	090°	Caucomgomoc Lake 15′ quadrangle	.55	1.82	9.6	17.5	Folded contact between the Northeast Carry Formation and overlying strata of the undivided part of the Seboomook Group.	Pollock (1985), part of section D–D′.
E	100°	Lobster anticline and Roach River syncline	.77	1.30	18.9	24.6	Folded contact at base of Silurian rocks.	Boucot and Heath (1969, pl. 14), part of section D1–D1′.
F	129°	Munsungun anticlinorium	.51	1.96	11.2	21.8	Folded base of the Spider Lake Volcanics.	Hall (1970), part of section A–A′.
G	140°	Phillips 15′ quadrangle	.47	2.13	5.3	11.3	Folded contact between the Smalls Falls and the Madrid and Carrabassett Formations.	Moench (1971), section B–B′.
H	140°	Kingsbury 15′ quadrangle	.38	2.63	5.4	14.1	Folded contact between the Madrid and Carrabassett Formations. Ignoring parasitic folds, stretch is 0.49 km, and initial width is 10.9 km.	Ludman (1978), part of section A–A′.
I	138°	Shin Pond 15′ quadrangle	.77	1.30	13.8	17.8	Folded contact between unnamed Silurian strata and the Seboomook Group.	Neumann (1967), part of section A–A′.
J	145°	Island Falls 15′ quadrangle	.34	2.94	5.6	16.5	Folded contact between the Mattawamkeag Formation and the overlying Allsbury Formation (to the west) and "rocks of Island Falls" (to the east).	Eckren and Frischknecht (1967), part of section A–A′.
K	150°	Maple and Hovey Mountain area	.58	1.72	.9	1.6	Folded contact of base of manganiferous beds. Constrained by exploratory drilling.	Pavlides (1962), section A–A′.
L	090°	Presque Isle 15′ quadrangle	.71	1.41	8.7	12.25	Weighted average of measurements across the Chapman syncline (base of unnamed Silurian limestone) and the Dudley syncline (base of the Perham Formation). Post–Mapleton stretch is 0.86 km.	Boucot and others (1964), parts of section A–B.

have suggested (McKerrow and Ziegler, 1971; Bradley, 1983). Nonetheless, they are linked by the chain of provenance relationships discussed above.

Second, evidence for the amount of fold-related Acadian shortening can be gleaned from some 1:62,500-scale quadrangle maps. We estimated shortening on 15 cross sections that were selected because they show at least one readily mapped contact, such as a carbonate-flysch contact. Using the sinuous-bed-length method, the ratio of the palinspastic cross-sectional length to length in the deformed state was found to range from 1.30 to 2.78 (fig. 13), with a mean of 1.83. These results, if taken as being representative of the entire orogen, suggest that the pre-Acadian across-strike width of figure 1 north of the Sennebec Pond fault was about twice the present width. The actual initial width was most likely even greater because we were able to estimate shortening only for those exceptional places with simple, traceable structure. In the southeasterly part of the Central Maine Basin in Maine (Osberg, 1988) and in the northwestern part of the same basin in New Hampshire (Eusden and others, 1996), early recumbent folds imply much greater amounts of shortening. Our estimate of shortening also does not account for the effects of penetrative strain.

Third, the only potentially problematic post-Acadian strike-slip fault in the study area (fig. 1) is the Devonian to Carboniferous Norumbega Fault System. The displacement on this fault system may be large, but not so large that the rocks on either side are unrelated. Near the Maine-New Brunswick border, turbidites assigned to the Lower to middle Silurian Flume Ridge Formation occur at the same chlorite grade both northwest and southeast of the fault (Ludman, 1981). Map relations reviewed by Ludman and West (1999) suggest 125 to 140 km of cumulative dextral displacement across the Norumbega Fault System.

In summary, the successive deformation fronts mapped in figure 10 are probably more than twice as close together as they would be on an accurate palinspastic base. Also, localities outboard of the Norumbega Fault System were probably about 125 to 140 km farther northeast during the Acadian orogeny. Because the Norumbega Fault System approximately parallels Late Silurian paleogeographic belts, however, it does not significantly affect any of the conclusions presented below.

Early Ludlow

The Acadian orogeny was underway by the beginning of the Ludlow. The deformation front in figure 10 was placed between the posttectonic Pocomoonshine pluton at locality 37 (fig. 1) and the graptolite-bearing Smalls Falls Formation and

Burtts Corners beds at localities 29 and 35. The boundary between the foreland basin and far foreland was placed northwest of localities 29 and 35 but southeast of such shallow-marine carbonate units as the Hardwood Mountain, Spider Lake, Ripogenus, Grand Lake Seboeis, and Fish River Lake Formations (at locs. 11, 18, 16, 20, and 21, respectively).

Although our main emphasis in this report has been on tracking the northwestward migration of the Acadian deformation front, we also note evidence that a second, outboard boundary between deformed and undeformed rocks migrated a much shorter distance to the southeast. In the Coastal volcanic belt, volcanism and sedimentation continued without interruption through most of the Silurian and into the Lochkovian. The southeast boundary between deformed and undeformed rocks must have lain somewhere between the Pocomoonshine pluton and locality 38 (fig. 1) by early Ludlow time. It does not appear to have moved southeast until after Lochkovian time—that is, after deposition of the youngest Acadian-deformed strata at Eastport (loc. 38).

Early Lochkovian

During early Lochkovian time, the Acadian deformation front was located along the northwest flank of the Central Maine Basin and near the eastern margin of the Aroostook-Matapedia Basin. Its position is bracketed by foreland-basin and far-foreland strata to the northwest and syntectonic and posttectonic plutons to the southeast. Plutonic rocks within the range 418.5 to 415 Ma delimit the zone of active deformation; these rocks include the syntectonic Lincoln sill (417 Ma) and the Hartfield pluton (415 Ma).

A Lochkovian foreland basin flanked the plutonic belt on the northwest. Strata assigned to this stage of the migrating foreland basin include the Hartin Formation (loc. 34, fig. 1), the Chapman Sandstone and Swanback Formation (loc. 31), a part of the Millimigassett Lake Formation (loc. 20), and the Madrid and Carrabassett Formations in the Central Maine Basin (for example, at locs. 26 and 30). The early Lochkovian position of the leading edge of the foreland basin (fig. 11) is

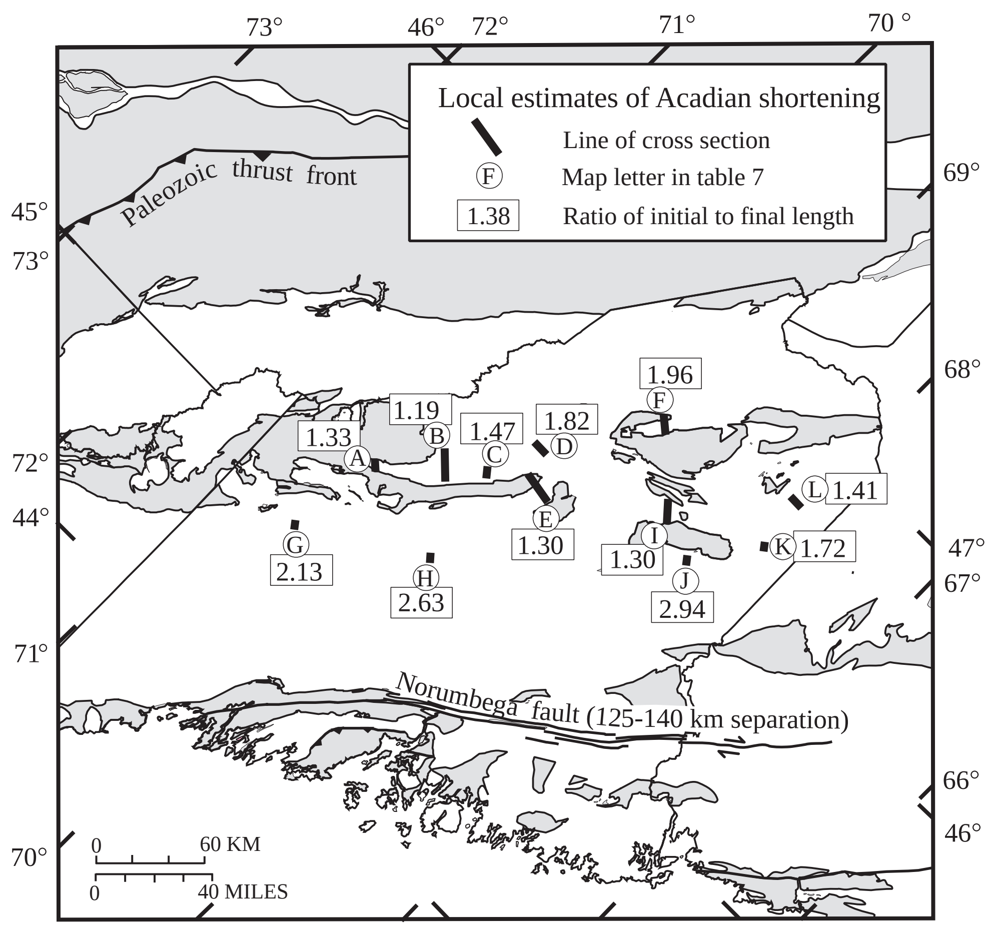

Figure 13.—Maine and adjacent parts of New Brunswick, Québec, New Hampshire, and Vermont, showing local estimates of Acadian and younger shortening, determined by sinuous-bed-length method, across 12 published cross sections. Boxed numbers are ratios of initial to final cross-sectional lengths; for example, a value of 2.63 corresponds to a cross section that is now 5.4 km long but is inferred to have been 14.1 km long, by unfolding one or more marker horizons. Data are listed in table 7.

accurately placed at localities 12 and 13, where the Beck Pond and McKenney Ponds Limestones grade into or interfinger with the Seboomook Group and Tarratine Formation, respectively. Carbonate rocks of Lochkovian age include parts of the lower Fish River Lake Formation (loc. 21), Spider Lake Formation (loc. 18), and Square Lake Limestone (loc. 23). These rocks presumably were deposited beyond the limits of the Lochkovian foreland basin.

Pragian

Two lines of evidence show that the Pragian foreland basin may have migrated 15 to 20 km (present distance) cratonward from its early Lochkovian position (fig. 11). Along the Munsungun anticlinorium (loc. 16, fig. 1), the Pragian Seboomook flysch encroached over Lochkovian far-foreland carbonates. Similarly, in the Pennington anticlinorium (loc. 23), Pragian and younger flyschlike siltstone of unit Dus overlies the Square Lake Limestone. Within the foreland basin itself, the deltaic Matagamon Sandstone prograded over prodelta turbidites of the Seboomook Group. Although we have no direct evidence for migration of the deformation front from its Lochkovian position, in constructing figure 10, we have assumed that the deformational front migrated the same distance as the foreland-basin margin did from its previous position (fig. 11). This assumption seems reasonable because the deformation front and foreland-basin front did migrate in tandem during every interval, such as Ludlow to Lochkovian, for which have adequate data.

Early Emsian

The clearest picture of Acadian paleogeography in Maine is in the early Emsian, at 407 to 406 Ma (Bradley and others, 1996). We can identify a foreland basin and orogenic wedge. Superimposed on both basin and wedge and straddling the deformation front was a volcanic-plutonic belt. Many Emsian plutons truncate Acadian map-scale structures, including the newly dated Sebec Lake, Russell Mountain, and Shirley-Blanchard (SL, RU, and SB, respectively, fig. 6). The deformation front was plotted to the northwest of such plutons. The early Emsian foreland basin includes the younger strata of the vast "Devonian slate belt" of Boucot (1970). Foreland-basin flysch units that have yielded Emsian fossils include the Temiscouata Formation (loc. 24, fig. 1), the Compton Formation (locs. 3, 4), and the Littleton Formation (locs. 9, 10). It seems likely, though unproved, that parts of the Seboomook Group in the Connecticut Valley-Gaspé Basin in northwestern Maine and the Gile Mountain Formation of Vermont are also Emsian. Early Emsian molasse deposits in the study area (fig. 1) include the Tomhegan Formation (loc. 15) and, as suggested above, the uppermost part of the Matagamon Sandstone (loc. 17). These molasse facies must have been laid down near the proximal, southeastern margin of the foreland basin. The cratonward boundary of the early Emsian foreland basin can be positioned by using negative evidence. As discussed below, the three occurrences of Eifelian limestone along the northwestern margin of the Connecticut Valley-Gaspé Basin rest everywhere on pre-Emsian rocks, and so Emsian foreland-basin deposits—which are kilometers thick where present—probably were never deposited that far northwest.

The deformation front must have corresponded to the Lobster Mountains anticlinorium (LMA, fig. 2), between the site of Tomhegan molasse deposition and the site of the plutons that were emplaced into the deforming or recently deformed orogenic wedge. The Moxie and Katahdin plutons (KA and MX, respectively, fig. 6) were emplaced in the general vicinity of the deformation front (Bradley and others, 1996).

Late Emsian to Early Eifelian

By the time of the Emsian-Eifelian boundary (~394 Ma), the Acadian deformation front had migrated some distance cratonward, to the approximate position of the Pennington anticlinorium (PA, fig. 2). Southeast of the deformation front (fig. 10), late Emsian and younger strata postdate Acadian deformation. To the northwest, foreland-basin and far-foreland deposits of this age range predate Acadian deformation. At locality 22 (fig. 1), the upper siliciclastic part of Boone's (1970) Fish River Lake Formation ("Nadeau Thoroughfare beds") were deposited in a foreland-basin setting. At localities 2, 5, and 7, Eifelian limestone was deposited in a far-foreland setting. The Mapleton Formation and Trout Valley Formation are similar in age to the upper part of the Fish River Lake Formation. Both the Mapleton and Trout Valley Formations, however, postdate the main phase of Acadian deformation, and so we place them southeast of the deformation front.

The stratigraphy of the Carlisle area (loc. 33, fig. 1) complicates an otherwise-straightforward paleogeographic picture. Though coeval with the Mapleton Formation, the Wapske Formation was involved in the same regional "Acadian" folding as the rest of the Silurian to Devonian section. One possible explanation is that the Carlisle area escaped Acadian deformation until the late Eifelian or later (figs. 10, 12). A second possible explanation is that the first phase of Acadian deformation, which in the Presque Isle area created the unconformity below the Mapleton Formation, bypassed the Carlisle area entirely. In this scenario, the folding at Carlisle would be related to the same, later event that caused gentle, late-stage or post-Acadian folding of the Mapleton Formation at Presque Isle. A third possible explanation is that the Wapske Formation was deposited farther to the northwest and was thrust into its present position during a late-stage Acadian event.

Late Eifelian to Early Givetian

By the time of the Eifelian-Givetian boundary (~387.5 Ma), the Acadian deformation front had again migrated cratonward from its previous position, especially on the basis of a belt of plutons in the Eastern Townships of Québec. These plutons intrude already-deformed Silurian and Devonian strata of the

Connecticut Valley-Gaspé Basin; the oldest and, thus, most significant pluton is the Scotstown pluton (384±2 Ma; SC, fig. 6; Simonetti and Doig, 1990). The deformation front must have been situated somewhere to the northwest by about 384 Ma.

Just a few million years before emplacement of the Scotstown pluton, Eifelian carbonates were still being deposited at locality 5 (fig. 1). These strata are penetratively deformed, indicating that the final cratonward limit of *Acadian* deformation front was at least that far northwest. (The *Paleozoic* deformation front shown on figure 1 could be as old as Ordovician.) Exactly when the deformation took place is unknown, but it was likely at least a few million years after deposition of the last carbonates (the youngest being the Famine Limestone, loc. 5), long enough to allow for an interval of foreland-basin sedimentation before deformation.

The late Eifelian to early Givetian stratigraphic record in the study area (fig. 1) is confined to a few clasts in the Ste. Hélène Island breccia, near Montreal (loc. 1). The clasts include calcareous siltstone, sandstone, and some limestone of late Eifelian to early Givetian age (Boucot and others, 1986). By comparison of these rock types with the Catskill clastic wedge in New York, Montreal was probably near the cratonward edge of the Acadian foreland basin at about the Eifelian-Givetian boundary (fig. 11).

Earliest Frasnian

The deformation front at the beginning of Frasnian time has been plotted in figure 10 on the basis of slender evidence. Deformed Eifelian carbonates at localities 2, 5, and 7 (fig. 1) show that the Acadian orogeny was felt at least that far to the northwest. The deformation front was extrapolated from its position at the beginning of the Givetian, assuming a nearly constant convergence rate. The final northwestward limit of Acadian shortening is unknown; it might correspond to the Paleozoic thrust front (fig. 10) or, alternatively, might lie somewhere between the Frasian deformation front and the Paleozoic thrust front. Along the Paleozoic deformation front near Albany, N.Y., strata as young as Eifelian were involved in thrust faulting, probably Acadian but conceivably Alleghanian.

Implications

Age of the Acadian orogeny

A zone of active deformation migrated across Maine and adjacent parts of New England and Canada from early Ludlow through Givetian time. Initially, the orogen was a narrow belt, but it widened over time, eventually encompassing the entire State of Maine (fig. 10). The overall pattern of cratonward migration of the orogen is similar to that suggested by Donahoe and Pajari (1973) but is much changed in light of our new data from Maine and our improved time scale.

The Acadian orogeny has traditionally been regarded as a Devonian event. During the 1990’s, however, new pluton ages were interpreted to show that the supposedly Devonian orogeny in parts of the northern Appalachians was actually Silurian (Bevier and Whalen, 1990b; West and others, 1992, 1995; Stewart and others, 1995a). Now, in light of revisions to the time scale that fix the Silurian-Devonian boundary at about 418 Ma, some of the purportedly Silurian plutons are back once again in the Early Devonian (for example, the North Pole Stream and Mount Elizabeth plutons of Bevier and Whalen, 1990b). The rest fall within the final 5 m.y. of the Silurian.

The northwestward migration of the deformation front (fig. 10) might be interpreted in two ways. One interpretation is that the orogeny occurred in pulses, each pulse separated by tectonic stability and each with some different cause. Some workers (for example, Boucot and others, 1964; Stewart and others, 1995a) have applied the term “Salinic” to Late Silurian orogenic events in Maine—events that, in their view, are distinct from the Devonian Acadian orogeny. Similarly, other names for orogenies might be coined for other positions of the deformation front (fig. 10). We prefer to explain the pattern in figure 10 as the result of a single plate-convergent event which swept the region over a long period of time. Existing data cannot resolve this question.

Rate and Trajectory of Deformation-Front Migration

In figure 10, the deformation front is shown to have migrated about 240 km across strike in about 40.5 m.y. Assuming continuous rather than episodic convergence, this implies a rate of about 6 km/m.y. (0.6 cm/yr). The actual rate must have been considerably faster because, as noted above, the base map is nonpalinspastic. Assuming, for the sake of discussion, that Acadian shortening reduced Maine to half its original across-strike width, the rate of thrust-front migration would be about 12 km/m.y. (1.2 cm/yr).

It is likely, but not assured, that the rate of thrust-front migration approximately equaled the rate of plate convergence (Bradley and Kusky, 1986; Bradley, 1989). This is generally the case at oceanic subduction zones. Similarly, in the snowplow analogy for orogenic wedges (wherein a snowplow pushes a wedge of snow ahead of the plowblade), the deformation front in the snow migrates at about the same rate as the plow advances. A less likely alternative is that the advance of Acadian thrusts was directly linked not to plate convergence but to extensional orogenic collapse, as in the Carpathians (Royden and Burchfield, 1989). Although little is yet known about Acadian hinterland tectonics between about 420 and 380 Ma (see section below entitled “Hinterland Deformational Regimes…”), there is no evidence for wholesale regional extension of the magnitude required by this mechanism.

On the basis of the along-strike diachronism of foreland-basin deposits in the northern Appalachians (younger toward the southwest), Bradley (1987, 1997) suggested that Acadian plate convergence had a dextral component. The migration of black-shale depocenters across the Acadian foreland basin in the central Appalachians led Ettensohn (1987) to the same conclusion. The present study area does not extend far enough along strike to provide clear evidence either for or against along-strike diachronism.

Physical Setting of Magmatism

One by-product of this study has been to show that many of the Acadian plutons of Maine occur in distinct age belts (fig. 6). Moreover, figure 12 shows how the various plutons and volcanic rocks were spatially related to the orogen as it existed at the time of magmatism.

Coastal Volcanic Belt and the Ludlow to Lochkovian Plutonic Belt

On geochemical grounds, Gates and Moench (1981), Hogan and Sinha (1989), and Seamans and others (1999) have suggested that the Ludlow to Lochkovian volcanic rocks and plutons of coastal Maine formed in an extensional setting. The migration pattern of the deformation front (figs. 10, 12), however, implies that magmatism took place in an overall setting of plate convergence. The same conclusion is suggested on a more local scale by the fact that the Lochkovian Lincoln sill was emplaced synkinematically (Tucker and others, in press). Judging from figure 12, magmatism took place in the upper plate of a two-plate system, presumably above an A-type subduction zone.

Piscataquis Volcanic Belt and the Emsian Plutonic Belt

The cause or causes of Ludlow to Emsian volcanism along the Piscataquis volcanic belt have long been debated, and although a resolution is beyond the scope of this report, our findings do provide an important new constraint: all of the volcanic rocks are broadly syncollisional and were erupted in a foreland setting. Most of the Ludlow to Lochkovian volcanic rocks (East Branch Group, West Branch Volcanics, and Fish River Lake, Spider Lake, Grand Lake Seboeis, and Allagash Lake Formations) were erupted in a *far*-foreland position. However, several volcanic units of Lochkovian and Emsian age are interbedded with or grade laterally into foreland-basin deposits and thus formed in a *proximal*-foreland position (Hedgehog Volcanics, Edmunds Hill Andesite, Hartin Formation, Traveler Rhyolite, and Kineo Rhyolite). Three alternative interpretations have been advanced for part or all of the Piscataquis volcanic belt. (1) On the basis of basalt discriminant plots, Hon and others (1992) and Keppie and Dostal (1994) interpreted the Spider Lake and West Branch volcanic rocks as having formed in an intracontinental-rift setting, related to so-called “transpressive rifting.” In light of our findings, any extension would necessarily have taken place in a collisional foreland setting, such as might have been caused by lithospheric flexure (Bradley and Kidd, 1991) or by Baikal-style cracking of the lower plate (Sengör and others, 1978). We note that the geochemical signature of volcanic rocks in foreland settings has not been adequately characterized and that discriminant plots used by the above-cited authors do not even identify fields for volcanic rocks formed in this tectonic setting. (2) Tucker and others (in press) have suggested that the Emsian foreland-basin rhyolites and the Emsian plutons just to the southeast were a product of lower-lithospheric delamination of the downgoing (North American) plate, which was being subducted toward the southeast. (3) Bradley (1983) interpreted these units as arc volcanic rocks formed over a northwest-dipping subduction zone. This model would explain magmatism in all three settings—far foreland, proximal foreland, and orogenic wedge—as the consequence of an orogenic wedge above a southeast-dipping subduction zone colliding with (and overriding) a magmatic arc formed over a northwest-dipping subduction zone. (An updated version of this tectonic model was illustrated by Hanson and Bradley, 1989.)

Syncollisional Versus Precollisional Plate Geometry

Tectonic interpretations of the Acadian orogeny have long been debated. Part of the controversy regarding Acadian plate geometry may be sidestepped for present purposes by separately considering the plate geometry during two intervals—before initial impact of the converging plates and during the actual collision. The paired migration of the orogen and foreland basin shown in figures 10 through 12 suggests that, during collision, an upper plate consisting of the growing orogenic wedge and its Avalonian backstop overrode the Taconic-modified margin of North America. The implied A-type subduction zone that operated during collision presumably evolved naturally out of a B-type subduction zone of the same southeast-dipping polarity, where ocean floor that once existed between Avalonia and North America was consumed during the late Llandovery and Wenlock.

What remains to be settled is the nature of the northwestern margin of the Central Maine Basin before collision. This is a much-debated topic, and we, the coauthors, are ourselves divided on the subject. In the two main tectonic models, both of which have variants, it was either (1) a “passive” or rifted margin on the backside of an arc that had collided with North America during Ordovician time (for example, Robinson and others, 1998; Tucker and others, in press), or (2) a convergent margin that took up some of the motion between Avalonia and North America during Silurian time (Bradley, 1983; Eusden and others, 1996). In the second model, the northwestern subduction complex and arc were overridden by, and downflexed beneath, the southeastern one described in the preceding paragraph (Bradley, 1983). In the one-subduction-zone model (Robinson and others, 1998), the northwestern passive margin was overridden by, and downflexed beneath, a southeastern subduction-collision complex. Although the two models differ significantly up until the time of collision, they are much the same afterward.

Hinterland Deformational Regimes and Regional Partitioning of Acadian Deformation

The present study has focused on tracking the first Acadian contractional deformation. In most places, however, Acadian deformation was polyphase. Figure 10 provides a framework for understanding those post-D1 events that occurred from about 423 to 382.5 Ma—that is, the ages of the oldest and youngest deformation fronts. Three examples will be discussed here.

Stewart and others (1995a) described a complex history of motion on a network of northeast-striking dextral faults in Penobscot Bay. The oldest of these, the Penobscot Bay-Smith Cove-North Blue Hill Fault, displaces isograds around the Segwick pluton (419.5±1 Ma) but is cut by the South Penobscot Intrusive Suite (419.2±2.2 Ma). Later, the South Penobscot Intrusive Suite was cut by the dextral Turtle Head Fault, which was intruded, in turn, by the Lucerne pluton (380±4 Ma; Wones, 1991). Where plate convergence has an oblique component, deformation can be partitioned into thrusting in the orogenic wedge and strike-slip faulting farther to the rear (Dewey, 1980). If this explanation is applicable to the latest Silurian episode of dextral strike-slip faulting on the Maine coast, subduction of Central Maine Basin crust beneath the Coastal volcanic belt would have had a dextral-oblique component.

Near Waterville (WA, fig. 6), a 399-Ma intermediate dike cuts the first-phase Acadian folds but is deformed by a second generation of folds that must be older than a static metamorphic event dated at 380 Ma (Osberg, 1988; Tucker and others, in press). The second folding must therefore have taken place when the Acadian deformation front was somewhere cratonward of its early Emsian position (fig. 10), perhaps 100 km to the northwest of Waterville. Accordingly, this contractional deformation was an out-of-sequence, hinterland event.

The Horserace pluton is a small body with a U-Pb age of 392 Ma (R.D. Tucker, unpub. data) within the 407-Ma Katahdin batholith. It was emplaced, probably synkinematically (Hon, 1980, p. 73), along a northwest-striking high-angle fault (the West Branch fault) that sinistrally offsets the western margin of the Katahdin pluton. This is one of many crossfaults that disrupt the bedrock map pattern in northern and central Maine (Osberg and others, 1985). Referring to figures 10 through 12, at about 392 Ma, the deformation front lay perhaps 70 km to the northwest. Like the second folding at Waterville, sinistral motion on the West Branch Fault took place in the Acadian hinterland, while the orogen was still advancing toward North America.

Foreland Deformational Regimes

Figures 10 through 12 also provide insights into pre-Acadian events that took place in what was then the orogenic foreland. Two northeast-striking high-angle faults at Spencer Mountain (loc. 12, fig. 1) cut the Hardwood Mountain Formation (Burroughs, 1979), to which we assign a late Ludlow to Pridoli age (table 2). The map pattern indicates that one of these faults is overlain by the Hobbstown Conglomerate, which ranges from possibly as old as Pridoli to definitely as young as Pragian (Boucot and Heath, 1969). The sense of motion is not well established. On the basis of the relative age of faulted strata, Burroughs (1979) showed both faults as having down-to-the-northwest motion. Whatever the exact age of faulting and the sense of motion, it is clear that faulting took place in a foreland setting and that the high-angle faults paralleled the approaching orogenic front. On this basis, they would appear to resemble normal faults of the Taconic foreland of New York, which formed in response to lithospheric flexure (Bradley and Kidd, 1991).

Suggestions for Further Work

Since an influential paper by Naylor (1970), the Acadian has commonly been described as “an abrupt and brief event.” This description appears to be only half-true. Although the first phase of deformation was, indeed, abrupt, lasting but a few million years at any given place, it took some 40 m.y. for the deformation front to make its way across Maine. Our study thus confirms the findings of Donahoe and Pajari (1973), in general if not in particular.

Figures 10 through 12 could be improved with better age control for several key units. The Flume Ridge, Madrid, and Carrabassett Formations each have a particular tectonic significance but are so loosely dated by fossils as to be barely useful for the present purposes. A dedicated search for fossils, especially palynomorphs, and (or) datable pyroclastic horizons in these units would be worth the effort; tuff horizons, 1 to 2 cm thick, commonly yield zircons, and many of these horizons have probably been overlooked. Tighter paleontologic control is also needed, though less urgently, for the Hartin, Perry, and Swanback Formations, the Chapman Sandstone, and the Seboomook Group in northernmost Maine. Likewise, modern geochronologic studies are needed for the West Branch, Spider Lake, and Hedgehog Volcanics, the Red Beach and Lucerne plutons in Maine, and the many Devonian plutons in the Northeast Kingdom of Vermont.

The Devonian time scale still has significant holes to be filled in the Pragian, late Emsian, and Givetian; until these intervals are better calibrated, correlations between isotopically and paleontologically dated rocks in these age ranges will remain equivocal.

The depositional environments, paleocurrents, and provenance of unnamed unit Dsus and the Sangerville, Smalls Falls, Hartin, Wapske, Fish River Lake, and Swanback Formations warrant detailed study; we have assigned these units to a foreland-basin setting on the basis of lithology and regional relations alone. The Wapske is of special interest because it was evidently deformed about 15 m.y. later than rocks along strike in Maine (fig. 10). Accordingly, the Wapske Formation may have been deposited in a “piggyback” basin, a suggestion that might be evaluated by using detailed sedimentologic studies.

Our assessments of the pretectonic, syntectonic, or post-tectonic age of plutons are admittedly cursory, based as they are on anecdotal observations of a few thin sections and one or two traverses from country rock into pluton. Detailed investigations would be especially useful for those plutons, such as the Moxie and Katahdin, that appear to have been emplaced very close to the deformation front.

The present study has emphasized evidence bearing on the northwestward migration of the Acadian orogen, but we have barely touched on its migration to the southeast. Did it migrate in a stepwise fashion, or gradually? Although the sedimentary record from the outboard side of the orogen is comparatively sparse, the Eastport Formation of coastal Maine and New Brunswick may preserve a record of synorogenic sedimentation off the backside of the growing Acadian mountains.

A palinspastic base will be essential for discerning the geographic relations between correlative rocks at the time they were formed and for quantifying the rate of plate convergence during collision.

Finally, one of the most intriguing recent developments in geodynamics involves previously unappreciated links between climate and tectonics. Judging from studies of the Southern Alps of New Zealand (for example, Beaumont and others, 1992), synorogenic rainfall and consequent erosion can profoundly affect the evolution of the underlying deforming rocks. Hoffman and Grotzinger (1993) suggested that an asymmetric orographic climate can influence the metamorphic field gradient, the width of the thrust belt, the extent of footwall involvement in deformation, and the thickness and compositional maturity of the foreland-basin fill. Acadian synorogenic climate and its possible influences on structure, metamorphism, and stratigraphy have not been investigated, but there is good reason to do so. Early Devonian paleogeographic reconstructions imply that the Acadian mountains had an asymmetric orographic climate, with the west side wetter (Witzke and Heckel, 1988). Such a climatic asymmetry, if borne out by paleoclimatologic studies of Devonian strata in the study area, might be explored as a possible control on the migration of the main Acadian deformation front.

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PLATES 1 AND 2

[Contact photographs of the plates in this report are available, at cost, from the U.S. Geological Survey Library, Federal Center, Denver, Colorado 80225]

Plate 1.—Late Silurian to early Middle Devonian (late Ludlow to Eifelian) conodonts from Maine and southern Quebec, Canada. Scanning-electron photomicrographs. Illustrated specimens are reposited in the U.S. National Museum (USNM), Washington, D.C. See text figure 1 for geographic location (locality no.), text figure 4 for stratigraphic position, and table 2 for faunal assemblage, age assignment, biofacies, and conodont-alteration index.

Figures 1–13. Beck Pond Limestone, locality 12. USGS colln. 12518–SD. Magnification, 50×.

- 1–6. *Ozarkodina remscheidensis eosteinhornensis* (Walliser), Pa, inner (figs. 1, 3) and outer (fig. 2) lateral views, Pb, M, and Sc (lateral views) elements. USNM nos. 508893–508897.
- 7, 8. *Dvorakia* sp., Sb element, posterior and anterior views. USNM no. 508898.
- 9, 10. *Icriodus* sp. indet., coniform Sc element, outer and inner lateral views. USNM no. 508899.
- 11, 12. *Decoriconus fragilis* (Branson and Mehl), inner and outer lateral views. USNM no. 508900.
- 13. *Belodella* sp., Sa element, lateral view. USNM no. 508901.

14–18. McKenney Ponds Limestone, locality 13. USGS colln. 12515–SD.

- 14, 15. *Pseudooneotodus beckmanni* (Bischoff and Sannemann), upper and lateral views. Magnification, 75×. USNM no. 508902.
- 16–18. *Icriodus* sp. indet., I elements (incomplete), upper views. Magnifications, 40× (figs. 16, 17) and 35× (fig. 18). USNM nos. 508903–508905.

19–26. McKenney Ponds Limestone, locality 13. USGS colln. 12516–SD.

- 19–24. *Icriodus* sp. indet., coniform elements. Magnification, 40×.
  - 19, 20. Sa element, opposite lateral views. USNM no. 508906.
  - 21, 22. M? element, inner and outer lateral views. USNM no. 508907.
  - 23, 24. Sd element, opposite lateral views, USNM no. 508908.
- 25, 26. *Ozarkodina remscheidensis* (Ziegler), Pa elements, inner lateral views. Magnification, 50×. USNM nos. 508909–508910.

27, 28. *Ozarkodina remscheidensis* (Ziegler), Pa elements, inner lateral and upper views. Magnification, 50×. USNM nos. 508911–508912, McKenney Ponds Limestone. USGS colln. 12517–SD.

29–31. Mountain House Wharf Limestone, locality 2. *Icriodus* sp. indet., I elements, upper views. Magnification, 45×.

- 29. USNM no. 508913, USGS colln. 12527–SD.
- 30, 31. USNM nos. 508914–508915, USGS colln. 12526–SD.

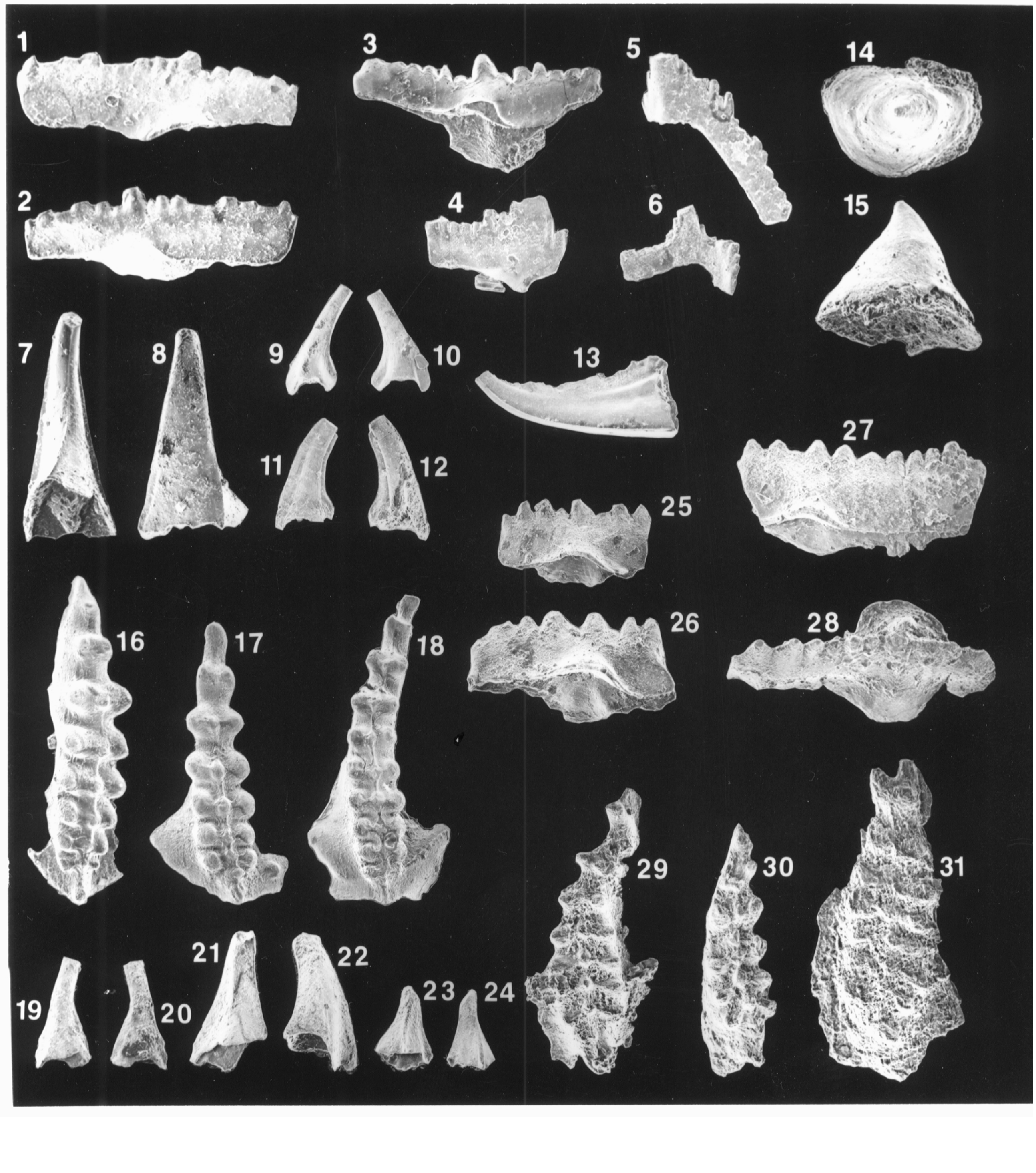
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Plate 2.—Late Silurian to earliest Devonian (late Ludlow to early Lochkovian) conodonts from Maine. Scanning-electron photomicrographs; illustrated specimens are reposited in the U.S. National Museum (USNM), Washington, D.C. See text figure 1 for geographic location (locality no.), text figure 4 for stratigraphic position, and table 2 for faunal assemblage, age assignment, biofacies, and conodont-alteration index.

Figures 1-5. Ripogenus Formation, locality 16. USGS colln. 12513-SD.

- 1, 2. *Dvorakia* sp. indet., inner and outer lateral views of Sb element. Magnification, 50×. USNM no. 508869.
- 3, 4. *Ozarkodina excavata* (Branson and Mehl), inner lateral views of Sb and Sc elements. Magnification, 50×. USNM nos. 508870-508871.
- 5. Digyrate apparatus, inner lateral view of Sb element. Magnification, 60×. USNM no. 508872.

6-9. Parker Bog Formation, locality 14. *Oulodus elegans*, inner lateral views of Pa, Pb, M, and Sb elements. Magnification, 60×. USNM nos. 508873-508876, USGS colln. 12521-SD.

10, 11. Parker Bog Formation, locality 14. *Pedavis* sp. indet., posterior and anterior views of coniform S element. Magnification, 75×. USNM no. 508877, USGS colln. 12521-SD.

12-20. Hardwood Mountain Formation, locality 11. USGS colln. 12523-SD.

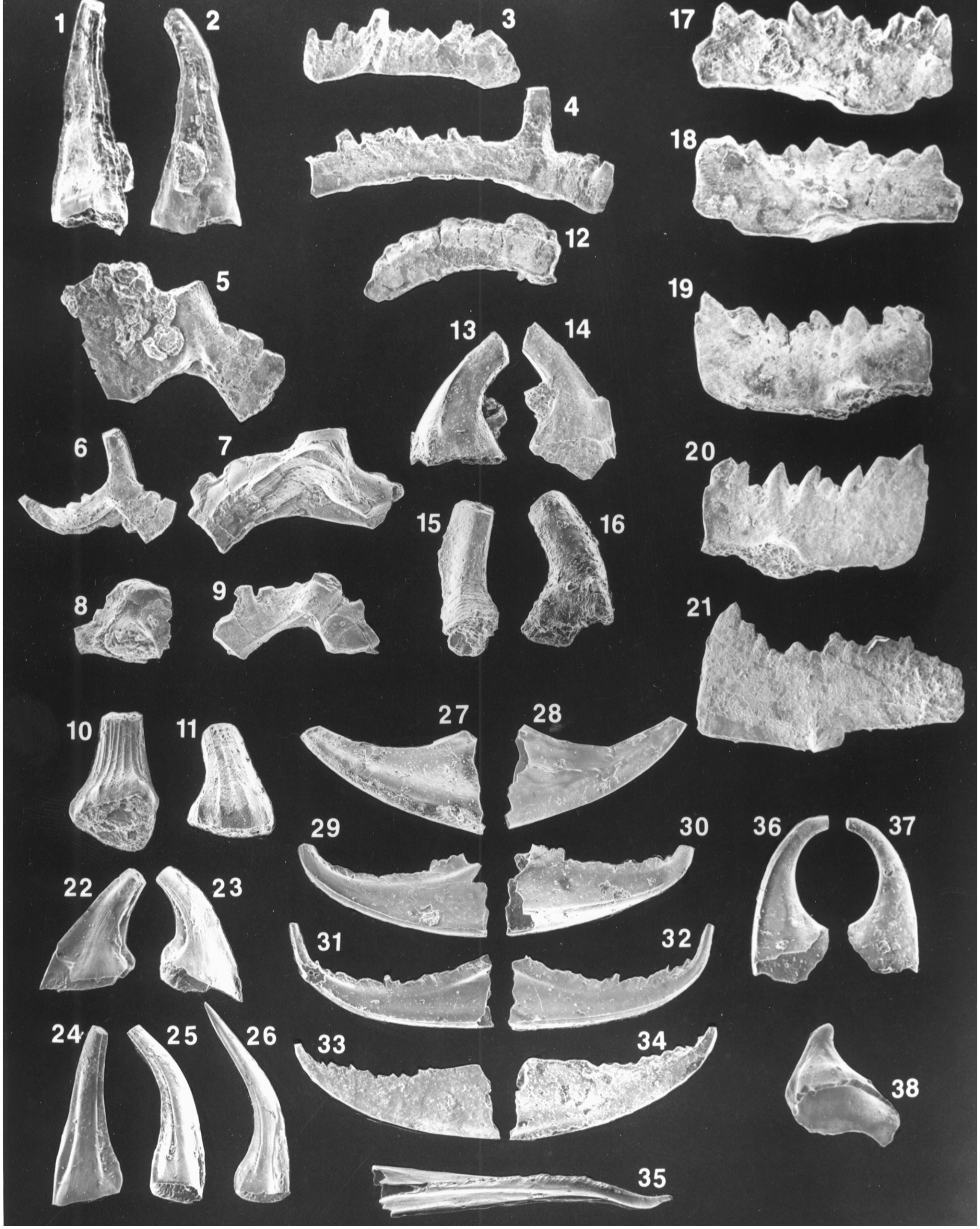
- 12. *Ozarkodina excavata excavata* (Branson and Mehl), Pa element, outer lateral view. Magnification, 50×. USNM no. 508878.
- 13-16. *Corryssognathus dubius* (Rhodes), Sb and Sc elements, outer and inner lateral and posterior and anterior views. Magnification, 50×. USNM nos. 508879-508880.
- 17-20. *Ozarkodina remscheidensis remscheidensis* (Ziegler), Pa elements, inner and outer lateral views. Magnifications, 45× (figs. 17, 18) and 50× (figs. 19, 20). USNM nos. 508881-508882.

21. Hardwood Mountain Formation, locality 11. *Ozarkodina confluens* Branson and Mehl, Pa element, inner lateral view. Magnification, 35×. USNM no. 508883, USGS colln. 12524-SD.

22-38. Square Lake Limestone, locality 23. USGS colln. 12511-SD.

- 22-26. *Decoriconus fragilis* (Branson and Mehl). Magnification, 75×.
  - 22, 23. Sc elements (23 lost after photography), inner lateral views, USNM no. 508884.
  - 24-26. Sb elements, inner and two outer lateral views of two specimens, USNM nos. 508885 and 508886.
- 27-35. *Belodella* cf. *B. resima* (Philip). Magnifications, 60× (figs. 27-34) and 110× (fig. 35).
  - 27, 28. M element, outer and inner lateral views, USNM no. 508887.
  - 29, 30, 35 Sb1 element, inner and outer lateral and posterior views. USNM no. 508888.
  - 31, 32. Sa element, opposite lateral views. USNM no. 508889.
  - 33, 34. Sc element, outer and inner lateral views. USNM no. 508890.
- 36, 37. *Dvorakia* sp., Sc element, inner and outer lateral views. Magnification, 50×. USNM no. 508891.

38. *Pseudooneotodus beckmanni* (Bischoff and Sannemann), oblique lower posterior view. Magnification, 75×. USNM no. 508892.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