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Plate Tectonic Model for the Evolution of the Central Andes

ABSTRACT

Data on the geophysics and geology of the central Andes are interpreted in terms of plate theory and a model for Andean evolution is presented. Analysis of upper mantle structure and seismicity shows that the underthrusting Pacific plate is now about 50 km thick and the overriding South American plate 200 to 300 km thick. Underthrusting of the Pacific plate probably began in Triassic time and has continued without substantial change to the present. Prior to underthrusting, the west coast of South America was quiescent, and great thicknesses of Paleozoic continental shelf deposits were laid down in an area east of the present volcanic arc. In Late Triassic or Early Jurassic time, an incipient arc began to form at or west of the present coast of South America. Igneous activity has since migrated eastward, culminating in the Pliocene-Pleistocene volcanic episode. The crust beneath the volcanic cordillera is more than 70 km thick and probably consists largely of rocks compositionally equivalent to those of the volcano-plutonic suites observed at the surface.

Increase in crustal volume of the volcanic arc between Cretaceous time and the present implies either that the mantle above the underthrust plate has undergone 18 to 36 percent partial melting or that 1 to 2 km of rock has been melted from the underthrusting plate. The intrusion of melt into the crust beneath the volcanic cordillera and the resultant crustal dilatation produced continentward compression of the Paleozoic sedimentary rocks which form an easterly belt of thrust and fold mountains. Here crustal shortening has produced crustal thicknesses of 50 to 55 km. Few deposits of the type normally termed eugeosynclinal, and no ophiolites, are observed between trench and volcanic arc; only in the intermontane foredeep behind the arc has a clastic wedge of geosynclinal proportions formed.

INTRODUCTION

The past decade has been one of revolution in the earth sciences, for the advent and general acceptance of the theory of sea-floor spreading and its corollary, plate tectonics, have led to new insights into many fundamental processes of geology and geophysics. Verification followed close upon the heels of the conceptual formulation of sea-floor spreading by Hess (1962), and the essential elements of the theory appear firmly established. More recently, the theory of plate tectonics has been advanced as a unifying explanation of observable global tectonic phenomena associated with sea-floor spreading (McKenzie and Parker, 1967; Morgan, 1968; Isacks and others, 1968; McKenzie and Morgan, 1969). According to this theory, the outer rind of the earth consists of rigid plates in motion relative to one another. These plates are generated at ocean ridges (or continental rifts) and consumed at trenches, and most global seismic and orogenic activity occurs at these plate junctures. The plates are perhaps as little as 0 to 10 km thick at the oceanic ridges and as much as 150 km thick or more far from the ridges. Where two plates collide, one is normally thrust under the other, and the upper part of the descending plate is marked by a Benioff zone. Although the theory of plate tectonics is by no means as secure as that of sea-floor spreading, it has been remarkably successful in clarifying the origins of many tectonic features associated with continental margins, island arcs, and orogenic belts. Recent applications of plate tectonics theory to continental areas have led to important revisions in concepts of the evolution of major geological features of orogenic belts (Dewey and Bird, 1970; Bird and Dewey, 1970; Hamilton, 1969a, 1970; Dickinson, 1970, 1971; Atwater, 1970). Specifically, efforts to interpret the structure and geology of both the Appalachian orogen (Bird and Dewey, 1970) and the cor-

Geological Society of America Bulletin, v. 82, p. 3325-3346, 10 figs., December 1971

dilleran system of the western United States (Atwater, 1970; Hamilton, 1969a) have been notably successful.

The present Andean orogenic belt appears to be a modern active analogue of inactive cordilleran mountain belts elsewhere in the world (Hamilton, 1969b), and it is these inactive belts that are now interpreted in terms of continental margin plate tectonics. In this sense, then, the Andean orogenic belt provides direct observational evidence concerning processes assumed to have played a major role in the evolution of continental margins through geological time.

This paper examines available geophysical and geological data pertaining to the central Andes, with a view both toward interpreting Andean evolution within the framework of plate tectonics and toward outlining and evaluating those features which bear upon plate tectonics in orogenic evolution. Some aspects of older cordilleran systems will be examined in the light of Andean structures; conversely, older, eroded cordilleran terranes may provide insight into lower structural levels beneath the Andes.

Physiographic Setting

The area of this study includes southern Peru, Bolivia, and northern Chile (Fig. 1). The Andean mountain belt between about 8° S. lat consists essentially of a single chain of closely spaced mountain ranges trending parallel to the coast. South of about 12° S. lat, the mountains branch: the western belt continues parallel to the coast, while the eastern belt trends southeast and forms a series of mountain ranges several hundred kilometers from the coast. A broad flat plateau about 4 km in elevation separates the two mountain belts. The crests of both the eastern and western ranges are generally above 5 km, although in the southern part of the area, the eastern ranges are somewhat lower. The western mountain belt (Cordillera Occidental) is here referred to as the western cordillera. The eastern belt consists of several ranges, among which are Cordillera de Vilcabamba, Cordillera de Vilcanota, Cordillera Real, Cordillera Oriental, and Cordillera Central, here referred to as the eastern cordillera. The high plateau between these cordilleras, termed "Puna de Atacama" in the southern part of the area and the "altiplano" elsewhere, is here referred to as the altiplano.

The Peru-Chile trench runs parallel to and about 300 km west of the volcanic crest of the central Andes. The trench is continuous in the area shown in Figure 1. It is deepest (about 7.6 km) and closest to shore (about 70 km) off the coast of northern Chile. Farther north, the trench is less deep and locally as much as 200 km from the coast. The maximum topographic relief between the deepest part of the trench and highest Andean peaks is about 15 km.

Geological Setting

The eastern cordillera (Fig. 2) consists primarily of Paleozoic dark shales, sandstones, and carbonates which probably were laid down on the continental shelf and slope of that time. These deposits may aggregate 10 to 15 km in thickness (Ahlfeld and Branisa, 1960) and have been extensively folded and faulted eastward by Cenozoic compression from the west. Eastward compression has also produced strong folding and thrusting of Mesozoic and Cenozoic continental deposits in the subandean ranges at the eastern margin of the Andean orogen (Ahlfeld, 1970). Little volcanism and no large-scale plutonism have occurred in the eastern cordillera, although in the high mountains of northern Bolivia, the Paleozoic sedimentary rocks are intruded by granitic rocks of mid-Mesozoic and mid-Tertiary age; in addition, hypabyssal and volcanic rocks of Tertiary age occur on the western flanks of the mountains.

The western cordillera consists mostly of volcanic and plutonic rocks of Mesozoic and Cenozoic age and associated continental sedimentary rocks. Shallow-water marine deposits of Jurassic age are widespread and some marine deposition persists into Cretaceous time. The crest of the western cordillera is dominated by volcanic peaks formed during the culminating stages of the most recent volcanic episode. The lavas of this episode appear mostly to be less than 10 m.y. old (Rutland and Guest, 1965). Flanking the volcanic crest to the west is the granitic Andean batholith. Although it had been widely assumed that the batholith is Cretaceous or early Tertiary in age (for example, Jenks, 1956), recent K-Ar age dating indicates that the picture is considerably more complex (Giletti and Day, 1968; Farrar and others, 1970). The study by Farrar and others for a small area in northern Chile between about 27° S. and 28° S. lat demonstrates that the plutonic rocks range in age from Lower Jurassic to upper Eocene, with the older rocks to the west.

During latest Mesozoic and the Cenozoic, the altiplano has been a depositional basin for continental clastic sediments derived from both

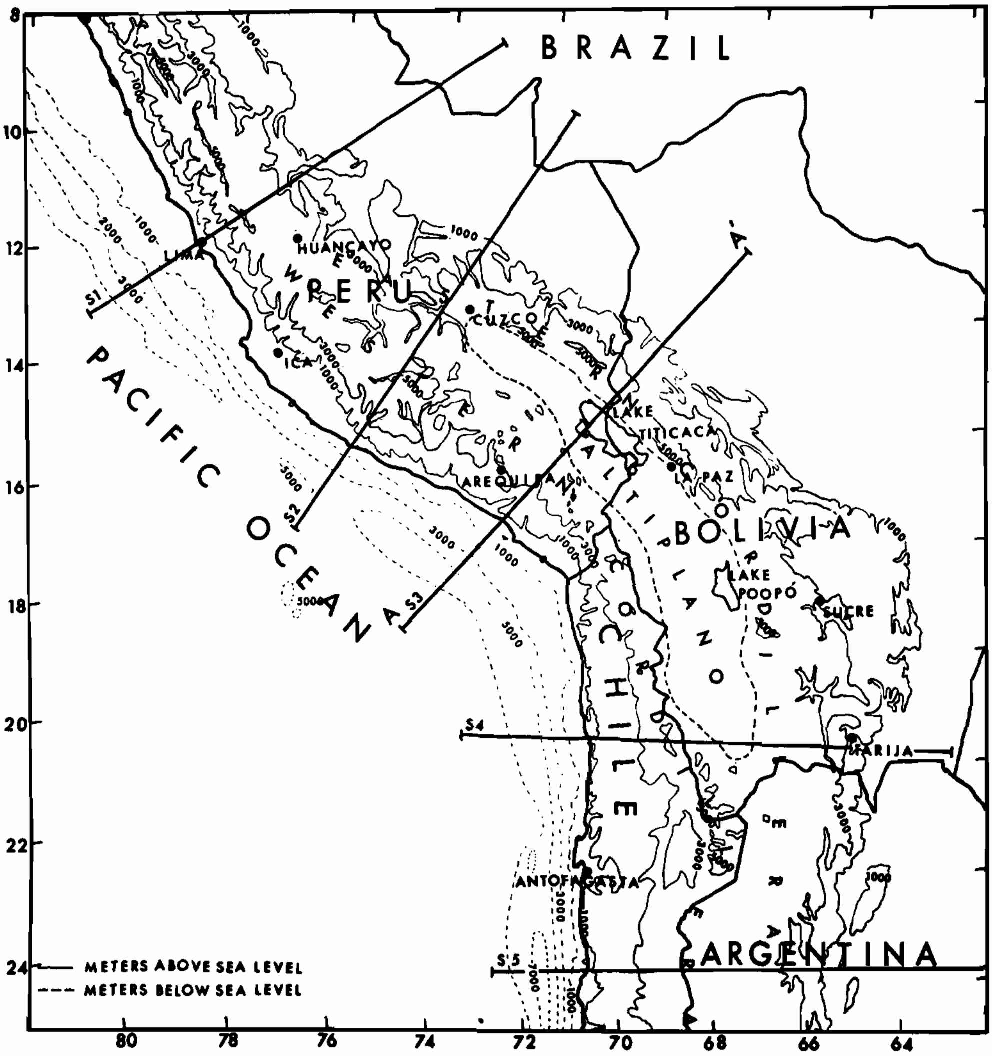

Figure 1. Map of central Andes showing the morphological provinces considered in this paper. Lines S1-S5 denote seismicity cross sections shown in Figures 5 and 6. A-A′ indicates line of structure cross section shown in Figure 7.

cordilleras. These sediments hide the pre-Cretaceous rocks whose history is unknown. The total volume of molasse (primarily continental) deposition in the altiplano is enormous. In the well-studied Lake Titicaca region, Newell (1949) reports that Cretaceous rocks may have an aggregate thickness of nearly 7 km. The Cenozoic sequence in Bolivian parts of the altiplano may aggregate as much as 15 km (Evernden and others, 1966). If a Paleozoic section comparable to that in the eastern cordillera lies beneath the Cretaceous rocks of the Lake Titicaca region, the total sedimentary section could be more than 30 km.

GEOPHYSICAL EVIDENCE

Most pertinent to an interpretation of the central Andes are the spatial distribution of earthquakes and the structure of the crust and upper mantle. Earthquake distribution is here taken from the U.S. Coast and Geodetic Survey hypocenter compilation for the period 1964 to the present. Many South American hypocenters are not well determined; nonetheless, when compared to more precise locations obtained by local networks (James and others, 1969), it is clear that the inaccuracy is not sufficient to spoil the over-all picture of regional seismicity.

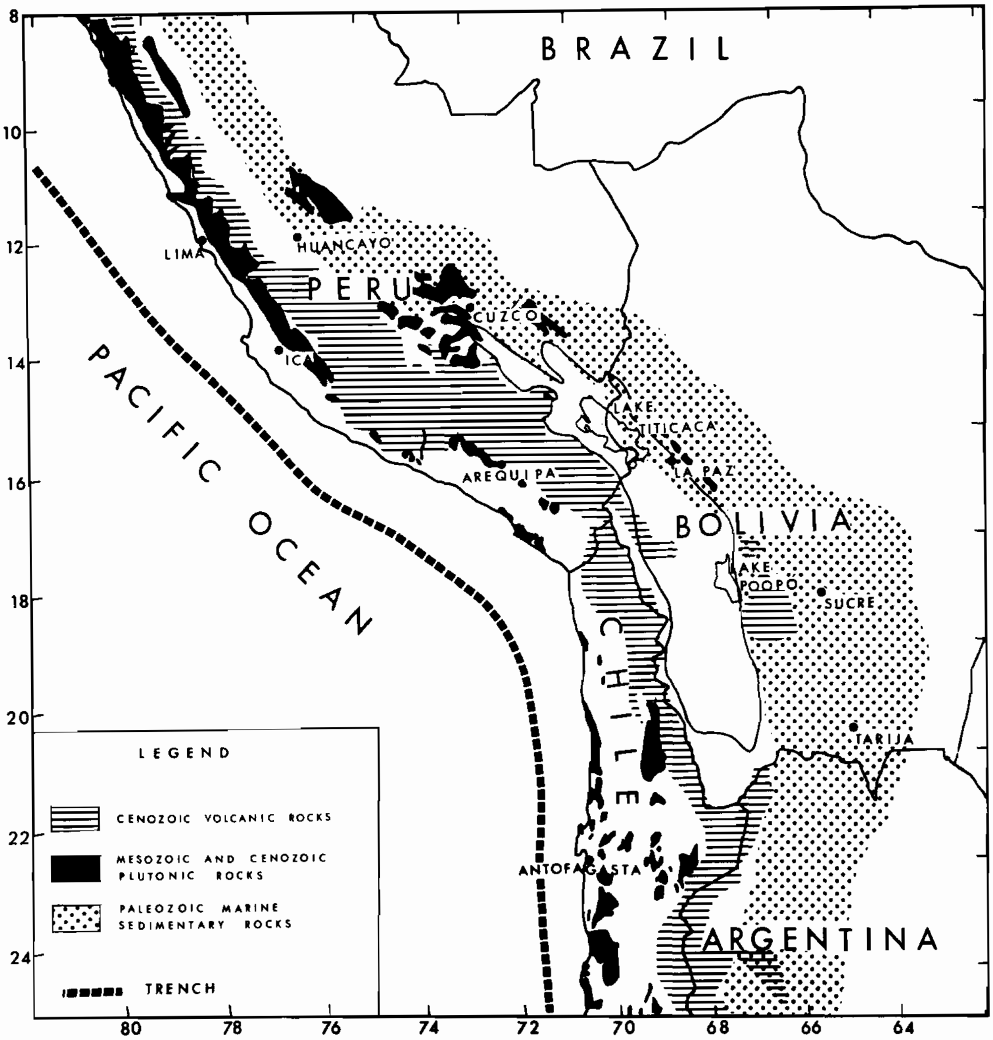

Figure 2. Outline geological map showing distribution of major rock types in the central Andes (*after* Geologic Map of South America, published by the Commission on the Geologic Map of the World, 1963).

Crustal and upper mantle velocity structure is based upon surface wave results presented previously (James, 1971) in which Rayleigh and Love wave phase and group velocities were determined for paths within the central Andes. Analysis of phase and group velocities yields shear velocity, compressional velocity, and density as functions of depth. Phase and group velocities of surface waves are controlled primarily by shear velocities, and both compressional velocity and density are commonly constrained to be simple functions of shear velocity. The layered structures derived from surface wave analysis are idealized and nonunique earth models—layered models are convenient for analysis, but there is no intent to suggest that the crust and upper mantle are in fact layered. The layers reflect on a gross scale what are probably gradational changes in elastic properties with depth.

Plate Geometry

Both upper mantle structure and seismicity provide evidence as to the geometry of the South American and Nazca plates. The top of the low-velocity channel for shear waves is here assumed to correspond to the base of the lithospheric plate; thus, the low-velocity zone is presumed to be a region of less mechanical strength and greater S-wave attenuation than the overlying lithosphere. Upper mantle velocities and S-wave attenuation thus provide evidence as to the thicknesses of the two converging plates. The Benioff zone is here taken to correspond approximately to the upper part of the underthrust plate. Figure 3 shows schematically the inferred structure of the plate junction.

Upper Mantle Structure. Surface-wave results argue against a low-velocity channel beneath the central Andes to a depth of at least 200 km in the upper mantle. Shear velocities in the uppermost mantle are about 4.4 to 4.5 km/sec and do not decrease with depth. Additional evidence for the absence of a low-velocity channel above 200 km is suggested by the comparatively slight attenuation of S-waves traveling between intermediate depth earthquakes (h ~ 200 km) and stations situated in the cordilleras. Figure 4 shows P-wave and S-wave arrivals from intermediate depth earthquakes recorded at stations situated above the Benioff zone and back of the volcanic arc in Tonga and South America. The comparison with the Tonga region is appropriate, because a major low-velocity zone exists there in the mantle above 200-km depth (Oliver and Isacks, 1967). Despite differences in instrumental frequency response of the two examples shown in Figure 4 (the Fiji seismograph system peaks at a higher frequency), it is clear that the shear waves recorded at Fiji have suffered substantially greater attenuation than those recorded at La Paz, Bolivia. The La Paz records generally exhibit stronger shear wave attenuation for intermediate earthquakes than records of other permanent stations in the central Andes and thus measure the approximate upper limit of attenuation in the mantle above the Benioff zone. These observations and conclusions are supported by the more thorough studies of Sacks (1971).

The absence of a clearly defined low-velocity or high-attenuation channel in the upper mantle beneath the cordillera prompts looking to greater depths. Sacks (1963, 1971) showed that the earthquake-free zone beneath South America in the depth range 300 to 500 km is also a zone of low Q (high attenuation) and probable low shear-wave velocity. This low-velocity and high-attenuation zone places a maximum depth constraint on plate thickness; accordingly, the lithosphere beneath the central Andes is more than 200 but less than 300 km thick.

The structure of the upper mantle beneath the Pacific basin adjacent to southern Peru and northern Chile is distinctly different from that beneath the cordilleras. Here, a major low-velocity channel for shear waves lies at a depth of only 50 to 60 km (James, 1971) indicating a Nazca plate thickness of 50 to 60 km at the juncture with the South American plate.

Seismicity. Patterns of global seismicity have assumed particular significance in recent years when examined in the light of plate tectonics. Seismic activity in island arcs appears to be confined largely to rather narrow Benioff

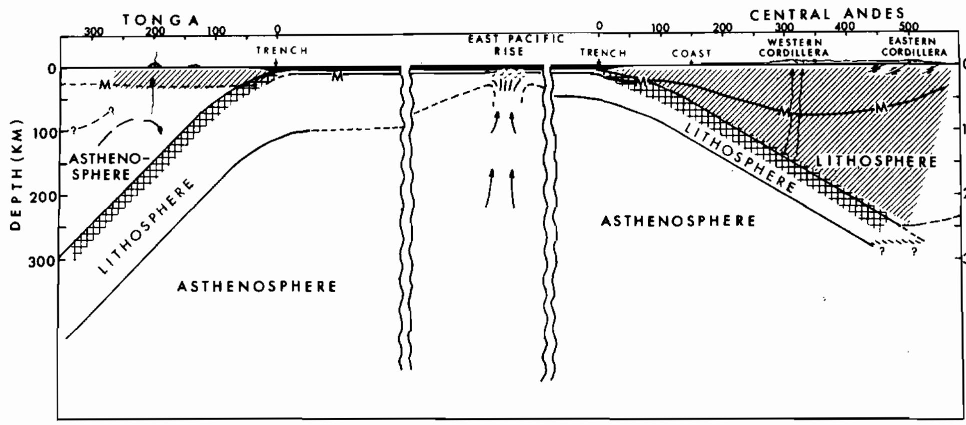

Figure 3. Schematic cross section, showing juncture between South American and Pacific plates. Plate configuration for Tonga arc is shown for comparison and is adapted from Oliver and others (1968). Diagonal hatching indicates region of moderate seismicity; square hatching indicates regions of high seismicity. Letter "M" denotes M-discontinuity.

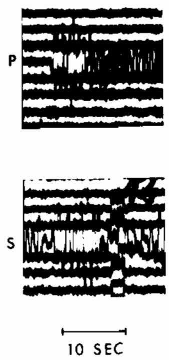

Figure 4A. Recording at World Wide Standard Station LPD (La Paz, Bolivia) for earthquake at 246-km depth and 6.1° epicentral distance. Note that the S-waves, while attenuated, do not exhibit the marked attenuation of the S-waves in Figure 4B.

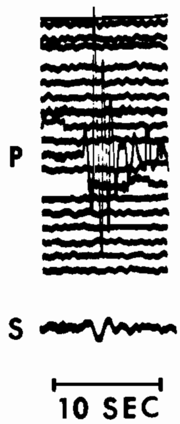

Figure 4B. Recording at Fiji station Vunikawa for earthquake at 208-km depth and 5.7° epicentral distance (*after* Oliver and Isacks, 1967). Note the almost complete attenuation of the S-wave arrival relative to the P-waves.

zones (for example, Sykes, 1966), and this observation has played a major role in developing some of the fundamental tenets of plate tectonics. The seismicity of the central Andes is of somewhat different style than that of island arcs and sheds light on some important differences between cordilleran and island-arc tectonics.

Earthquake hypocenters in the central Andes were projected onto a series of vertical sections oriented perpendicular to the arc structure (Fig. 5). The locations of the five sections are shown in Figure 1 and each projection contains only those events within 90 km of the section. The discussion that follows is limited to activity above 300 km.

The depth distribution of earthquakes shown in Figure 5 provides for several important observations. First, there is a total cessation of seismic activity at a depth of no more than about 300 km; in some regions, activity terminates at depths of little more than 100 km. The greater depth limits (200 to 300 km) at which activity ceases correspond with what may be the top of a low-velocity zone. An interesting feature of the earthquake distribution is that although seismic activity is most intense in a rather narrow belt (the Benioff zone) some activity occurs within the entire region between the Benioff zone and the earth's surface. This observation is especially important, as it contrasts sharply with those in the Tonga-Fiji region (Sykes, 1966; Sykes and others, 1969; Mitronovas and others, 1969) and other island arcs where earthquake activity is confined to comparatively narrow Benioff zones. Thus seismic activity occurs within the rigid South American lithospheric plate as well as in the upper parts of the underthrusting Pacific plate. If the South American plate is between 200 and 300 km thick, it is plausible that stress interaction between the South American and Pacific plates would produce earthquake strain release within the brittle South American plate.

Comparison with the Tonga-Fiji Island Arc. Figure 3 compares the South American structure as it has been interpreted above with the structure of the Tonga-Fiji island arc as interpreted by a group of workers at the Lamont-Doherty Geological Observatory of Columbia University (Sykes, 1966; Oliver and Isacks, 1967; Isacks and others, 1968). Behind (west of) the Tonga arc, there may actually be a gap in which there is no lithospheric plate, whereas behind (east of) the Peru-Chile trench, the South American plate is continuous and of considerable thickness. This important contrast may explain some principal differences. In the Tonga-Fiji region, the downgoing lithospheric plate is bounded both top and bottom by comparatively mobile asthenosphere. Thus, at depths greater than about 100 km, virtually all of the earthquakes occur within the downgoing lithospheric plate (Sykes, 1966). Moreover, overturn may occur in the asthenosphere above the descending slab, producing secondary spreading behind the arc (Karig, 1970, 1971). In contrast, earthquakes are not confined to the Benioff zone beneath South America nor is there any evidence of spreading behind the arc. These observations suggest that the thick lithospheric plate beneath South American precludes secondary spreading (or convection) and that, therefore, the stresses induced by the downgoing slab produce seismic activity rather than convective overturn above the Benioff zone.

Crustal Structure

The pertinent structure of the crust beneath the central Andes is shown in Figures 6 and 7. Figure 6 is a somewhat speculative contour map of the M-discontinuity beneath the land area, and Figure 7 is a vertical cross section. The

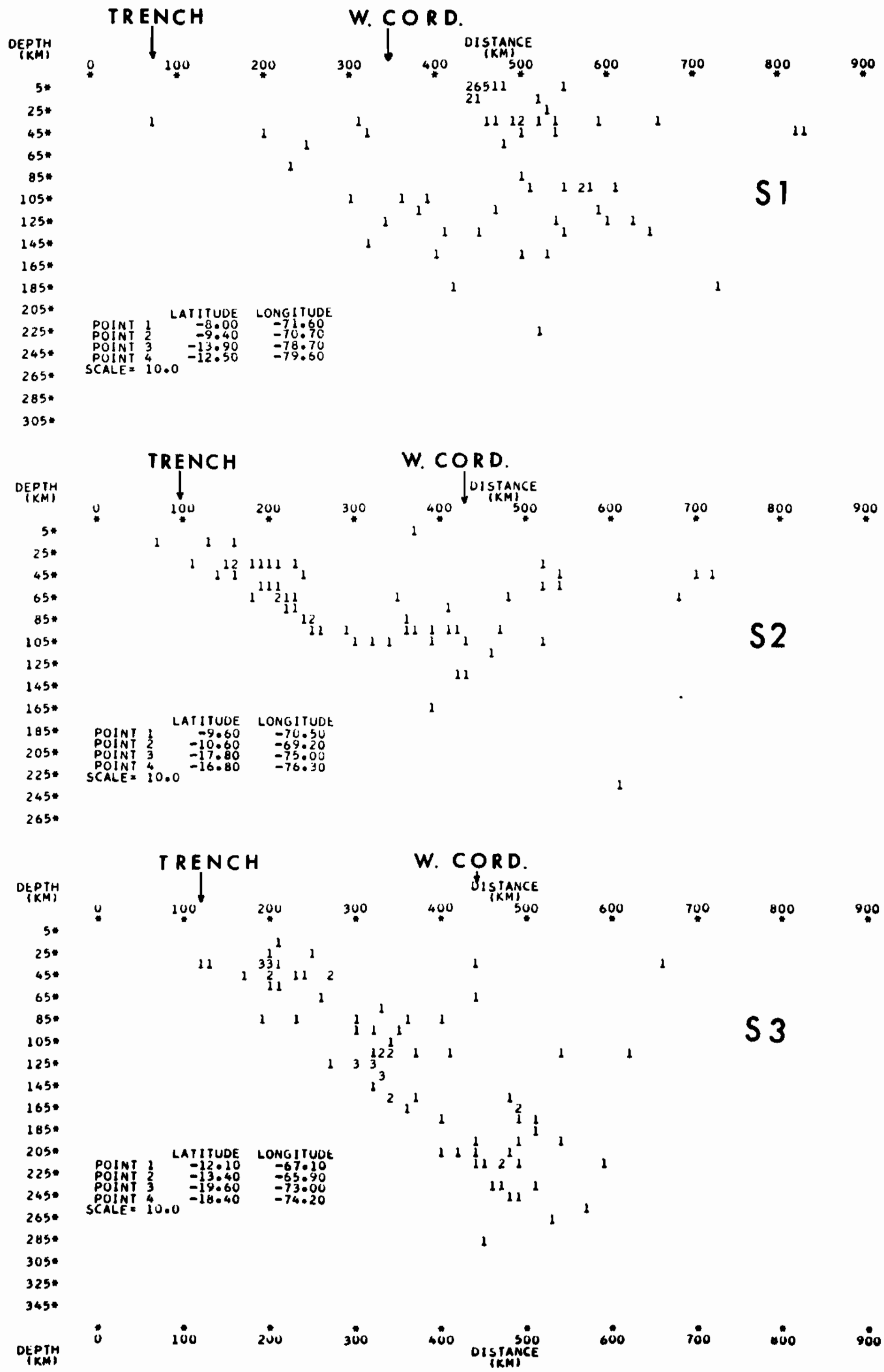

Figure 5A. Seismicity cross sections for lines S1 through S3 shown on Figure 1. Hypocenters projected from distances within $\pm$ 90 km of each cross section. The numbers denote number of hypocenters within each 10-km square block. Data from U.S. Coast and Geodetic Survey compilation for period January 1964 to June 1970.

following summary is based on results given elsewhere (James, 1971).

The most striking aspect of the Andean structure is crustal thickness. Over a distance of little more than 500 km, it varies from about 11 km (including water layer) in the Pacific basin to 30 km along the coast to more than 70 km beneath the western cordillera and western part of the altiplano. The crust thins rapidly eastward, and beneath the eastern cordillera is only about 50 to 55 km thick. The crustal thickness of more than 70 km beneath the western cordillera and western altiplano is among the greatest yet measured anywhere in the world.

The crustal velocities (Fig. 7) are derived from surface-wave analysis, but they are consistent with available refraction results (Woollard, 1960; Ocola and others, 1971). The crust beneath the cordilleras can be approximated by a three-layer model; the crust beneath the coastal areas by a two-layer one. The petrologic implications of these models bear critically

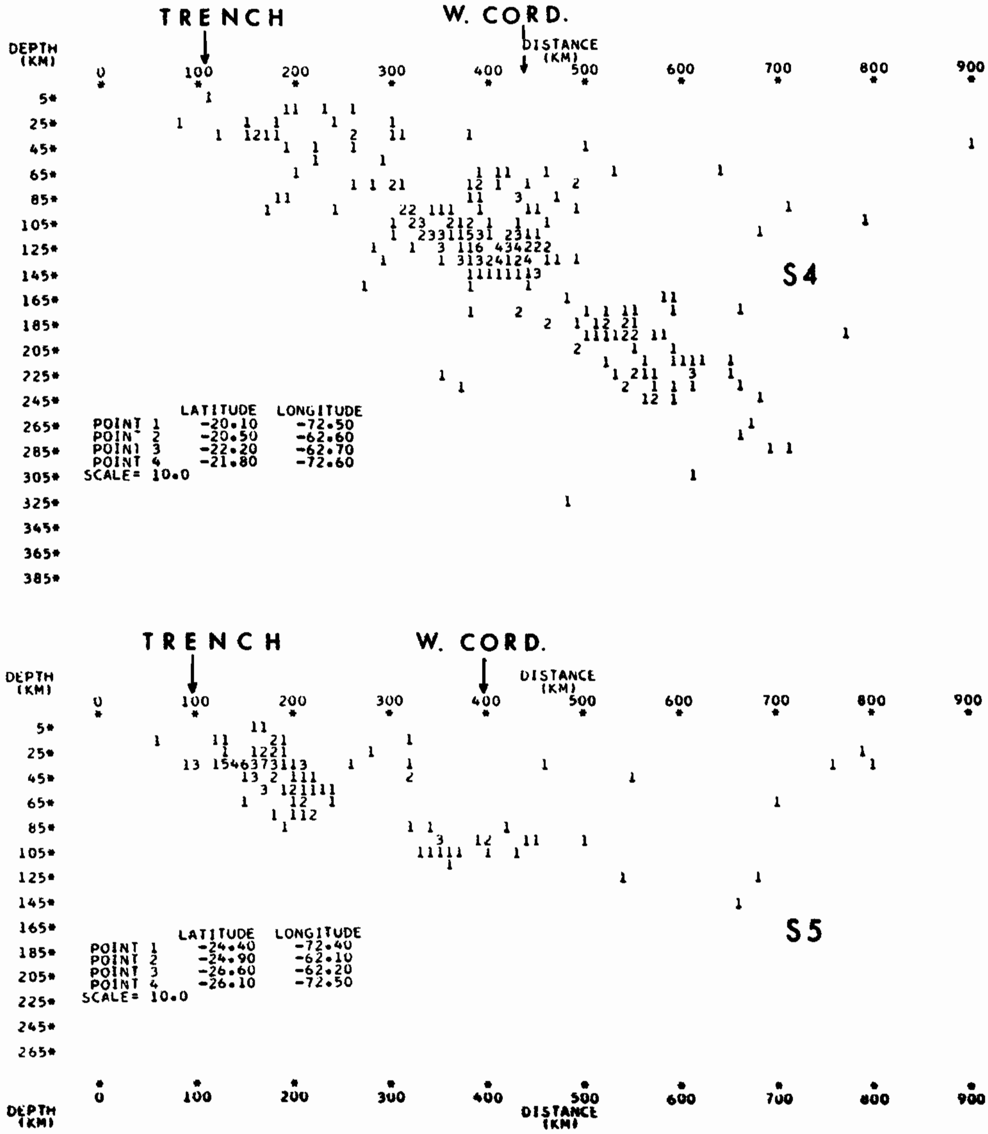

Figure 5B. Vertical sections for lines S4 and S5. Explanation same as for Figure 5A.

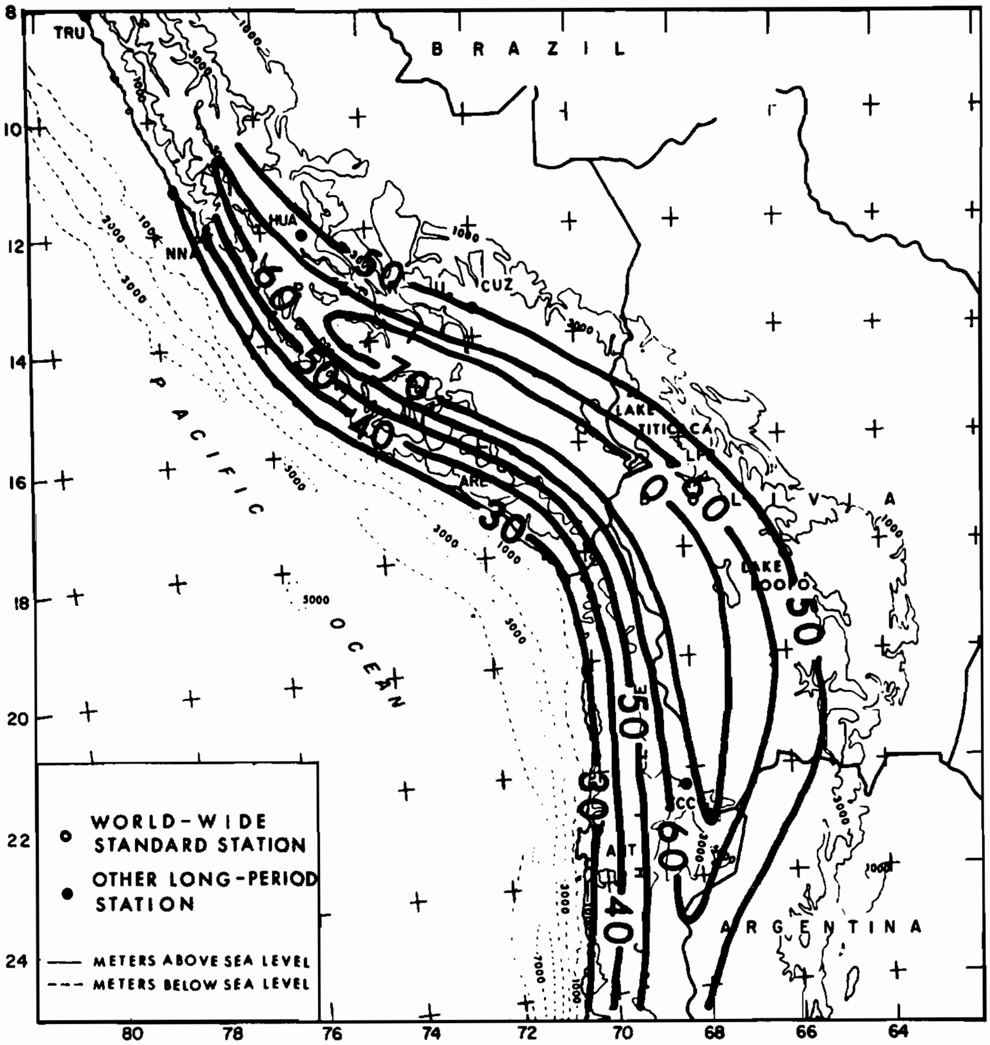

Figure 6. Contour map of depth (in kilometers) to the M-discontinuity beneath the central Andes (*from* James, 1971).

upon the evolution of the Andean arc, and it is appropriate to consider the crustal velocities in the context of the information and constraints they provide upon rock types. Discussion below is confined to P-wave velocities, as they are well known and more easily related to rock types than shear-wave velocities. Only shear-wave velocities, however, have been obtained explicitly.

The rocks in the lower parts of the crust, with P-wave velocities consistently near 6.6 km/sec, could be of gabbroic composition beneath the ocean basin, but may well be of intermediate composition beneath the land area. It is a common misconception that gabbroic compositions must be characteristic of the lower crust because they exhibit the appropriate laboratory velocities and densities. Ringwood and Green (1966, p. 612) challenge this notion and assert that on the basis of chemical equilibrium studies, "the inference that the lower crust is composed of gabbroic or basaltic rocks is almost certainly wrong...." Their experimental studies demonstrate that mafic rocks, when subjected to pressures and temperatures in the range suggested for the lower parts of the crust, are metamorphosed to the eclogite facies or garnet-clinopyroxene-granulite subfacies if the lower crust is anhydrous. Since these rocks should have substantially higher velocities and densities than have been inferred for the lower crust, they conclude that an anhydrous lower crust cannot be composed of mafic rocks but instead must consist of silicic to intermediate high-grade metamorphic rocks. On the other hand, if the lower crust is hydrous and low temperature (T < 700° C), the rocks could be amphibolites with densities and seismic velocities consistent with those observed. The likelihood of a hot lower crust and the absence of large volumes of exposed rocks of mafic composition give little encouragement to the notion of a mafic lower crust, and it seems plausible that the lower crust is both anhydrous and compositionally similar to the intermediate and silicic volcanic and plutonic rocks which form the upper levels of the crust.

Rocks at intermediate to shallow depth with P-wave velocity around 6.0 km/sec may be either granitic rocks or metamorphosed sedimentary rocks. Beneath the western cordillera and its western flanks, the intermediate-depth crustal rocks may be comagmatic with the granitic or intermediate rocks exposed at the surface; beneath the altiplano and the eastern cordillera, the 6.0 km/sec rocks could be slightly metamorphosed sedimentary sequences, although igneous rocks may be mixed with these metamorphic assemblages.

Within the uppermost crust beneath the altiplano and the western part of the eastern cordillera, there is approximately 10 km of rock with a velocity of about 5.0 km/sec. In the altiplano, the presence of this layer is confirmed by explosion refraction studies (Ocola and others, 1971). The only rocks with measured velocities consistently this low at depths greater than a few kilometers are sedimentary and volcanic (Press, 1966). This suggests, then, that at least the upper 10 km of the crust in this region consists of sedimentary rocks with possible associated volcanic sequences. There is strong evidence for a layer about 5 km thick with P-wave velocity near 5.0 to 5.2 km/sec in the

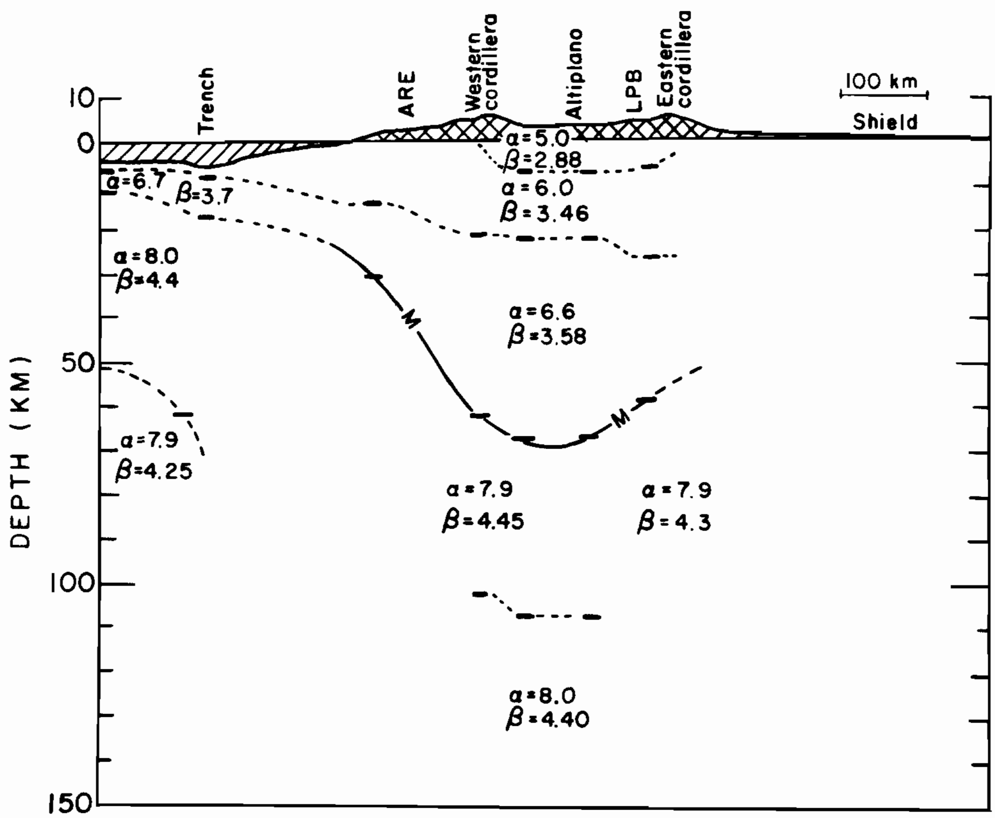

Figure 7. Cross section of crustal and upper mantle structure along line A-A′ of Figure 1. Crustal structure beneath trench from Fisher and Raitt (1962) and remaining structure from James (1971). α denotes P-wave velocity; β denotes S-wave velocity.

uppermost crust beneath the western cordillera. This layer could consist of either volcanic or sedimentary rocks; geological evidence suggests that it may be a combination of both. No seismic velocities less than 6.0 km/sec have been observed in the coastal areas.

In the zone situated between the trench and the shoreline, it is readily apparent from studies by Fisher and Raitt (1962) and Scholl and others (1970) that this region is not a site of "eugeosynclinal" deposition of terrigenous sediments. Indeed, basement rocks with velocities of 5 to 6 km/sec are nearly everywhere found at depths of less than 1 km beneath a thin sedimentary cover. At the bottom of the trench off Peru, sediment accumulations appear to be less than 1 km (Fisher and Raitt, 1962) and off northern Chile may not be present at all (Scholl and others, 1970, profile D-8, and Figs. 6 and 7). The seismic velocities in the basement rocks underlying the thin sedimentary cover are appropriate for granitic or metamorphic rocks.

GEOLOGY OF THE CENTRAL ANDES

It is probably always somewhat hazardous to attempt to integrate a large amount of geological information, the work of many people with varying biases, into a coherent pattern suitable for analysis in terms of broad and unquestionably over-simplified concepts of tectonic evolution. It should be stated at the outset that no attempt has been made to be complete, and emphasis is given only to those aspects of the geology that seem to bear most critically on the orogenic evolution of the Andes. Moreover, much of the geology of this part of South America has never been investigated; nearly every aspect of the geology, from the details of the Paleozoic stratigraphy to the chemistry of the recent volcanic rocks, is sadly riddled with unknowns. In spite of these limitations (although ignorance seldom hinders imagination), it is my intent here to place the major geological features known into a workable framework of plate tectonics. The simplicity of the discussion that follows is not meant to deny the complexity or even contradictions more readily apparent on finer scale.

A geologic map is shown in Figure 8. The stratigraphic and tectonic development of the central Andes is described below in more or less chronological outline. The following section is devoted to a consideration of the igneous activity associated with the orogenic belt and to the implications of magma source and genesis.

The oldest rocks in the central Andes are of Paleozoic age, mostly confined to a broad belt in the eastern cordillera. The only known pre-Mesozoic rocks in the Peruvian western cordillera are some scattered orthogneisses and paragneisses of dioritic to granitic composition in a belt along the southern coast (Jenks, 1956). These rocks were metamorphosed more than 400 m.y. ago (Aldrich and others, 1958). Other probable Precambrian or Paleozoic metamorphic rocks have been reported in northern Chile, but the ages are not well established.

The Paleozoic sequences of the eastern cordillera of the central Andes have been best studied in Bolivia (Ahlfeld and Branisa, 1960; Ahlfeld, 1956) and in the Lake Titicaca region of southern Peru (Newell, 1949). Only meager information exists on the Paleozoic rocks of the Cordillera Oriental of southern Peru (Steinmann, 1930; Newell and others, 1953; Hosmer, 1959). The following descriptions of Paleozoic stratigraphy pertain mostly to Bolivia and unless otherwise referenced are based upon the work of Ahlfeld (1956, 1970) or Ahlfeld and Branisa (1960).

The rocks of the eastern cordillera of Bolivia and southern Peru consist predominantly of Ordovician, Silurian, and Devonian deposits belonging to a single clastic facies of dark-gray shales and sandstones, and their low-grade metamorphic equivalents (Newell, 1949). These "miogeosynclinal" rocks are here interpreted as continental shelf deposits. Widespread deposition began in the Ordovician and continued into the Early Permian, and the aggregate thickness of Paleozoic deposits is thought to exceed 10 km. The only tectonic activity during this period appears to have been gentle epeirogenic movements.

Rocks of Ordovician age may aggregate 3 to 5 km and consist largely of sandy shales, quartzitic sandstones, and occasional black shales (Ahlfeld, 1956). Silurian strata are poorly developed and consist mostly of nonfossiliferous sandstones. Devonian deposits 5 km or more in thickness lie conformably on the older Paleozoic sequences. The Devonian in southern Peru and Bolivia is an extremely uniform succession of dark shales, sandy shales, and quartzitic sandstones. Limestones are almost lacking, as in the Ordovician and Silurian strata. In the eastern Andes, the Devonian is rich in plant remains and bituminous shales, suggesting near-shore environment. Carboniferous

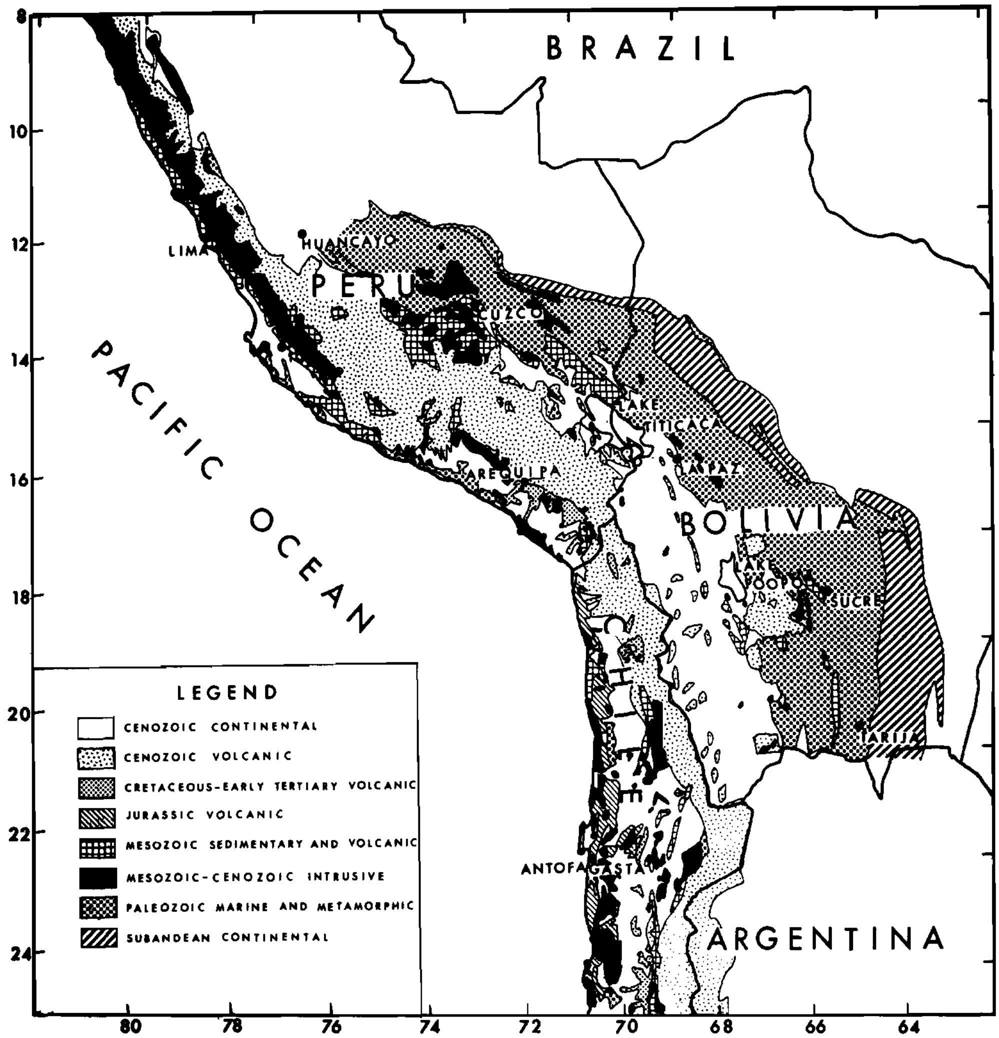

Figure 8. Geologic map showing distribution of principal rock units of the central Andes (*adapted from* Geologic Map of South America; Geologic Map of Bolivia, Bolivian Geol. Survey, 1968; and *from* Geologic Map of Peru, Peruvian Geol. Soc., 1956).

marine deposits are rare or absent; in Early Permian, the last of the shelf deposits were laid down in Peru and Bolivia. Permian rocks are predominantly limestones, carbonaceous shales, and arkosic and continental sandstones, representing a littoral facies of alternating marine and continental deposits (Newell, 1949).

No rocks of mid-Permian to mid-Cretaceous age are reported in the eastern cordillera. Minor volcanism and hypabyssal intrusion occurred both in Peru and Bolivia during this hiatus in deposition. Some small granitic batholiths in the Cordillera Real give Early Jurassic or Late Triassic K/Ar ages (Evernden and Kistler, 1970). Possibly the low-grade metamorphic phase that affected the Paleozoic strata of the Cordillera Real was contemporaneous with this plutonism.

Formation of the western cordilleran volcanic arc began in Late Triassic or Early Jurassic time near the present coast of northern Chile and southern Peru. The rocks, consisting largely of pillow lavas, tuffs, breccias, and agglomerates of basaltic to andesitic composition (Hosmer, 1959), probably represent submarine stages of development of an island arc which formed immediately west of the Paleozoic continental shelf. These rocks occur only in the coastal areas as shown in Figure 8. Along the southern Peruvian coast, the volcanic rocks may attain a thickness of 4 km (Hosmer, 1959). Middle and Upper Jurassic marine limestones, shales, and sandstones occur abundantly in a zone within and west of the present cordilleran crest. These deposits indicate that a shallow sea existed between the volcanic arc and the eastern ranges. Volcanism continued in Middle and Upper Jurassic times along the coast of central Peru near Lima (Hosmer, 1959) and may have extended farther south at times.

The oldest plutonic rocks of the Andean batholith are at least as old as Lower Jurassic (Farrar and others, 1970) and may be contemporaneous with Triassic/Jurassic volcanism. The Lower Jurassic granitic rocks occur along the coastline of northern Chile (lat ~ 28° S.) less than 100 km from the trench. The plutonic rocks in this area exhibit progressively younger ages eastward, and the small batholiths just west of the cordilleran crest are of Eocene age. The age of the Andean batholith of central and southern Peru is Late Cretaceous or Early Tertiary (Giletti and Day, 1968), and here, too, the older rocks are to the west. The measurements cited are all K-Ar mica ages and are, therefore, measurements only of the final stage of cooling (or most recent metamorphism) of the rock and not necessarily of the times of emplacement.

Volcanic rocks, predominantly andesites with some dacites and rhyolites, of Late Jurassic or Early Cretaceous age are found extensively along coastal central Peru (Hosmer, 1959). Interlayered with them are continental and shallow water marine deposits. In northern Chile, volcanic and intercalated sedimentary rocks of Jurassic and Cretaceous age are as much as 15 km thick (Zeil, 1964), but the rock types are poorly known; in general, they appear to be largely keratophyres and andesites (Zeil, 1964), probably laid down in submarine conditions similar to those inferred for the equivalent rocks in southern Peru. Generally, volcanic rocks of Cretaceous age are comparatively rare (Hosmer, 1959), although Late Cretaceous or early Tertiary rocks are especially widespread in Peru south of Arequipa and in northern Chile.

Except for a few thin marine limestone beds of mid-Cretaceous age, Cretaceous and younger deposits within the altiplano and flanking areas of the eastern cordillera are continental or volcanic in origin. The continental deposits are of molasse type and derived from both western and eastern cordilleras. The Cretaceous rocks are predominantly red beds with abundant conglomerates in both middle and Upper Cretaceous sections. The thickness of Cretaceous deposits varies greatly from place to place, but northeast of Lake Titicaca, the section may be 4 km thick (Newell, 1949); southeast of Lake Poopó, along the eastern edge of the altiplano, the Cretaceous section may total more than 7 km (Kriz and others, 1965).

The Cenozoic rocks of the central Andes are almost entirely volcanic and continental. The few marine deposits are confined to coastal areas. Cenozoic volcanic rocks, mostly Miocene or younger, have been most extensively studied in the western cordillera of northern Chile. Widespread Tertiary volcanism began here with pyroclastic eruptions. The areal extent of the pyroclastics is variously estimated between 110,000 km^2 (Petersen, 1958) and 150,000 km^2 (Zeil and Pichler, 1967) with an average thickness of nearly .5 km. Numerous K/Ar age dates are published and most range from at least 10 m.y. (upper Miocene) to 4.25 m.y. (late Pliocene) (Rutland and Guest, 1965; Clark and others, 1967). The pyroclastics range from andesite to rhyolite and are predominantly dacites or rhyodacites (Guest, 1969). They were apparently erupted from fissures which occupied the approximate crest of the modern Andes, and it has been proposed that the period of recent uplift in the Andes began with the volcanic eruptions (Rutland and Guest, 1965; Hollingworth and Rutland, 1968). Quaternary eruptions in northern Chile produced the "andesitic" strato-volcanoes that now dominate the crest of the Andes (Pichler and Zeil, 1969). The predominant rock type of these flows is quartz-bearing latite-andesite.

The major Tertiary volcanism in southern Peru is contemporaneous with that in northern Chile. Activity began in late Miocene time, increased in Pliocene/Pleistocene time, and continued with waning intensity to historical time. The extrusives have been best described for the region around Arequipa, Peru (Jenks, 1948; Jenks and Goldich, 1956), where andesite is by far the predominant rock type. Rhyolitic ignimbrites constitute less than 1 percent of the total volume of the volcanic rocks.

Cenozoic volcanism occurs in the altiplano and as far east as the western flank of the eastern cordillera. The volcanic rocks are variously and vaguely described as dacitic or andesitic flows and tuffs. The volcanic rocks are intercalated in thick sections of continental debris. Molasse deposition that began in the altiplano in Cretaceous time continued throughout Tertiary and Quaternary times. Near Lake Titicaca, a lower Tertiary section about 7 km thick is followed by 4 km of younger Tertiary and Quaternary rocks, a large part of which is volcanic (Newell, 1949). The altiplano in Bolivia may be underlain by even thicker Tertiary deposits. Evernden and others (1966) estimate that Tertiary continental deposits, lava flows, and pyroclastic debris may exceed 15 km in the Bolivian altiplano; in the west-central Bolivian altiplano, Meyer and Murillo (1961) gave an average thickness of 10 km for an incomplete Tertiary section.

Miocene compression produced east-directed folding and faulting for the first time in the continental section of the altiplano and eastern cordillera. This compression affected the entire eastern ranges of the central Andes (Jenks, 1956) and was followed by much more intense compression in Pliocene time (Ahlfeld and Branisa, 1960), possibly coinciding with the Pliocene-Pleistocene volcanism. Continentward thrust faulting occurs throughout the eastern cordillera (Ahlfeld and Branisa, 1960). Newell (1949) believes that much of the uplift of the eastern cordillera near Lake Titicaca has been accomplished by thrusting.

DERIVATION OF THE IGNEOUS ROCKS

Theories of the genesis of arc-related magmas have undergone considerable change in the light plate tectonics. Recently it has been proposed that andesitic melts derive ultimately either from the descending lithospheric plates themselves, or, by release of water from these plates, by partial melting of the mantle above the plate (*see,* for example, Dickinson, 1970). These propositions are now hotly debated, but evidence is mounting that the melts are derived from near the Benioff zone (Dickinson, 1970). It is additionally presumed that the large composite batholiths of orogenic belts are emplaced in the upper crust beneath a cover of their own volcanic ejecta (Hamilton and Myers, 1967; Hamilton, 1969b; Dickinson, 1970). The emplacement of plutonic rocks and the extrusion of volcanic rocks are assumed to be roughly contemporaneous events and the parent magmas cogenetic. By this hypothesis, the two kinds of igneous associations are no more than expressions of different structural levels in orogenic belts. An essential tenet of these theories is that most of the plutonic and volcanic rocks of orogenic belts derive ultimately from a zone near the top of the descending plate. The volcano-plutonic complexes of the Andes exhibit many examples of the close relations in time and space between volcanic rocks and small plutons, and they support in a general way the theory of consanguineous intrusive and extrusive rocks. Without pursuing the argument further, I shall assume as a basis for discussion that the plutonic rocks and volcanic rocks are derived from magmas initially generated near the Benioff zone. Within this framework of assumptions, then, there are a number of observations which seem especially to bear upon plate tectonics. This discussion and those that follow are not meant to deny the possibility that crustal differentiation and anatexis play a significant role in determining the nature and distribution of crustal igneous rocks.

Eastward Migration of Igneous Activity

Igneous activity in the central Andes has migrated progressively eastward (or northeastward) with time, in a direction normal to the trench. In a very general way, this migration is apparent in that the Andean batholith lies to the west of the present volcanic chain. But eastward younging has been observed even in detail within the large composite batholiths (Hosmer, 1959).

K-Ar ages yield a more concrete measure of the rate of eastward migration (if we assume that these ages correspond roughly to periods of emplacement and crystallization). The data of Farrar and others (1970) for an area in northern Chile indicate that the oldest rocks are Lower Jurassic in age and crop out very near the coast. Since Lower Jurassic time, igneous activity has migrated eastward at a rate of about 1 km/m.y. In central and southern Peru, results presented by Giletti and Day (1968) yield a similar rate of migration. The oldest rocks in Peru are Late Cretaceous or early Tertiary, yet these rocks and rocks of similar age in northern Chile are approximately equidistant from the trench, suggesting that the plutonic equivalents of the Lower Jurassic volcanic rocks of southern Peru lie beneath the continental shelf. Here, uppermost crustal velocities are appropriate for granitic rocks.

Migration of activity away from the trench has been observed in other island arcs (Kawano and Ueda, 1967), and it is perhaps appropriate to speculate as to factors that might produce the migration:

1. Progressive depression of isotherms near the upper margin of the descending plate would cause the zone in which melting or water release occurs to migrate down the subduction zone. This isotherm depression might be produced by continued underthrusting of a comparatively cold plate into lower and hotter parts of the mantle (*see*, for example, Oxburgh and Turcotte, 1970).

2. If melt is derived from the mantle above the downgoing plate, the migration may represent the time constant necessary to form sufficiently large batches of magma to enable upward movement; in the zone where large quantities of water are released from the plate, partial melting of the mantle may occur more rapidly than at greater depths along the subduction zone where less water is released and magma generation proceeds more slowly. Presumably, activity ceases when the adjacent mantle has been bled of its partial melt fraction.

3. The position of the subduction zone may change with time. In this event, the trench would have to move landward, implying stoping and underthrusting of the continental margin. The close approach (~ 70 km) of the Peru-Chile trench to the oldest plutonic rocks in northern Chile does suggest possible eastward migration of the trench.

4. The dip of the subduction zone changes, decreasing with time.

Present data do not permit a clear choice between these various speculations. Some potash relations presented in the following section seem to indicate that the dip of the Benioff zone may not have changed much during Cenozoic time, but the data are not conclusive. Of the speculations given, perhaps the most plausible mechanism for eastward migration is progressive depression of the isotherms.

K-h Relations

In a series of papers, Dickinson and Hatherton have shown that the potash content of island-arc andesites bears a simple functional relation to the depth to the Benioff zone beneath the area in which they were erupted (Dickinson and Hatherton, 1967; Dickinson, 1968, 1970). This correlation, termed K-h by Dickinson, suggests a genetic link between the inclined seismic zone and magma generation. Dickinson (1970) states further that the K-h correlation exhibits little or no dependence on the thickness or composition of the crust through which the magma penetrated, lending further support to the assertion that the magma is neither derived from nor significantly contaminated by the crust.

The crust beneath the Andean volcanic chain is in excess of 70 km, and it is reasonable to suppose that if crustal origin or contamination of andesitic magma is important, the Andean rocks will exhibit quite a different K-h relation than that found for island arcs. Some 60 analyses of modern volcanic rocks in northern Chile were used to obtain average potash content as a function of SiO_2 content (Pichler and Zeil, 1969; Zeil and Pichler, 1967; Guest, 1969; Jenks and Goldich, 1956). The depth to the Benioff zone is taken beneath the eruptive center and not the sites of sample collection. The depth to the Benioff zone beneath the eruptive centers is about 175 km. Figure 9 shows the northern Chile K-h values for three SiO_2 contents. The lines on this figure represent approximate best fits for the K-h diagrams

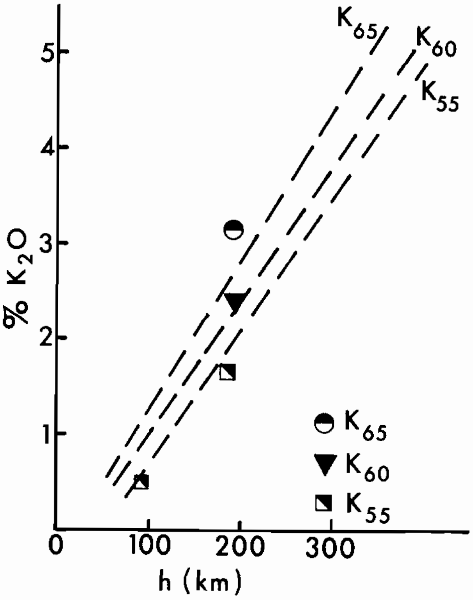

Figure 9. Averaged results of 60 analyses of northern Chile volcanic rocks plotted on a K-h diagram (Dickinson, 1970, p. 832). The percentage of K_2O for a given percentage SiO_2 is obtained from a variation diagram for the volcanic suite. Dashed lines labeled K_{55}, K_{60}, and K_{65} are approximate fits to Dickinson's (1970, Fig. 3) combined K-h plots for 55, 60, and 65 percent of SiO_2. Depth to Benioff zone, h, is taken beneath probable source of eruption, not beneath sample locality. The single additional value given for K_{55} (h = 80 km) is taken from two analyses of the Late Cretaceous/early Tertiary Acari pluton (*see* text).

presented by Dickinson (1970). The Chilean analyses fit well with and are easily within the range of scatter of the empirical lines obtained by Dickinson. This simple observation does not prove, of course, that the magmas originated near the Benioff zone, but it does pose an unhappy coincidence for those who would derive the rocks anatectically.

Analyses are rare of unaltered older plutonic rocks that could provide evidence of earlier positions of the Benioff zone. Some recent analyses given by Dunin-Borkowski (1970) of rocks of the Late Cretaceous or early Tertiary Acari pluton of southern Peru (~ 15.2° S., 74.7° W.) include potash measurements in agreement with those predicted on the basis of depth to the present Benioff zone (Fig. 9), suggesting that the position and dip of the Benioff zone have not changed greatly since the end of the Cretaceous. These few data are not definitive but do point to a comparatively stable Benioff zone, at least during the Cenozoic.

Volumetric Considerations

Some insight into the origin of the igneous rocks that make up the crust beneath the western cordillera can be gained through the simple volume calculations. If we suppose that most of the western cordilleran crust consists of igneous rocks or their sedimentary derivatives, we can easily estimate the total volume of melt that was needed to produce the crust. For simplicity, I have here restricted the calculations to cross-sectional areas along the line A-A′ (Fig. 1). The computations pertain to the vertical section shown in Figure 7 and include that part of the crust to which igneous material has been added since latest Mesozoic. Within this cross section, the area of added crust is ~ 4000 to 8000 km^2, depending upon whether the initial crust was continental (h ~ 30 km) or oceanic (h ~ 6 km, exclusive of water layer). The cross-sectional area of the underlying mantle between the M-discontinuity and the top of the descending plate is ~ 22,000 km^2. Thus, if the magma which produced the Andean crust was derived entirely by partial melting of a noncirculating mantle above the Benioff zone, 18 to 36 percent partial melting is required. This would require an extraordinary amount of partial melting in the mantle and does not warmly support the thesis that the mantle is the sole source of melt. If the magma is generated within the lithospheric plate, the source problem is less severe. Assuming underthrusting of about 6 cm/yr (Le Pichon, 1968) for 60 m.y., the volumetric requirements could be satisfied if 1 to 2 km of the slab was melted. Possibly magma is generated both by melting in the slab and by partial melting of the underlying mantle.

MODEL FOR THE EVOLUTION OF THE CENTRAL ANDES

In this section, I propose a model for the evolution of the central Andes. The presentation of a single model is not meant to exclude the multiple alternatives that exist. Rather, it is my intent here to present what seems, on the basis of the limited observations outlined above, the most plausible of many possible scenarios. Against this backdrop of disclaimers, then, a schematic representation of events from Paleozoic to Quaternary times is presented in Figure 10. The evolution of the central Andean orogen is here envisioned to have resulted by interaction between the South American and Nazca plates which has continued essentially unchanged from Mesozoic to the present time. The model pertains principally to the central Andes situated between about 14° and 22° S. lat; although other areas, particularly in northern Chile, have provided missing elements to the synthesis. The cross section itself is roughly along line A-A′ of Figure 1. Geographic terminology relates to present day.

In late Paleozoic time, the ancestral continental shelf and slope of western South America were blanketed by undeformed Paleozoic deposits totaling 10 to 15 km in thickness. To the west of this shelf area lay a region of sialic crust. Relicts of this crust are now exposed as metamorphic rocks along the coast of southern Peru and northern Chile. The origin of this sialic crust remains something of an enigma. The rocks may have occupied their present position during the Paleozoic time or they may be fragments swept into South America from the Pacific plate, probably near the beginning of subduction. The apparent absence of suture-zone rock assemblages seems to argue against the latter possibility, but evidence of suture zones is commonly not obvious, and there are not yet adequate data in the central Andes to distinguish between the alternatives. The picture is further complicated by post-Paleozoic sedimentary deposition and volcanism which have obliterated much evidence of the crust in the region situated between the coastal metamorphic rocks and the area of Paleozoic shelf deposition.

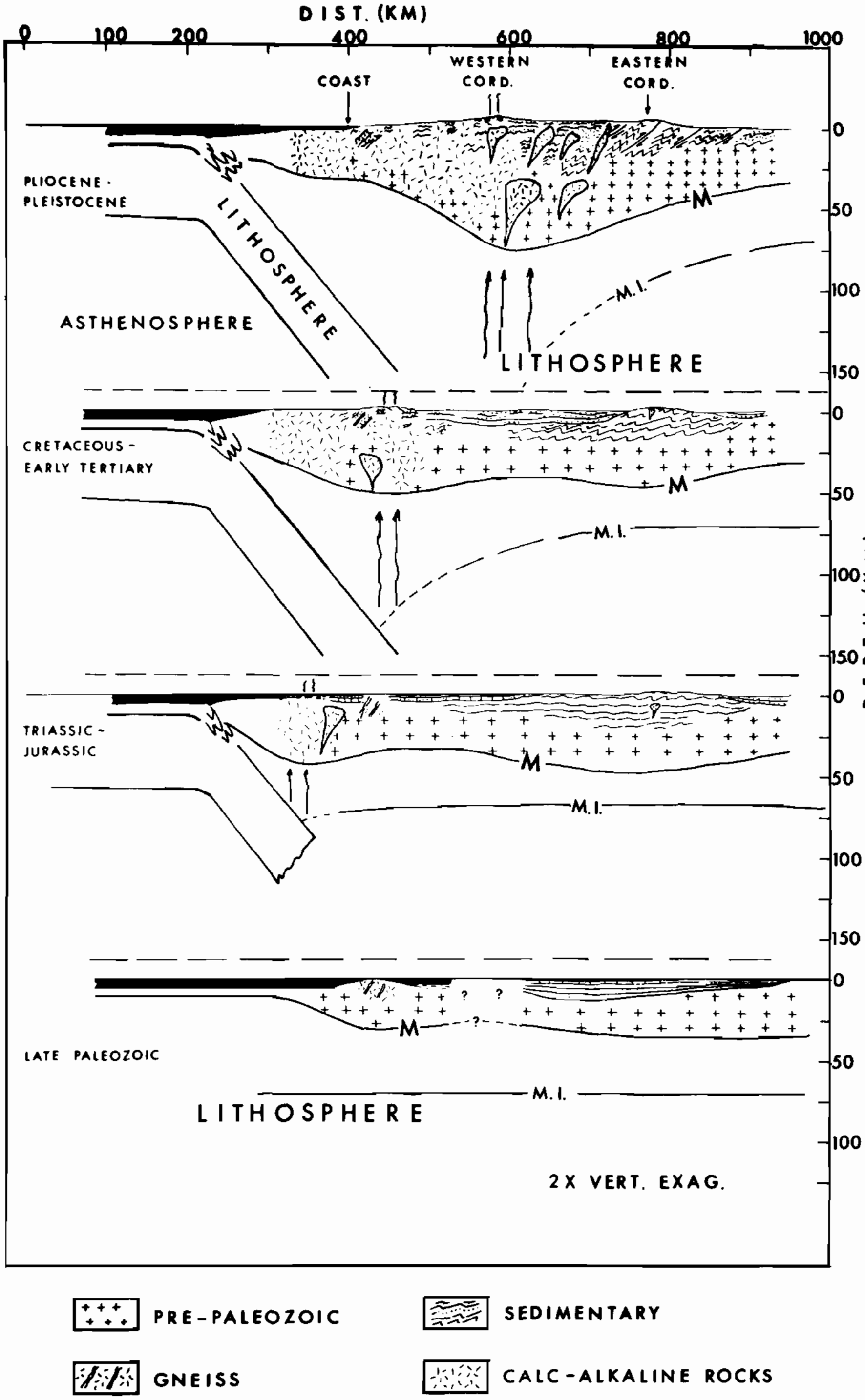

Figure 10. Schematic sequence of cross sections (appropriate for line A-A′ of Fig. 1) illustrating a model for the evolution of the central Andes. Letter "M" marks M-discontinuity. "M.I." denotes "magic isotherm" and indicates the presumed termperature at which melting or water release occurs in the lithospheric slab. Arrows denote rising magma.

Marine deposition in the eastern cordillera ended in mid-Permian time, and underthrusting of the Pacific plate began no later than Late Triassic or Early Jurassic. The start of underthrusting may have coincided with the beginning of spreading on the East Pacific rise, but definitely precedes the breakup of South America from Africa some 135 to 140 m.y. ago (W. Pitman, 1971, personal commun.). During Late Triassic–Early Jurassic time, an incipient volcanic arc developed in the ocean west of the Paleozoic continental shelf (Fig. 10) with the eruption of submarine basaltic and andesitic lavas near the coast of northern Chile and probably along the present continental shelf area of southern Peru. Approximately contemporaneous minor granitic plutonism occurred in the Cordillera Real of the eastern cordillera. Shallow-water marine deposition through mid-Jurassic time in the western cordilleran zone indicates that the crust beneath that area was still comparatively thin.

A major phase of orogeny occurred during Late Cretaceous or early Tertiary time. The zone of this activity is clearly displaced eastward from that of Triassic-Jurassic time (Fig. 10). It was during the Late Cretaceous–early Tertiary episode that the major part of the Andean batholith was emplaced. The crust in the zone of orogeny was invaded by considerable volumes of intermediate to granitic melt; the resultant dilatation of the western cordillera produced compressive stresses which were transmitted laterally to the crust to the east, causing the rocks of the eastern cordillera to be folded and further uplifted. Molasse deposits in enormous quantities were dumped into the intervening altiplano basin, some 7 km in Cretaceous time alone (Newell, 1949). The absence of comparable deposits to the west of the western cordillera suggests that even during the Cretaceous, desert conditions may have prevailed on the western flank of the Andean chain.

The volcanism and emplacement of granitic plutons that began in Miocene time record the beginning of the modern phase of igneous activity in the Andes. This activity appears to have been cyclical, and the Miocene stage was followed by the much stronger Pliocene-Pleistocene orogenic phase. Emplacement of magma into the crust beneath the present Andean crest produced great lateral compressive stresses that caused intense folding and thrust faulting in the altiplano and eastern cordillera. The eastern ranges during the Pliocene-Pleistocene compression were jammed up to form narrow high mountain belts. The altiplano during the Cenozoic received some 15 km of molasse derived from both eastern and western mountain ranges. During late Quaternary time, the waning of igneous activity resulted in relaxation of the compressive stresses and graben structures formed within the altiplano.

It is appropriate and instructive to examine the evolution of the crust in some detail. In the interpretation portrayed in Figure 10, virtually the whole of the crust beneath the western cordillera consists of deep-seated equivalents of the andesitic to silicic volcanogenic sequences and granitic batholiths that are exposed at the surface. Two conclusions are implicit in this interpretation: (1) the lower crust beneath the western cordillera is compositionally equivalent to the upper crust, and (2) the crustal section consists predominantly of rocks derived from near the subduction zone. The first conclusion derives from a consideration of the crustal velocities and appropriate experimental work. Results of Ringwood and Green (1966) and Green (1970) have been summarized at length above, and it remains here only to reiterate the important conclusion of their work. That is, that under the P-T conditions appropriate to the lower crust, the deep crustal velocities of the central Andes are representative of intermediate, not mafic, compositions. It is on this basis that I am encouraged to conjecture that the lower and upper parts of the crust are compositionally similar.

The second conclusion, that most of the western cordilleran rocks are derived from the subduction zone or mantle beneath the crust, is partly conjecture but is supported by evidence of large volumetric additions to the crust. Specifically, well-documented tensional deformation of the western cordillera (Katz, 1970; Kausel and Lomnitz, 1968) indicates that crustal dilatation has accompanied volcanism and magmatic intrusion. In the eastern cordillera, on the other hand, crustal shortening by folding and thrust faulting has probably been the major factor in crustal thickening. In the altiplano, crustal thickening may have been achieved both by molasse deposition and crustal shortening, as well as by igneous intrusion in the western altiplano. The crustal shortening in the eastern ranges and altiplano is here interpreted to have resulted from compression produced by crustal dilatation of the volcanic arc, not by compressive interaction between plates.

It is worth reviewing the implications of deriving the crustal rocks of the western cordillera from the subduction zone. If, as shown earlier, the melt is derived by partial melting of the mantle between the top of the underthrust plate and the base of the crust, some 18 to 36 percent of the mantle material must be melted. Partial melting on this scale has been proposed for areas such as Hawaii where magmas of predominantly tholeiitic composition are erupted; however, even 20 percent partial melting of mantle rocks to produce intermediate to silicic melt seems improbable. Alternatively melting of 1 or 2 km of the downgoing plate could produce the necessary volumes of material. Possibly, crustal rocks are derived by a combination melting of downgoing plate and overlying mantle. It may be also that some source material derives from erosion of the South American plate. If the leading edge of the continental plate is progressively crumpled and underthrust at the trench, the crustal rocks dragged down could be melted to provide magma for the continental crust. There are obvious problems, however, in transporting light sialic material to great depths in the mantle.

DISCUSSION

The results and interpretations that have been presented in this paper leave many questions unresolved and pose a number of new ones. Traditionally, the Andean orogenic belt has been interpreted in terms of classical concepts of geosynclines. Major revision of ideas of eugeosynclinal-miogeosynclinal evolution of cordilleran belts (Dickinson, 1971) leads by necessity to a re-evaluation of classical concepts. It is for this reason that I have avoided use of geosynclinal terminology in this paper. It is useful, however, to examine briefly what insights the Andes may provide as to the nature of geosynclines.

Classical concepts of geosynclines held that elongate basins of deep subsidence were necessary precursors to later mountain belts which were presumed to have formed by deformation and anatectic melting of the thick eugeosynclinal sedimentary section. Although the Andean orogenic belt has been interpreted in geosynclinal terms, it has never fit readily into the mold, for it is clear that no depositional eugeosynclinal basin has ever existed. Dickinson (1971) has reconstructed geosynclinal theories within the framework of plate tectonics and in so doing has recast the concepts into a form that bears a clear relation to the Andean orogen. By this revised view of geosynclines, the so-called eugeosynclinal sequences within the western cordillera have been derived from the volcanic arc itself as the arc has evolved.

In the normal cordilleran arc setting, there are at least three distinct sedimentary assemblages (Dickinson, 1971): (1) the complex mélange of trench turbidites and ophiolites which are ground under at the trench; (2) marine sedimentary and graywacke deposits that accumulate in sedimentary troughs situated between arc and trench; and (3) the volcaniclastic wedge deposited in foredeeps behind the arc, away from the trench.

Of these three types of "eugeosynclinal" assemblages, only the foredeep deposits are of the great consequence in the Andes. This is probably due in part to the fact that the western slopes of the Andean volcanic chain are desert, so that erosion proceeds slowly. These desert conditions have been cited to explain the absence of turbidite deposits in the trench off northern Chile (Scholl and others, 1970). The remarkably high correlation between rainfall and thickness of turbidites in the trench appears too good to be coincidental and provides a simple explanation for an otherwise puzzling observation. Similarly, there appears to be little sedimentation between trench and coast. No serpentinite or ophiolite belt has been observed in the central Andes. If ophiolites and related mélange assemblages represent underthrust oceanic crust and trench deposits, observation of these rocks is presumably not likely until subduction ceases and the underthrust assemblages rebound isostatically.

Because desert conditions preclude much erosion and deposition on the western flanks of the mountains, the ranges are asymmetrically eroded from their wet eastern flanks. The deposition of detritus derived from both western and eastern cordilleras has resulted in an abnormally thick "foredeep" clastic wedge in the intervening altiplano trough.

The "miogeosynclinal" deposits are here interpreted to be the Paleozoic marine continental shelf deposits that accumulated during a long period of quiescence at the western margin of the Paleozoic South American continent. A modern analogue of these deposits are the thick sequences that lie on the continental shelf of the eastern United States (Drake, 1966).

Finally, it is important to emphasize the eastward migration of igneous activity. If this migration is characteristic of island arcs, polarities of dormant subduction zones can be deduced from the direction in which igneous rock ages decrease. Moreover, it is possible that even the rate of underthrusting can be determined from the rate at which ages migrate. The younging of igneous rocks provides, together with such other empirical observations as K-h relations, yet another means of untangling the tectonic histories of older orogenic belts.

ACKNOWLEDGMENTS

The interpretations and conclusions presented in this paper are entirely my responsibility, but many of the essential ideas and concepts would not have been possible without extended discussions with many colleagues. I especially thank S. R. Hart for innumerable stimulating discussions and suggestions and H. L. James both for enthusiastic interest and discussion and for pointing out that igneous activity commonly youngs with distance from the trench. Special appreciation is due R. P. Morrison, who not only expressed active interest but generously supplied preprints of parts of his forthcoming book on South American geology. Finally, I wish to thank L. M. Dorman, O. B. James, W. Hamilton, and W. R. Dickinson for critically reading the manuscript and providing several important suggestions.

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MANUSCRIPT RECEIVED BY THE SOCIETY APRIL 26, 1971
REVISED MANUSCRIPT RECEIVED JULY 8, 1971