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Sierra Nevada Batholith

The batholith was generated within a synclinorium.

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How were the great batholiths of granitic rock formed? Geologists have pondered this question for two centuries and still cannot agree on an answer. The Sierra Nevada includes one of the world's largest and best exposed batholiths—a splendid subject for study. Only recently, however, has enough systematic and detailed geologic mapping been completed to permit deciphering of its internal structure and its setting. The emerging geological picture has been supplemented during the past few years by geophysical and geochemical studies, and hypotheses as to the origin of the batholith can now be based on data from several different disciplines.

The Sierra Nevada and the Sierra Nevada batholith are somewhat different entities. The Sierra Nevada is a magnificent mountain range in eastern California that was formed by the westward tilting and consequent dissection, in late Cenozoic time, of a huge block of the earth's crust (*1*). It is composed chiefly of Mesozoic granitic rocks and metamorphosed Paleozoic and Mesozoic sedimentary and volcanic rocks. These rocks are discontinuously overlain in the north by Cenozoic volcanic sheets that extend southward from the Cascade Range, and they are overlapped on the west by sedimentary rocks of the Great Valley of California (Fig. 1).

The Sierra Nevada batholith, on the other hand, is the part of the Sierra Nevada that is underlain chiefly by granitic rocks. These rocks, which range in composition from quartz diorite to granite, make up most of the southern half and the eastern part of the northern half of the range. They are part of a more or less continuous belt of granitic rocks that extends from Baja California northward through the Peninsular Ranges and the Mojave Desert, through the Sierra Nevada, and into western Nevada.

In a world view, the Sierra Nevada batholith is part of a belt of Mesozoic plutonic rocks that encircles the Pacific Basin (Fig. 2). Mesozoic plutonic rocks are as characteristic of the Pacific margins as are oceanic deeps, the belt of fire, active earthquake faults, and graywacke sandstones. When enough information becomes available so that a geologic history of the Pacific Basin can be written, one very significant chapter will be based on studies of the Mesozoic batholiths.

Most of the detailed studies of the Sierra Nevada batholith have dealt with a particular region—a wide belt that crosses the Sierra Nevada between latitudes 36°45′N and 38°00′N, and much of what we have to say pertains specifically to this belt (Fig. 3). Work in other parts of the range indicates, however, that this belt is representative and that concepts developed here can be extended to the range as a whole.

Country Rocks

The Sierra Nevada batholith was intruded into strongly deformed but weakly metamorphosed strata of Paleozoic and Mesozoic age, which are preserved in the walls of the batholith and in remnants within the batholith. In the northern half of the range, the batholith is flanked on the west by the western metamorphic belt, which is the site of the famed Mother Lode, but the eastern contact of the batholith is hidden beneath surficial deposits of Cenozoic age. However, the composition, structure, and age of the eastern wall rocks are shown in roof pendants along the crest of the range and in desert ranges east of the southern Sierra Nevada (Fig. 1).

The Paleozoic strata along the east side of the batholith consist chiefly of fossiliferous shale, quartzite, limestone, and dolomite, whereas the Paleozoic strata in the western metamorphic belt are characterized by sparsely fossiliferous or unfossiliferous graywacke sandstone, slate, chert, and andesitic volcanic rocks. This distribution of rock types indicates a transition from miogeosynclinal facies east of the batholith to eugeosynclinal facies in the western metamorphic belt. The line of transition may trend southwestward diagonally across the batholith, on the extension of the boundary between the carbonate "eastern" facies and the volcanic and clastic "western" facies of the Paleozoic rocks in Nevada (Fig. 1).

The Mesozoic strata are graywacke, slate, and volcanic rocks, and they are eugeosynclinal. Crossbeds and fossiliferous sequences in miogeosynclinal rocks and graded beds in eugeosynclinal rocks of both Paleozoic and Mesozoic ages provide a means of determining the top directions of beds and of unraveling the complex structures of the rocks. The major structure in

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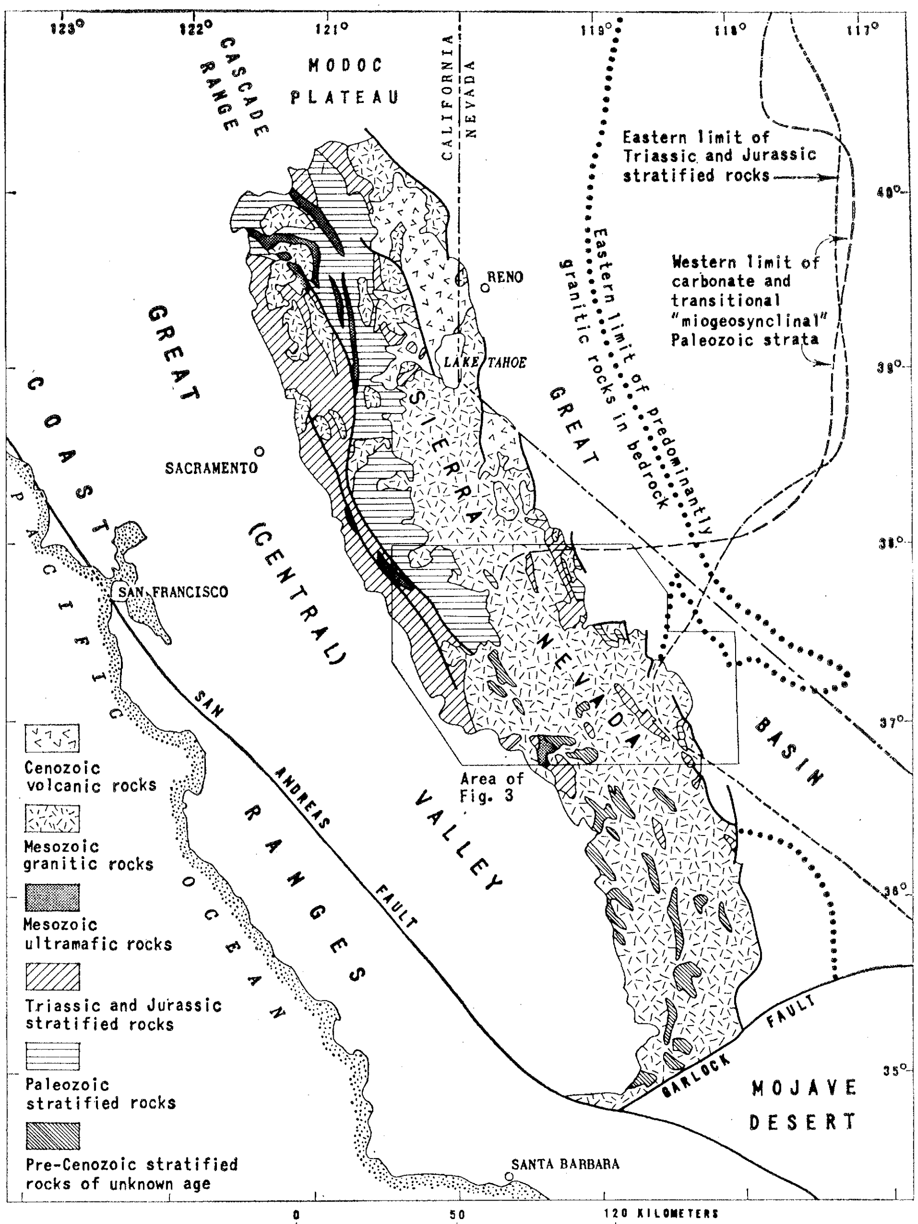

Fig. 1. Generalized geologic map of the Sierra Nevada and adjacent areas.

the stratified rocks of the central Sierra Nevada (Fig. 3) is a complex faulted synclinorium (*2*). This structure is indicated by the predominance of eastward-facing over westward-facing bedding tops in the tightly folded strata of the western metamorphic belt; by westward-facing bedding tops in thick, steeply dipping homoclinal sequences in roof pendants of the eastern Sierra Nevada; and by the progressively decreasing age of strata toward the middle of the batholith from both sides, except where faults interrupt the sequence. This synclinorium is not readily apparent in the patterns of geologic maps, chiefly because only fragments of the framework rocks remain and because strike faults of large displacement interrupt the sequence of strata in the western metamorphic belt.

Unconformities within and between the Paleozoic and Mesozoic units and structures in the stratified rocks indicate repeated movement since middle Paleozoic time. Probably the faulted synclinorium began to take form in Permian or Triassic time, and intermittent disturbances occurred throughout the Jurassic and into the Cretaceous. The very severe disturbance that took place near the close of the Jurassic and caused the principal folds in the Upper Jurassic strata of the western metamorphic belt is referred to as the Nevadan orogeny, but both earlier and later disturbances have been identified (*3*, *4*).

In any given area, the first deformation, regardless of when it occurred, generally preceded the emplacement of adjacent plutonic rocks and produced the largest folds in stratified rocks. In many places plutonic rocks emplaced after a first deformation were deformed during later deformations, indicating overlap of episodes of magmatism with episodes of deformation.

Structure of the Batholith

The Sierra Nevada batholith is composed chiefly of quartz-bearing rocks that range in composition from quartz diorite to granite, but it includes scattered smaller masses of darker and generally older plutonic rocks and remnants of metamorphic rocks. Rocks in the compositional range of quartz monzonite and granodiorite predominate and are about equally abundant. In general, the plutonic rocks in the western part of the batholith are more mafic than those in the eastern part.

The granitic rocks are in discrete masses or plutons, which generally are in sharp contact with one another or are separated by thin septa of metamorphic or mafic igneous rocks. Most large plutons are elongate in a northwesterly direction, parallel to the long axis of the batholith, but many smaller ones are elongate in other directions or are irregularly rounded. Individual plutons range in outcrop area from less than 1 square kilometer to more than 1000 square kilometers; the limits of many large plutons have not yet been delineated. The rocks in the different plutons can be readily distinguished from one another in the field by differences in mineral composition and texture. Where two plutons meet, it usually is possible to determine which is the older, from inclusions and dikes of one in the other and from truncated structures (*5*). The following observations indicate that the granitic plutons rose from below as magma (partial melt): sharp contacts of plutons with wall rocks and with one another; dikes and inclusions along contacts between plutons, from which the relative ages of plutons can be determined; finer grain size in apophyses and in the margins of some plutons; dilated walls of dikes; and deformation that can be attributed to the forcible emplacement of plutons.

In the western foothills, granitic

rocks intrude Upper Jurassic strata, which in turn are unconformably overlain by Upper Cretaceous strata. On the basis of these relations the granitic rocks of the Sierra Nevada have been referred to as Jurassic or as Cretaceous, no breakdown within the period being made other than assignment of relative ages as indicated by contact relations. Calkins (*6*) mapped two series of granitic formations in the Yosemite region in which the plutons are successively younger toward the east, and in the past many geologists have felt that this pattern, together with the generally more felsic compositions of the granitic rocks in the eastern Sierra, indicated that the plutons were progressively younger toward the east. However, isotopic age determinations indicate that the pattern of intrusion is more complicated. The dates obtained, including many unpublished dates, from the central Sierra Nevada indicate that most of the granitic rocks along the east side of the batholith and in adjacent areas to the east are 170 to 210 million years old (Late Triassic or Early Jurassic); that plutons in lower Yosemite Valley and others intruded into the western metamorphic belt are 125 to 145 million years old (Late Jurassic); and that plutons along and just to the west of the range crest are 80 to 90 million years old (early Late Cretaceous) (*7*). In addition, one large, possibly composite, unit in the western half of the batholith has yielded potassium-argon age determinations on hornblende ranging from 88 to 115 million years; many of these age values are doubtless too recent, as a result of reheating of the plutons at the time of intrusion of younger magmas. The ages of many other mapped plutons have not been firmly established, and some of these plutons may have been emplaced during magmatic episodes that have not yet been recognized.

The major features of crustal struc-

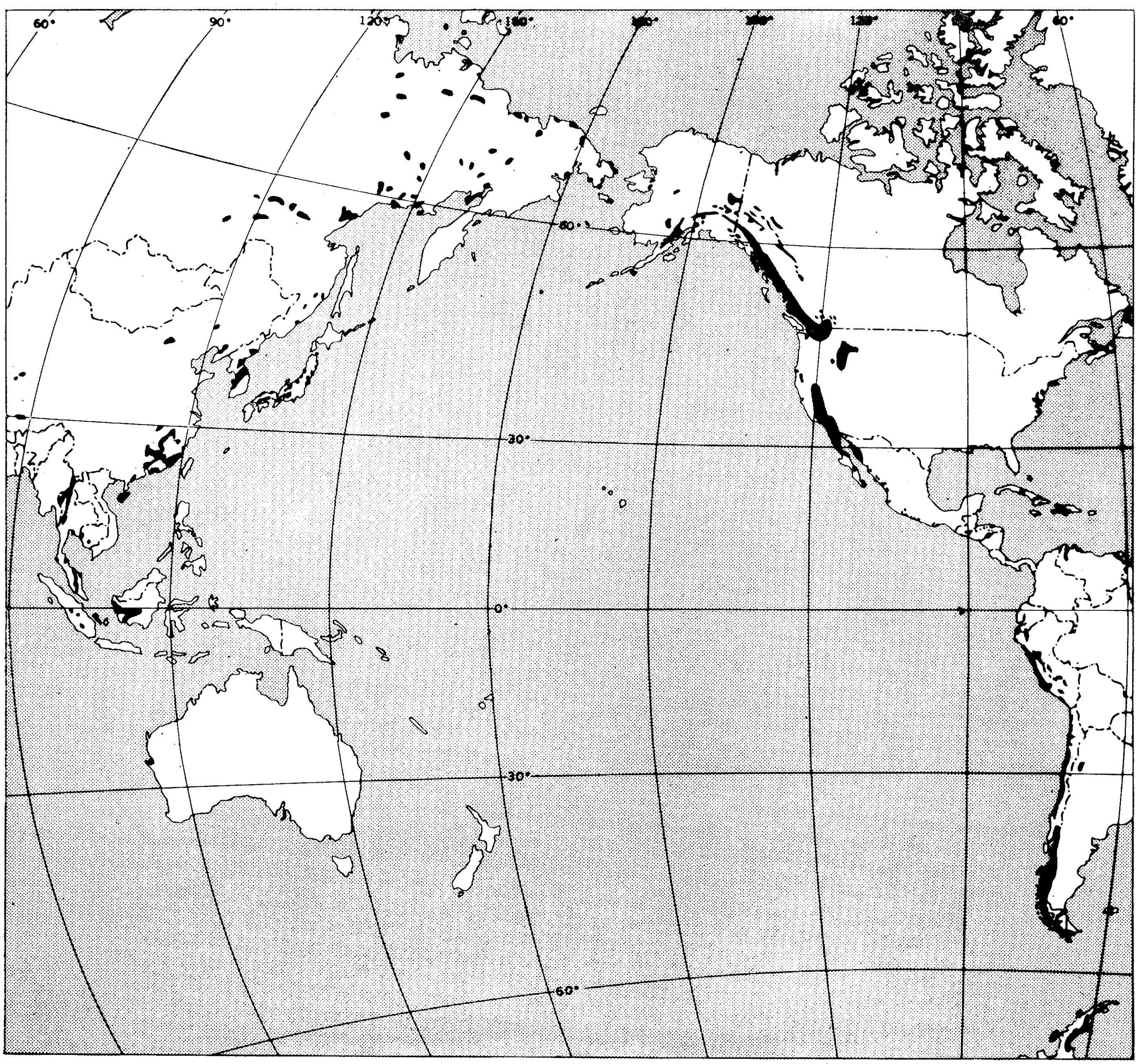

Fig. 2. Map showing that the principal granitic batholiths of Mesozoic age are marginal to the Pacific Basin.

ture beneath the Sierra Nevada and neighboring provinces have been outlined during the past 6 years by seismic refraction profiles. Profiles that pass longitudinally through these ranges have determined seismic P-wave velocities and thicknesses of principal crustal units beneath the Sierra Nevada and central Coast Ranges, and transverse profiles crossing the Coast Ranges, Great Valley, and Sierra Nevada, from shot points offshore near San Francisco and San Luis Obispo, have set limits on the manner in which crustal structure varies between the Coast Range and Sierra profiles (*8*). Profiles in the Basin and Range province, including some that cross the boundary between the Basin and Range and the Sierra Nevada provinces, delineate the crustal structure east of the Sierra Nevada in considerable detail (*9*).

From these data we have constructed a seismic cross section transverse to the Sierra Nevada from the Pacific shore to the central Basin and Range province (Fig. 4). The cross section passes through Fresno and crosses the steep eastern face of the Sierra Nevada 22 kilometers northwest of Bishop. Its most striking feature is a trough-like depression of the Mohorovicic discontinuity beneath the Sierra Nevada batholith. This depression extends about 50 kilometers east of the crest of the Sierra Nevada proper into the western edge of the Basin and Range physiographic province and underlies the "Sierran Highland Region" of Thompson and Talwani (*10*), which is approximately coincident with the batholith. From a depth of about 24 kilometers below sea level beneath the Coast Ranges, the base of the crust descends eastward to slightly deeper than 52 kilometers below sea level beneath the crest of the Sierra Nevada and then rises to only 27 kilometers below sea level beneath the western edge of the Basin and Range province. The velocity of P-waves in the upper mantle decreases from 8.0 km/sec beneath the Coast Ranges to 7.9 km/sec beneath the Sierra Nevada and to 7.8 km/sec beneath the Basin and Range province. Just east of the center of the Great Valley, coincident with a magnetic and gravity high that runs the length of the valley (*11*), the top of the mantle may bulge upward several kilometers, perhaps to a depth as shallow as 18 kilometers.

In both the Coast Range and the Basin and Range provinces, the crust is divided into two zones or "layers." The upper three-fifths of the crust has a P-wave velocity of about 6.0 km/sec in both areas, whereas the lower two-fifths has a P-wave velocity of about 6.8 km/sec beneath the Coast Ranges and about 6.7 km/sec beneath the Basin and Range province. Beneath the Sierra Nevada the lower part of the crust constitutes slightly more than half the entire crust and

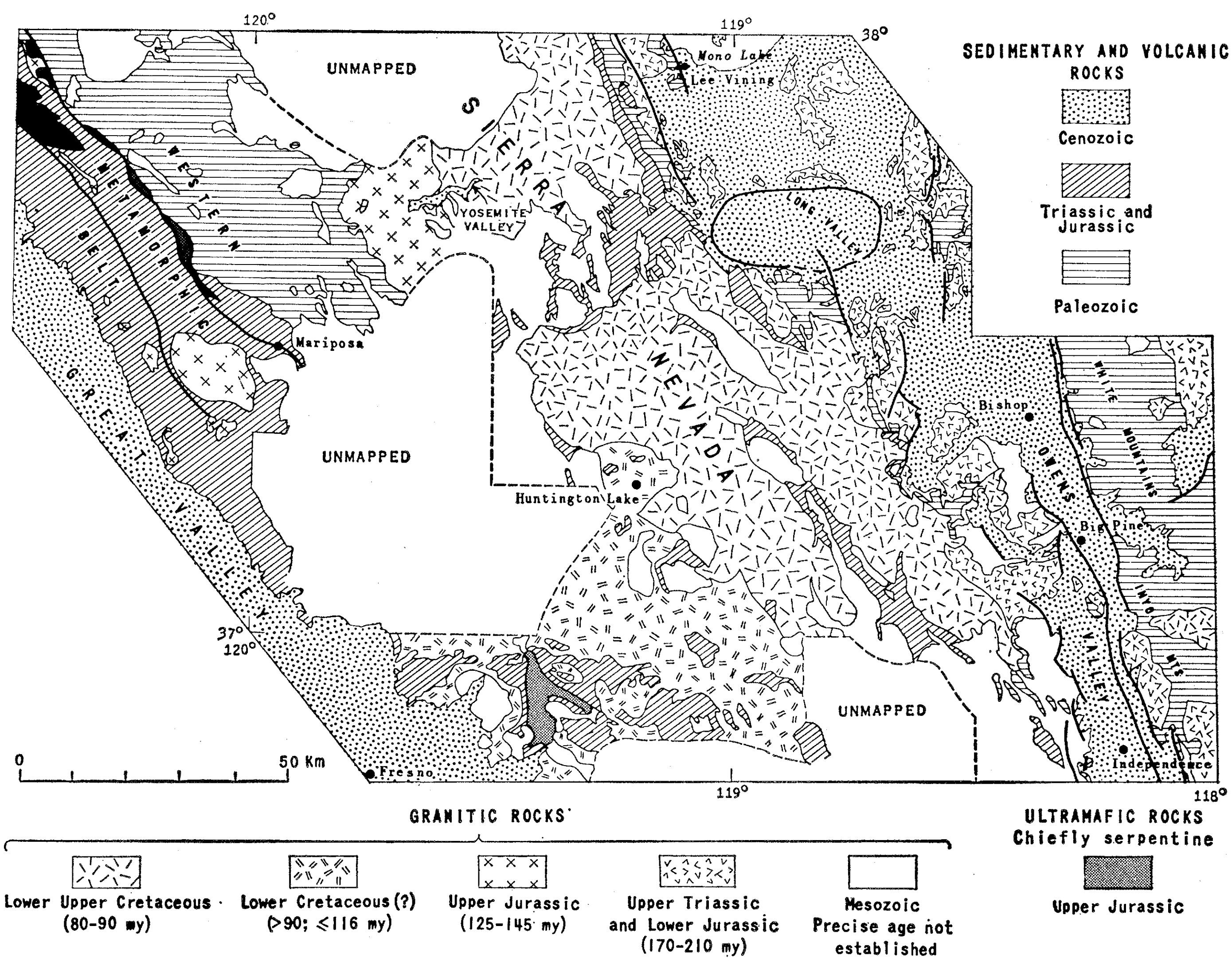

Fig. 3. Geologic map of the central Sierra Nevada showing the distribution of plutonic and metamorphic rocks.

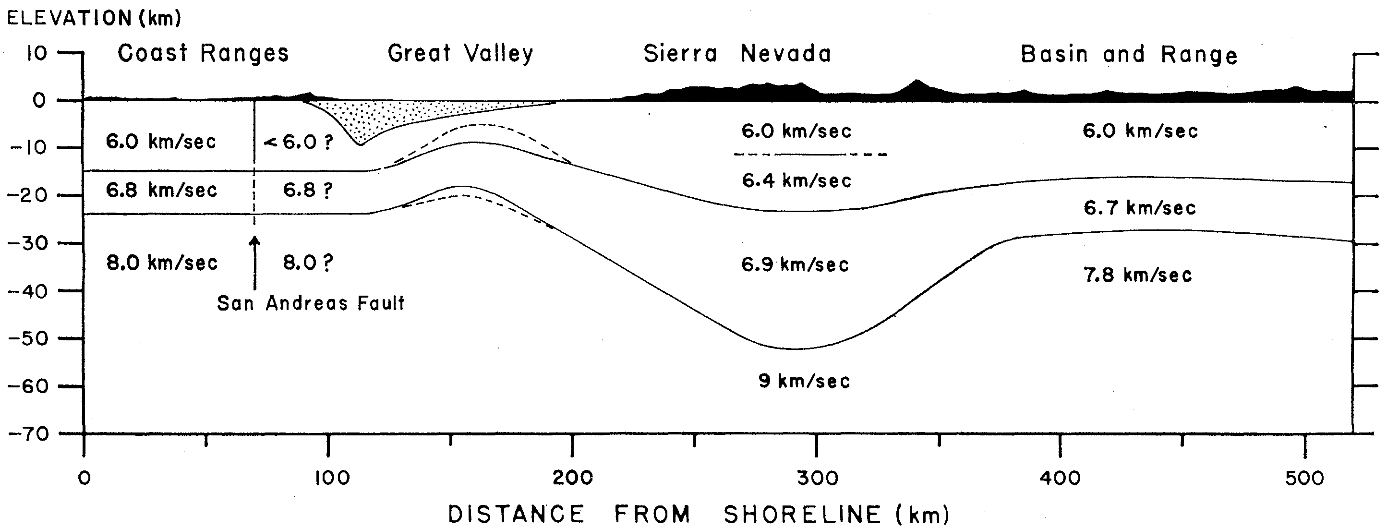

Fig. 4. Transverse section, based on seismic refraction measurements, of the Coast Ranges, Great Valley, Sierra Nevada, and Basin and Range province. The line of section passes through Fresno, California, and crosses the eastern face of the Sierra Nevada 22 kilometers northwest of Bishop, California. Solid and dashed lines beneath the eastern part of the Great Valley show alternative explanations of early P-wave arrivals in that region.

has a P-wave velocity of about 6.9 km/sec. The upper part of the crust is further subdivided: rocks with a P-wave velocity near 6.0 km/sec extend to a depth of 11.5 kilometers, where they are underlain by rocks with a P-wave velocity of 6.4 km/sec which extend to a depth of nearly 24 kilometers.

Properties and structure of the crust between the San Andreas fault and the western flank of the Sierra Nevada are not well established. Beneath the thick, unconsolidated sediments of the Great Valley, which descend to a depth of about 10 kilometers, the "Sierran" basement rocks on the east give way to rocks of the Franciscan Formation on the west. Moreover, the early arrival of seismic P-waves on the transverse profiles at the eastern edge of the Great Valley, which is interpreted above as indicating an upward bulge of the Mohorovicic discontinuity (Fig. 4, solid curves) may actually be caused by the virtual elimination of the upper layer of the crust by thickening of the rocks of the lower crust (Fig. 4, dashed curves).

Crustal Thickness and Isostatic Balance

It is common practice to test the crustal model obtained from the interpretation of a seismic refraction profile by comparing values of gravity computed from the seismic model with actual gravity observations along the profile. One perilous step in making such a comparison is the choice of rock densities for the seismic model on the basis of observed P-wave velocities and some empirical relationship between velocity and density. Another problem is the choice of an appropriate scale and method of comparing the gravity consequences of the seismic model with observed gravity. Small shallow features that are not adequately delineated on the seismic profile may have pronounced local gravity effects, and assumptions made in reducing observed gravity values to Bouguer or isostatic gravity anomalies can obscure the features that are to be compared (*12*).

Some ambiguity in the dependence of density on velocity in rocks is removed by the introduction of mean atomic weight as a parameter: for rocks of the same mean atomic weight, a simple linear relationship, with much reduced scatter of data points, is obtained (*13*). However, the rocks of the crust and upper mantle appear to vary somewhat systematically in mean atomic weight with depth, so a velocity-density curve for rocks of a single mean atomic weight cannot be applied indiscriminately to an entire velocity-versus-depth profile (*14*).

Measurements of P-wave velocities in a wide variety of rocks under confining pressures up to 10 kilobars have been reported by Birch (*15*) and Christensen (*16*), who presented compositional and textural data on the rocks they studied. Data from these sources for a sequence of igneous and metamorphic rock types that may constitute major units of the crust and mantle are presented in Fig. 5. Approximate corrections were applied to reported density and velocity values to account for changes in dimensions of test specimens subjected to high confining pressures. For the corrections, compressibilities were computed from zero-pressure densities, 2-kilobar velocities, and V_p/V_s ratios (V_p and V_s are P-wave and S-wave velocities, respectively) at 1 to 4 kilobars (*15*). Each rock type is represented by five points on the graph, corresponding to confining pressures of 1, 2, 4, 6, and 10 kilobars.

Data on the dependence of velocity in rocks on temperature, at high temperatures and pressures, are inadequate. Some measurements for temperatures up to 400°C and pressures up to 6 kilobars have been reported (*17*). For one specimen (San Marcos Gabbro), results for the range 2 to 6 kilobars and 25° to 300°C suggest that velocity variations correlate directly with changes in density, regardless of which environmental variable produces them. Increase of temperature with depth was not considered in the data displayed in Fig. 5.

In drawing a curve to represent density as a function of velocity in rocks of the upper crust and mantle, the approximate lithostatic pressure at which a given velocity was actually encountered was taken into consideration. Approximate pressures at the tops of refracting horizons beneath all four regions are indicated in Fig. 5. The curve was also drawn in accordance with the hypothesis that the composi-

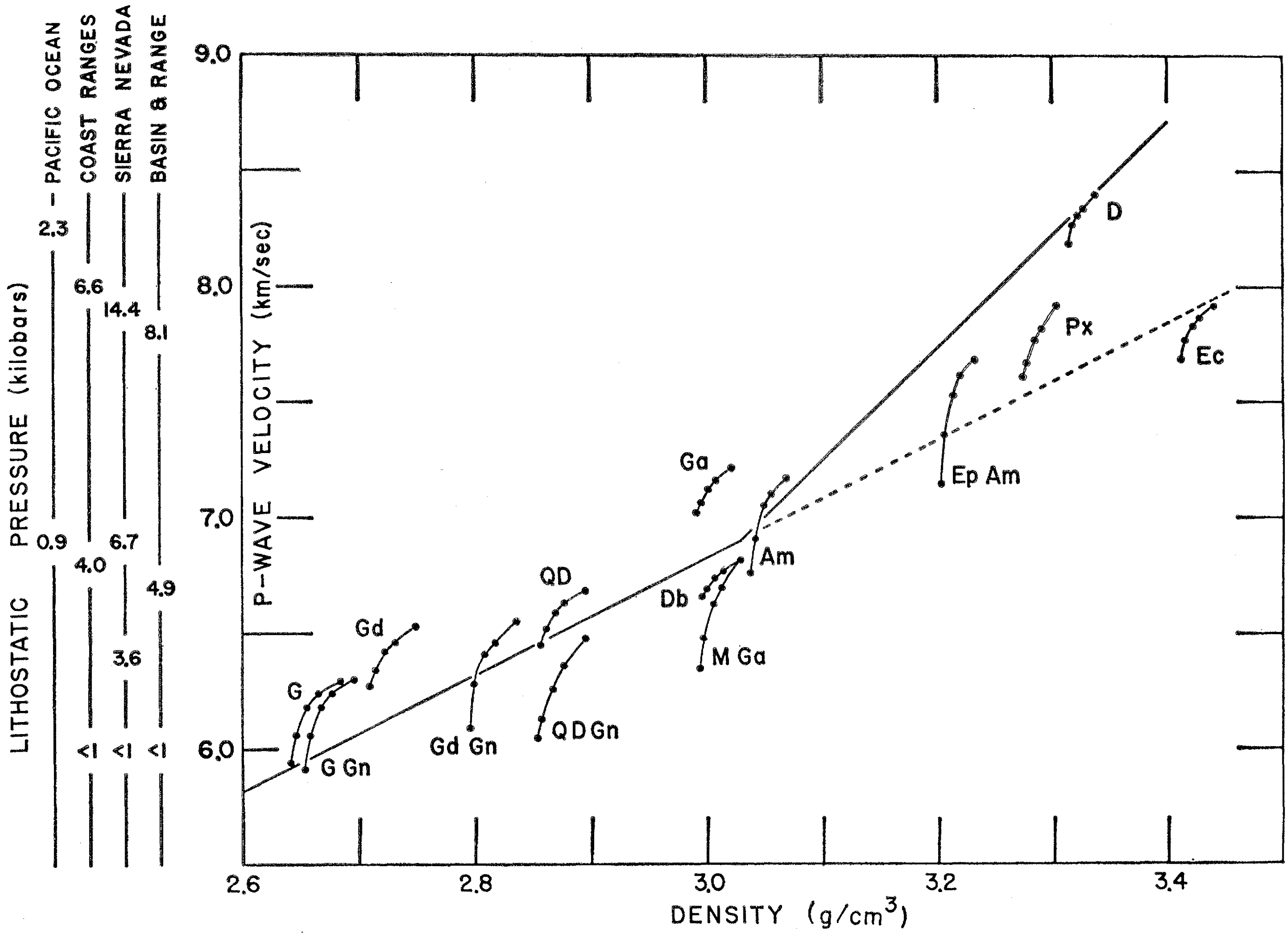

Fig. 5. Experimental data (*15, 16*) on P-wave velocities as a function of confining pressure for selected igneous and metamorphic rocks. From bottom to top, data points for each rock type correspond to pressures of 1, 2, 4, 6, and 10 kilobars. The solid curve based on these data was used to estimate densities of the rocks composing the seismic columns from their velocities. Approximate lithostatic pressures corresponding to the tops of the seismic "layers" are plotted opposite the appropriate layer velocities at left. *G*, Granite from (i) Westerly, Rhode Island, (ii) Chelmsford, Massachusetts, and (iii) Barre, Vermont, and quartz monzonite from Porterville, California (*15*); *GGn*, gneiss 1 and gneiss 2 (*16*) from Berkshire Highlands; *Gd*, granodiorite from Butte, Montana (*15*); *GdGn*, gneiss 3 and gneiss 4 (*16*) from Berkshire Highlands; *QD*, quartz diorite from San Luis Rey, California, and Dedham, Massachusetts (*15*); *QDGn*, gneiss 5 from Berkshire Highlands (*16*); *Ga*, gabbro from Mellon, Wisconsin, and French Creek, Pennsylvania, and norite from Pretoria (*15*); *Db*, diabase from (i) Holyoke, Massachusetts, (ii) Centerville, Virginia, (iii) Sudbury, Ontario, and (iv) Frederick, Maryland (*15*); *MGa*, metagabbro from Hodges Mafic Complex (*16*); *Am*, amphibolite 1 and amphibolite 2 (*16*) from Mount Tom; *EpAm*, epidote amphibolite 1 and epidote amphibolite 2 (*16*) from Hodges Mafic Complex; *Px*, pyroxenite from Sonoma County, California, and bronzitite from Stillwater Complex and Bushveld Complex (*15*); *D*, dunite from Twin Sisters Peaks, Washington (*15*); *Ec*, eclogite from Healdsburg, California, and Kimberly (*15*).

tion of the continental crust, which is observed to be in the range of granite to granodiorite in its upper part, becomes progressively more mafic with depth, and that the composition of the upper mantle is approximately that of dunite.

The observation that the free-air gravity anomaly averages nearly zero, when averaged over large areas, indicates that large areas are generally in isostatic balance with one another. Isostatic balance requires that the total weight, per unit area, of the rocks above some reference level (called depth of compensation) be the same from one region to the next, and it provides a simple means of subjecting crustal models from seismic interpretations to a simple gravity test (*18*).

Averaged seismic structures of the Coast Ranges (0 to 70 kilometers from the Pacific shoreline), high Sierra Nevada (250 to 350 kilometers), and Basin and Range (400 to 500 kilometers) portions of the seismic cross section in Fig. 4 were compared with each other and with a Pacific Ocean "seismic" crust (*19*), according to the criterion of isostatic balance. Crustal units were assigned densities on the basis of P-wave velocities according to the curve in Fig. 5. The mass per unit area of the rocks above each boundary in each crustal column, as well as at a standard depth within the mantle beneath each column, was then computed. Finally, values for the oceanic profile were subtracted from those for the continental profiles to obtain curves showing the mass excess per unit area as a function of depth in the continental profiles with respect to the oceanic profile. Velocity-density columns for all four regions and the curves for mass excess plotted against depth for the continental columns are shown in Fig. 6.

Despite large differences within the crust, the mass-excess curves converge toward zero at depths of 50 to 60 kilometers. Much of the "compensation" for differences in surface elevation and crustal structure is accomplished through lateral variations in density within the upper mantle. The P-wave velocities on which the deepest parts of the mass-excess curves are based apply strictly only to the uppermost mantle, and the depths at which these curves decrease to zero suggest minimum values for the depth of compensation. The actual depth of compensation depends on the manner in which regional differences in the upper mantle diminish with depth.

Gross Crustal Structure

The bottom cross section in Fig. 7 illustrates our concept of the gross structure across the central Sierra Nevada. The section is highly generalized and is composite; the western half is drawn at the latitude of Yosemite to show the wall rocks west of the batholith, and the eastern half is drawn farther south at the latitude of Bishop to show the wall rocks on the east side in the White Mountains. The deeper crustal rocks are shown as amphibolite or gabbro-basalt in accordance with their P-wave velocity of 6.9 km/sec (Fig. 5) and as having the composition of the mafic inclusions, which are ubiquitous in most plutons of granodiorite and probably are mostly refractory material brought upward from depth (*2*). Downward increase in the P-wave velocity in the upper half of the crust from 6.0 km/sec near the surface to 6.4 km/sec suggests downward increase in the proportion of diorite, quartz diorite, and calcic granodiorite (or their gneissic equivalents) relative to quartz monzonite and granite.

Both the Precambrian and the Paleozoic strata are shown as thinning to westward and the Precambrian is shown as dying out beneath the Sierra Nevada, because farther west, in the Coast Ranges, the Mesozoic Franciscan Formation appears to rest directly on oceanic crust (*20*).

The two reverse faults that cut the wall rocks west of the batholith represent the two largest faults identified by Clark (*3*). Serpentinized ultramafic rocks occur along these faults and suggest deep penetration, possibly into the upper mantle.

The localization of the batholith in the axial region of a synclinorium of great size and depth leads us to consideration of the hypothesis that the granitic magmas were generated by the melting of relatively fusible crustal rocks as a result of their being depressed into deeper regions of higher temperature (a process known as anatexis). This hypothesis has enjoyed wide popularity because the rocks of the upper crust yield a high proportion of granitic magma whereas basaltic and ultramafic rocks of the lower crust and mantle yield embarrassingly small proportions of granitic magma. In the Sierra Nevada, correspondence of compositional differences in the granitic rocks in the two sides of the batholith to compositional differences in the country rocks suggests that the crustal rocks provided significant amounts of material for the generation of the granitic magmas.

The great amplitude of the synclinorium virtually requires the conclusion that the Sierran root was formed during the Mesozoic, contemporaneously with the synclinorium, rather than during the late Cenozoic uplift of the range as has been suggested by Christensen (*21*) and by Hamilton and Myers (*22*). Formation of the root during the Mesozoic is supported by the fact that the root, as determined both by seismic and by gravity studies (*10*), coincides geographically more closely with the batholith than with the range. Also, the upper crust and lower crust thicken coincidently toward the axis of the batholith, suggesting vertical redistribution of material, which implies that the lower crust originated at the same time as the upper crust, during the Mesozoic.

Arguments advanced for the formation of the root during the late Cenozoic are motivated largely by need for a mechanism to explain the uplift of the range at that time. However, the Sierra Nevada was tilted westward during the time interval in which adjacent ranges to the east were tilted and uplifted, so the Cenozoic uplift of the Sierra Nevada appears to be related to the broader problem of the origin of Basin and Range structure. The hypothesis that the root formed beneath the Sierra Nevada during the Cenozoic would not solve this broader problem.

In the Sierra Nevada the rocks that were melted doubtless included Paleozoic and Mesozoic strata and, in the eastern half of the batholith, Precambrian quartzose and micaceous schist, amphibolite, and granitic gneiss. In addition to these sialic rocks, mafic material from the lower crust and perhaps the mantle must have been incorporated in the parent granitic magmas to supply the required calcium, iron, and magnesium, although upper Precambrian and Paleozoic carbonate formations like those exposed in roof pendants of the eastern Sierra Nevada and in desert ranges farther to the east doubtless supplied some calcium and magnesium. Ultramafic and mafic intrusions in the western metamorphic belt, and mafic flows, tuffs, and intrusives there and elsewhere, indicate that mafic material was introduced into the upper crust and must have been depressed with the enclosing rocks into the zone of melting. Probably of less importance, mixing of sialic and mafic material could also have occurred at the base of the sial, in the zone of melting. Experimental studies show that, under water-vapor pressure of 10 kilobars, basalt begins to melt at about 625°C (*23*), almost the same temperature at which synthetic granite begins to melt at that pressure.

Differences in the composition of the granitic rocks in the two sides of the batholith can be explained in terms of the rocks that were melted. Moore (*24*) inferred that systematically higher K_2O/Na_2O ratios in the granitic rocks east of his quartz diorite line were caused by the rocks there having been generated within a thick sialic layer, in contrast to those west of his line, which were generated within the sima or within a thinner sialic layer containing a great thickness of associated geosynclinal sedimentary and mafic volcanic rocks. Moore's quartz diorite line runs through the Sierra Nevada parallel to the long axis of the range and, in the central Sierra Nevada, coincides with the west boundary of the lower Upper Cretaceous granitic rocks.

In support of Moore's argument is the finding that the granitic rocks in the western part of the central Sierra Nevada are more mafic and apparently have a lower initial Sr^{87}/Sr^{86} ratio than rocks in the eastern part; basalt contains much more calcium, iron, and magnesium, and has a substantially lower initial Sr^{87}/Sr^{86} ratio, than sialic rocks. A recent isochron based on whole-rock rubidium-strontium determinations, by Kistler (*25*), for the El Capitan Granite and related granitic rocks of Late Jurassic age in the lower

Yosemite Valley indicates an initial Sr^{87}/Sr^{86} ratio of 0.704 ± 0.0008, whereas Hurley *et al.* (*26*) obtained an initial ratio of 0.707 ± 0.0004 for lower Upper Cretaceous and Triassic and Lower Jurassic rocks of the east-central Sierra Nevada.

A reasonable interpretation of the compositional, geographic, and temporal relations of the granitic rocks is that magma was generated several times, each time in a somewhat different place as a result of shift in the locus of downfolding (Fig. 7). The picture envisaged is this: The upper crust was downfolded along the continental margin, possibly as the result of convective overturn in the earth's mantle and the load of sedimentary rock that was deposited in the downfold. Further periodic downfolding then resulted in the formation of magma in one part or another of the downfold, and the magma rose to form plutons and volcanic rocks.

In our model we have assumed three epochs of magma generation, but there were doubtless other epochs. For all three episodes, the rise of magma into the upper crust is shown in Fig. 7 to be accompanied by thickening of the basalt layer by residual concentration of refractory rocks and minerals and by the settling of crystals from the rising magmas.

In drawing the sections, shown in Fig. 7, that represent our interpretation of the different magmatic episodes, an important consideration was the amount of erosion that has taken place since each episode occurred, for this erosion must approximate the additional depth to which the crustal rocks and the Mohorovicic discontinuity were then depressed, provided there was no interchange of crustal and mantle material. This is an even stickier problem than that of determining the subsurface structure, because the wall and roof rocks are tightly folded and provide no means for determining their former distribution above the present ground surface, and because geophysical measurements cannot be made of material that is no longer there. At present the problem can be approached in two ways—through study of the volume of ma-

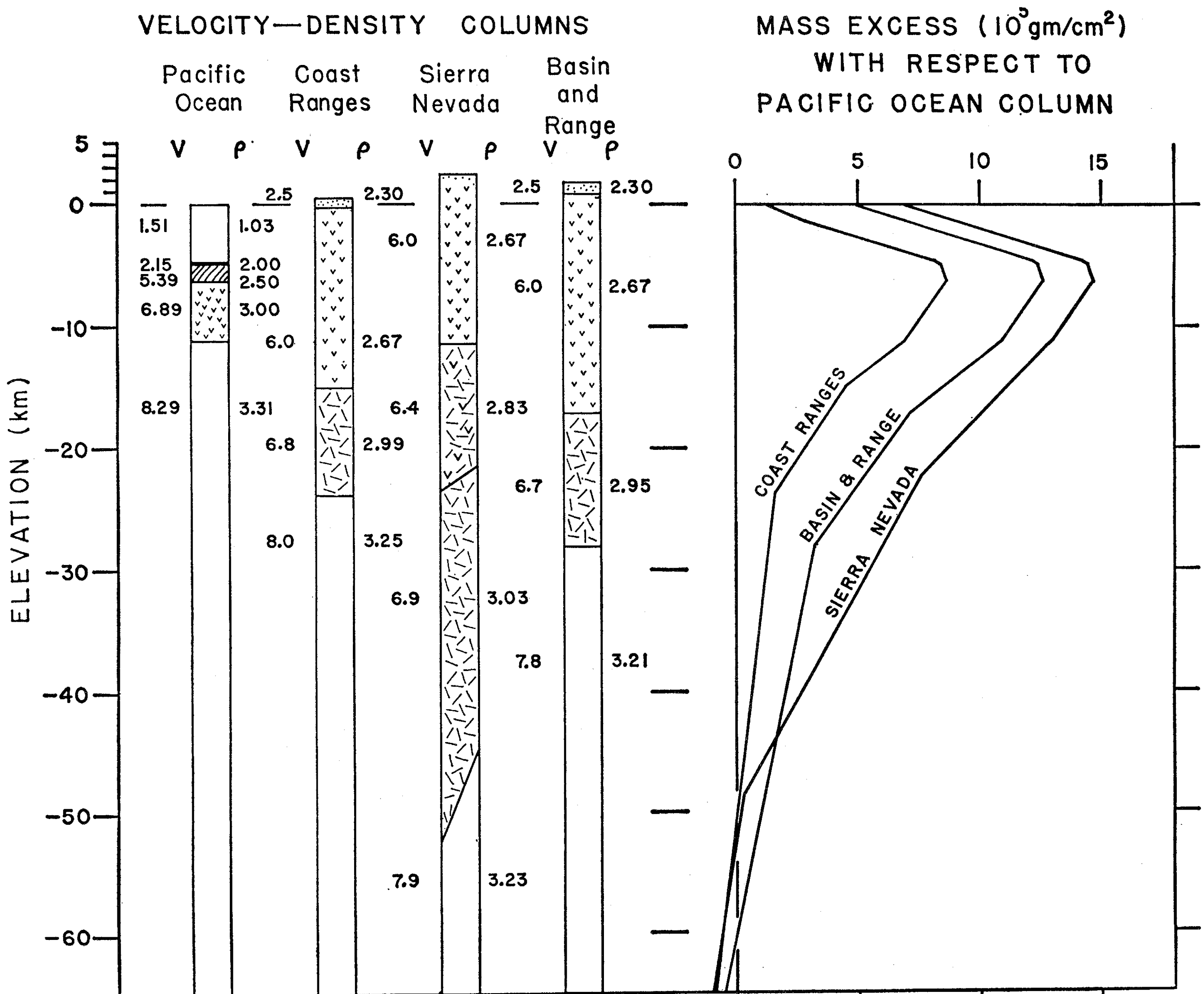

Fig. 6. (Left) Velocity-density profiles for the Pacific Ocean, Coast Ranges, Sierra Nevada, and Basin and Range province. *V*, P-wave velocity in kilometers per second; ρ, rock density in grams per cubic centimeter. Although errors in density values obtained by means of the curve of Fig. 5 are difficult to estimate and may be large, the densities are shown to the nearest 0.01 gram per cubic centimeter to indicate the values actually used in computing the mass-excess curves. (Right) Mass excess (in 10^5 grams per square centimeter of the earth's surface) as a function of depth, for the three continental columns with respect to the Pacific Ocean column.

terial deposited in areas adjacent to the Sierra Nevada, and through study of mineral assemblages that may act as load-pressure barometers. Bateman (*1, 2, 5*) has repeatedly explored this problem, with less than satisfactory results. Rough appraisal of the volume of material that has been eroded from the Sierra Nevada suggests that an amount of rock sufficient to form a crustal layer 8 to 16 kilometers thick has been removed since Late Jurassic time.

Several workers (*27*) have estimated the water-vapor pressure (P_{H_2O}) in granitic and contact metamorphic rocks at about 2 kilobats; because P_{H_2O} may be less than P_{total}, this figure probably represents the minimum possible load pressure at the present ground surface at the time the magma crystallized. The maximum possible load pressure is given by the upper pressure limit of stability of andalusite, the common polymorph of Al_2SiO_5 in the contact metamorphic rocks. Unfortunately, experimental investigations of the aluminosilicates published during the last 4 years indicate wide disagreement as to the upper pressure limit of stability of andalusite. Three recent investigators place it at 4.2 (*28*), 5.9 (*29*), and 6.6 (*30*) kilobars, respectively. Minimum and maximum pressures of 2.0 and 4.2 kilobars, equated to load pressure, correspond to depths of about 7 and 16 kilometers, an average density of 2.67 grams per cubic centimeter for the eroded rocks being assumed. Tectonic overpressures during metamorphism are unlikely because any overpressure should have been relieved by upward movement of the magma and plastic flow of the wall rocks. The cross sections that illustrate conditions at the various times of magma generation indicate the removal by erosion of 10 to 12 kilometers of rock since the Triassic or Early Jurassic and Late Jurassic magmatic episodes and of 7 to 10 kilometers of rock since the early Late Cretaceous episode. These estimated amounts of erosion would place the lowest part of the Mohorovicic discontinuity at a depth close to 60 kilometers during each magmatic episode.

Experimental studies show that, at atmospheric pressure, sialic rocks begin to melt fractionally at 960°C, but that an increase in water-vapor pressure causes the melting temperature to drop spectacularly (*31*). At pressures above 4 kilobars and in the presence of enough water to saturate the melt generated, melting begins at temperatures between 630° and 660°C, the melting temperature varying inversely with the pressure (*32*). If certain other substances, such as fluorine or chlorine, are present, the temperature at which melting begins is lowered further.

The first melt to form in a sinking synclinorium with rising temperature would be composed chiefly of quartz, orthoclase, and albite. Whatever water was available in pore spaces or in minerals would enter into this melt until the melt was saturated or the available water was exhausted. With increase of temperature the volume of melt would increase through the melting of progressively more refractory materials, including those that contained calcium, iron, and magnesium. At some volume of melt all the available water would be used up, and as the volume increased

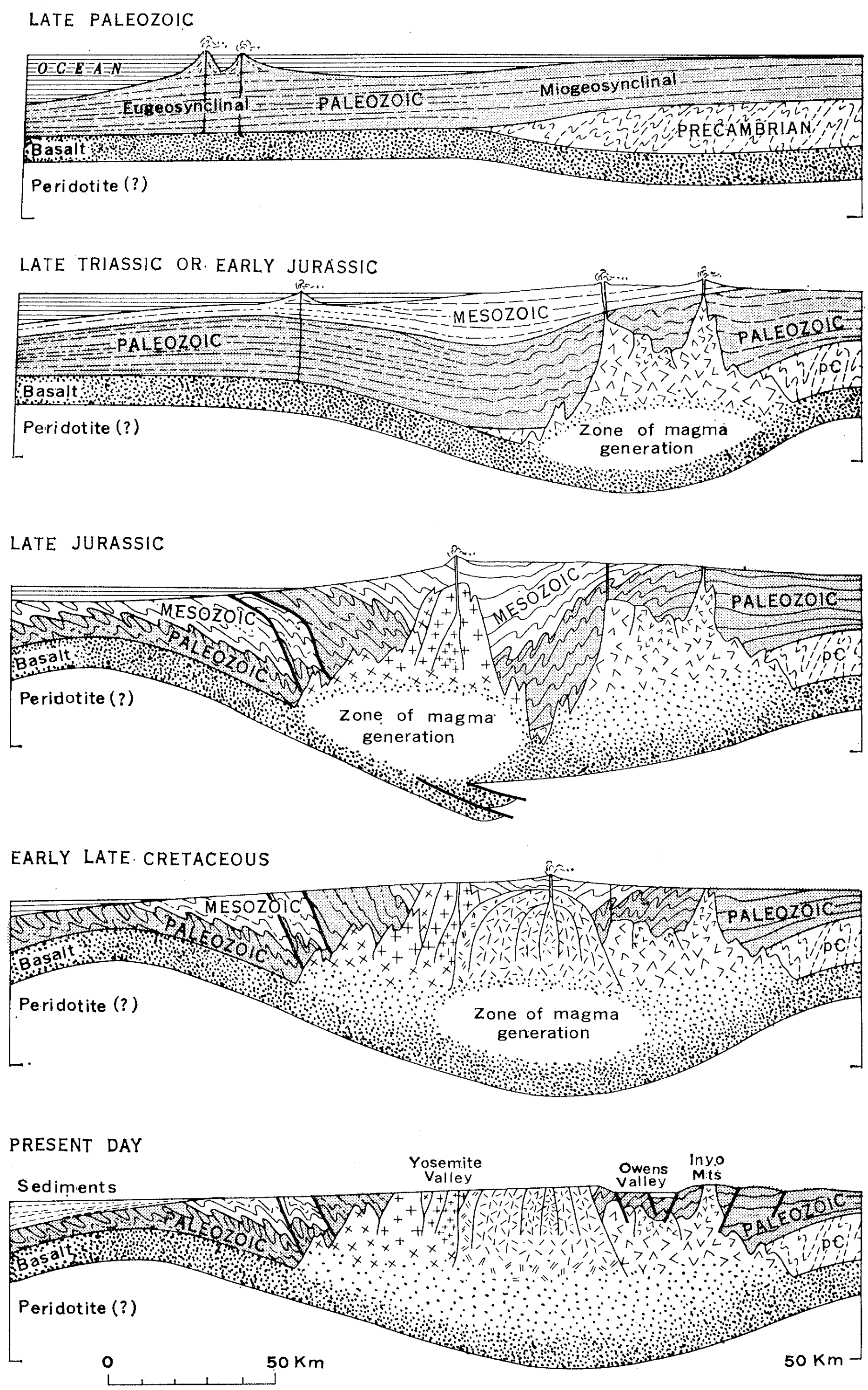

Fig. 7. Cross sections illustrating (bottom) the existing crustal structure of the Sierra Nevada and (top four sections) a model for the evolution of that structure.

beyond that point the melt would be increasingly unsaturated. Deficiency of water would adversely affect the formation of melt because higher temperatures are required to produce unsaturated melt than are required to produce saturated melt. That most Sierran magmas were unsaturated is indicated by a general lack of evidence of former volatile constituents either in the country rocks or in the granitic rocks, except for quartz veins and the contact-metasomatic rocks called skarns, which are abundant locally but not widespread.

Because the anorthite content of the plagioclase component in a granitic melt increases as the ratio of melt to crystals increases, this mineral provides an index to the temperature of the melt. Many plagioclase crystals in typical granodiorite have a homogeneous or slightly zoned core in the compositional range An_{55-45}, and a body that is progressively zoned (but with superimposed oscillations) outward from the core in the compositional range from An_{55-45} to An_{30-20}. The most likely explanation of this arrangement is that the nearly homogeneous core represents unmelted material that was brought more or less into equilibrium with the melt during its formation and rise into the upper crust, and that the zoned body crystallized around the core during cooling.

Plagioclase crystals of An_{50} composition can coexist with granitic melt containing about 15 percent of anorthite in the plagioclase component. At a temperature of about 870°C and a pressure of about 400 bars, a water content of 2 percent (by weight) is sufficient to saturate such a melt. As the total pressure is raised and the water content is held constant, the melting temperature is raised. The melting temperature of albite (a component of granite) increases at an average rate of 10°C per kilobar between 0 and 20 kilobars (*33*). If the melting temperature of granite rock with 2 percent water increases at half this rate, the melting temperature at a pressure of 14 kilobars (equivalent to a depth of about 50 kilometers) would be about 940°C. The melting interval thus spans the range from about 630°C, the temperature at which a small amount of water-saturated melt lacking an anorthite component first appears, to the melting temperature of the unsaturated melt—around 900°C for our example of melt containing water (2 percent, by weight), anorthite (15 percent) in the plagioclase component, and plagioclase crystals of composition An_{50}. Most Sierran magmas probably fell within this temperature range, but quartz diorite magmas may have been hotter.

It is difficult to say at what depth crustal rocks can be expected to melt to granitic magma because our knowledge of the generation and distribution of heat in the earth is still in a primitive state. In stable parts of the crust, where all the heat is carried to the surface by conduction, a temperature of 630°C may be reached at depths of 30 to 50 kilometers, but a temperature of 900°C would require much greater depth. However, the situation in a downfolding synclinorium is not ordinary, because crustal rocks relatively rich in radioactive uranium, thorium, and potassium are greatly thickened in the downfold.

Computations by Lachenbruch (*34*) indicate that the required temperatures can be produced by conduction and heat generation within strata accumulating in a synclinorium such as exists in the stratified rocks of the Sierra Nevada, reasonable amounts of uranium, thorium, and potassium and a modest heat flux from the underlying rocks being assumed. If the rate of heat production in the geosynclinal material were 1.67×10^{-13} cal cm^{-3} sec^{-1} and the flux from the surface of the pre-geosynclinal crust were 1.2 μcal cm^{-2} sec^{-1}, the temperature near the bottom of a 25- to 30-kilometer pile of geosynclinal volcanic and sedimentary rocks [conductivity = 5 mcal cm^{-1} sec^{-1} (deg C)$^{-1}$] would rise to 630°C during the time that must have elapsed between deposition and magma generation. The distribution of subjacent sources is immaterial to the foregoing calculation. With only 0.3 μcal cm^{-2} sec^{-1} coming from the upper mantle and the remainder of the pre-geosynclinal flux generated in the original crust, a temperature of 900°C would be attained at a depth of about 45 kilometers at the end of the same time interval.

The picture that emerges from these considerations is this: In a developing synclinorium a zone of partial melting would begin at a depth of 25 to 30 kilometers and extend downward to the lowest level of depressed sial. At the top of the zone of melting, the temperature would be about 630°C, and the ratio of hydrous felsic magma to solid rock would be very small. If no migration of material occurred within the zone of melting, the ratio of magma to rock would increase downward as the temperature increased, and the melt would be progressively less hydrous and richer in calcium, iron, and magnesium.

Because granitic magma is significantly less dense than granitic rock of the same composition, and certainly less dense than the residual refractory rocks, the magma would tend to move upward in the manner of a salt intrusion, exploiting lines of structural weakness wherever possible. Some magma would doubtless break through to the surface in volcanic eruptions, but much of it probably would crystallize at depth to form the plutons we now observe. Closely related sequences of granitic rocks that are successively more felsic with age and concentrically zoned plutons that are progressively more felsic toward the center indicate magmatic differentiation. Although lateral compositional variations are common, vertical variations have been observed in only a few plutons close to their roofs. No evidence has been found to indicate that the bottom of even one pluton lies close beneath the present ground level.

Most radioactive elements would be concentrated in the melt and would be carried upward with it. Lachenbruch and his associates (*35*) have pointed out that such an upward concentration of the radioactive constituents is, in fact, required by the high radioactivity of surface rocks and the relatively low heat flow from the Sierra plutons. Lachenbruch's calculations show that a reasonable upward concentration of the heat-producing elements (uranium, thorium, and potassium) during plutonism would lead to cooling on the order of 200°C in the zone of magma generation over a period of 50 million years. Postulating removal of the heat-generating surface rocks by erosion provides one means of reconciling the low heat flow that has been measured in the Sierra Nevada with the great crustal thickness and the observed radioactivity of surface rocks (*35*).

Summary

The Sierra Nevada batholith is localized in the axial region of a complex faulted synclinorium that coincides with a downfold in the Mohorovicic discontinuity and in P-wave velocity

boundaries within the crust. Observed P-wave velocities are compatible with downward increase in the proportion of diorite, quartz diorite, and calcic granodiorite relative to quartz monzonite and granite in the upper crust, with amphibolite or gabbro-basalt in the lower crust, and with periodotite in the upper mantle. The synclinorium was formed in Paleozoic and Mesozoic strata during early and middle Mesozoic time in a geosyncline marginal to the continent. Granitic magmas are believed to have formed in the lower half of the crust at depths of 25 to 45 kilometers or more, primarily as a result of high radiogenic heat production in the thickened prism of crustal rocks. Magma was generated at different times in different places as the locus of downfolding shifted. It rose into the upper crust because it was less dense than rock of the same composition or residual refractory rocks. Refractory rocks and crystals that were not melted and early crystallized mafic minerals that settled from the rising magma thickened the lower crust. Wall and roof rocks settled around, and perhaps through, the rising magma and provided space for its continued rise. Erosion followed each magmatic episode, and 10 to 12 kilometers of rock may have been eroded away since the Jurassic and 7 to 10 kilometers since the early Late Cretaceous.

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Hemoglobin and its Reaction with Ligands

The alpha-beta dimer appears to be the important unit in hemoglobin function.

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For almost a century hemoglobin has been the object of intense study by workers in the natural sciences. There is widespread interest in this protein not only because it has a prominent role in respiration, but also because it exemplifies many fundamental aspects of protein behavior.

Hemoglobin is contained in the red blood cells of all the vertebrates, and it accounts for more than 95 percent of the total proteins in these cells. Its physiological function is that of carrier of the respiratory gases (oxygen and carbon dioxide) between the tissues and the outer environment. The transport of oxygen depends on the specific property of the iron within the prosthetic group, which undergoes a reversible reaction with molecular oxygen; carbon dioxide, on the other hand, is bound, as such or in ionic form, by the amino acid side chains of the protein. In this article I describe a few aspects of the reaction of hemoglobin with ligands, with particular emphasis on the interrelation between the functional behavior and the structural properties of the protein.

Structure of Hemoglobin

Hemoglobin from red blood cells of mammals and most vertebrates, as far as is known, exists under many conditions as a tetrameric molecule of molecular weight 65,000; the molecule consists of four subunits, identical in pairs (*1–5*). The two different polypeptide chains have been called α- and β-chains so that the full molecule is indicated as $\alpha_2\beta_2$. Each polypeptide chain is bound to a prosthetic group, the heme, which is an iron II complex of protoporphyrin IX (*6*); in the native protein the ferrous iron binds reversibly molecular oxygen, carbon monoxide, nitric oxide, and other nongaseous ligands, such as alkylisocyanides and nitrosoaromatic compounds,

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