The Nature of Batholiths

By WARREN HAMILTON *and* W. BRADLEY MYERS

SHORTER CONTRIBUTIONS TO GENERAL GEOLOGY

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A survey of features of U.S. batholiths leads to the interpretation that these complexes are generally thin and that they crystallized beneath covers consisting largely of their own volcanic ejecta

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By Warren Hamilton and W. Bradley Myers

ABSTRACT

A survey of features of batholiths in the United States is interpreted to indicate that batholiths generally are thin, having spread out laterally at shallow depth, and that many of them reach the surface and crystallize beneath a cover of their own volcanic ejecta. It is inferred also that the magmas originate in the lower crust or upper mantle at depths greater than any ever exposed by erosion. Such conclusions agree with those reached by many geologists, but disagree with the concepts that batholiths are masses of great thicknesses, form beneath deep cover of metamorphic rocks, and crystallize from melts mobilized at the levels exposed in gneissic and migmatitic terranes.

Successively older Phanerozoic batholiths display in a broad way successively deeper sections into batholithic complexes. Mid-Tertiary batholiths are largely still capped by their volcanic crusts. The Late Cretaceous Boulder batholith of Montana is only a few kilometers thick; it spread across a floor of prebatholithic rocks, and preserves discontinuously its roof of almost exclusively volcanic rocks, which are contemporaneous and consanguineous with the plutonic ones. The Idaho batholith, largely of middle Cretaceous age, has a few small areas of possible volcanic roof rocks, although the correlative Sierra Nevada batholith has none remaining; but the more than a million cubic kilometers of volcanic ash in Cretaceous strata in the continental interior has no apparent source other than such batholiths. Seismic and gravity data indicate the Sierra Nevada batholith to be probably thin, and gravity data indicate the same for the Boulder batholith. The large Mesozoic batholiths were unroofed within a small fraction of a geologic period after their formation; since unroofing, they have been incised but not greatly eroded, and shallow depths are indicated. Late Cenozoic uplift correlates with Mesozoic batholithic rock type, apparently increasing with the proportion of radioactive components; the present crustal roots of the batholiths are responsible for the uplifts, are more mafic than the exposed rocks, and must somehow have formed because of the overlying batholiths. The uplift is now resulting in selective erosion of Mesozoic batholiths.

Batholiths are abundant in Precambrian and Paleozoic terranes, but probably none of these batholiths are as large as the great late Mesozoic batholiths of western North America. One possible interpretation of this contrast is that most pre-Mesozoic batholiths have been selectively eroded away. Gneiss terranes such as the sillimanite "plateau" of New Hampshire may form beneath batholiths, as plutons of magma rise bubblelike and displace heated wallrocks, which flow downward and beneath the plutons and become intensely metasomatized and injected. The metamorphic gradients flanking many gneiss terranes are far too steep to be explained in terms of geothermal heat conducted from the mantle, and the heat may have been introduced in magmas that largely rose through the crust and coalesced into surficial batholiths.

The largest Phanerozoic batholiths are partly in eugeosynclinal terranes, and the tectogene hypothesis of melting in downbuckled geosynclines is based on this association. Many Phanerozoic batholiths, however, intrude miogeosynclinal and platform sedimentary rocks and Precambrian basement rocks, and even have formed in the oceanic environment of island arcs. The tectogene hypothesis cannot be applied to such noneugeosynclinal batholiths—if batholiths have a common cause, it cannot be a tectogene. Strontium isotope data indicate that granitic magmas are melted from rocks poorer in rubidium than are exposed basement rocks and thus are derived from the lower crust or upper mantle or (in eugeosynclines) from volcanic materials derived in turn from such sources.

Batholithic and silicic-volcanic magmas become in general more silicic and more potassic as the continental crust becomes thicker, so the lower crust may be increasingly involved in melting as its depth increases. The magmas produce batholiths capped by volcanic fields in some places but produce high-alumina volcanic fields alone in others, depending upon local factors. Laboratory high-pressure data require that high-alumina batholithic and volcanic magmas be equilibrated with crystals above the depth at which basaltic rock undergoes pressure-phase transformation to eclogite; this can be achieved by partial melting, differentiation, or assimilation. Zone melting—whereby volatile components rising in response to pressure gradients within magmas lower the melting temperature of the roof while forcing crystallization low in the chamber—can cause great assimilation; indeed, very little of the final high-level magma need represent material present in the initial melt.

Much deformation conventionally ascribed to either crustal compression or gravity sliding may be due to the shouldering aside of wallrocks by rising batholith magmas. Batholiths, once formed, resist fragmentation by younger structures and hence greatly influence the subsequent deformation of their regions.

INTRODUCTION

Batholiths are composite masses of granitic rocks having areas ranging from tens of square miles to tens of thousands of square miles. Some batholiths that cut sharply across their wallrocks and that are surrounded by contact-metamorphic aureoles clearly formed from magmas intruded from greater depths. Other batholiths are largely concordant and lie within terranes of uniformly high-grade gneisses, and the origin of such batholiths and the source of heat for metamorphism of the associated gneisses are less obvious.

We attempt no broad review here of the descriptive features of batholiths; Buddington (1959) has done that ably, and the reader is referred to his work. We note features of batholiths and related rocks in the United States which appear to need explanation in any general theory of the origin and emplacement of batholiths. Much interpretation is incorporated with the individual descriptions in the first section of the paper. General synthesis and speculation follow in the second section. The rationale developed is that batholiths form from magmas generated in the upper mantle and lower crust, beneath any depths exposed by erosion; that pluton magmas rise in detached balloonlike forms through the crust, frequently reaching the earth's surface, and coalesce into shallow and fairly thin complexes; and that gneissic terranes form in the zones, beneath the final batholiths, through which the plutons rise, as wallrocks flow beneath the rising magmas and are heated and metasomatized by them. Elements of this synthesis have been suggested by other geologists, who are cited in the appropriate places. The third section of the paper discusses the influence of batholiths, once formed, upon the subsequent structural evolution of their regions.

Rock names applied on the basis of petrographic criteria to intermediate volcanic rocks generally connote compositions markedly more mafic and calcic than are connoted by names applied to plutonic rocks of identical chemical compositions. One is likely to think of andesite and diorite, dacite and quartz diorite, and rhyodacite and granodiorite as being of the same compositions, but this is not generally true. For example the postbatholithic andesites of Mount Rainier contain 60 to 64 percent SiO_2 and 1.6 to 1.9 percent K_2O (Fiske and others 1963, table 2); plutonic rocks of the same composition would be classed petrographically as granodiorite or at least quartz diorite. Chemical analyses rather than petrographic names should be employed when comparisons are made between intermediate plutonic and volcanic rocks.

REGIONAL DESCRIPTIONS AND INTERPRETATIONS

SIERRA NEVADA BATHOLITH

The Sierra Nevada batholith is known best across its central part (Bateman, 1965; Bateman and others, 1963; Calkins, 1930; Ernst Cloos, 1932, 1935a, 1935b; Durrell, 1940; Hamilton, 1956; Krauskopf, 1953; Macdonald, 1941; Moore, 1963; Rinehart and Ross, 1964; D. C. Ross, 1958; Sherlock and Hamilton, 1958; and others). Bateman, Clark, Huber, Moore, and Rinehart (1963, pl. 1) compiled a geologic map of the central Sierra. The batholith is 55 to 110 kilometers wide, has an exposed length of 650 km, and is a composite of plutons of Late Jurassic and Cretaceous ages (Kistler and others, 1965). Large plutons are elongate parallel to the northwesterly regional strike and small plutons lie between and within them. Younger plutons in many places cut sharply across structures of older plutons but in some places are nested concordantly inside them. Adjacent plutons can be of markedly different compositions within the spectrum quartz diorite-granodiorite-quartz monzonite-alaskite. The larger plutons and the bulk composition tend to become more silicic and richer in alkalis eastward; the dominant rock types are mafic calcic quartz diorite and granodiorite in the west and leucocratic granodiorite and quartz monzonite in the east, but exceptions are numerous on both sides.

Contacts between plutons and wallrocks and between adjacent plutons are typically so sharp that a hand specimen can be taken across them, although dike-injection zones make some contacts gradational at mapping scales and contact migmatites are extensive along some contacts with metamorphic rocks. Contacts may be quite irregular in detail but generally are broad curves at map scale. Thin discontinuous screens of metamorphic rocks locally separate plutons, and large pendants of metamorphic rocks separate groups of plutons.

The batholith is bounded by irregular belts of lightly to moderately metamorphosed Paleozoic and Mesozoic sedimentary and volcanic rocks. The proportion of mafic to silicic volcanic rocks is higher in the western border belt than in the eastern belt. The rocks of the western border belt in the central Sierra dip steeply toward the batholith and consist of long fault blocks of which those nearest the batholith contain the oldest rocks, although in each block the youngest rocks tend to be on the side toward the batholith; the eastern border belt of the central Sierra dips steeply in either direction but its rocks become in a gross way younger westward toward the batholith (Bateman and others, 1963). One pluton of the west part of the batholith was interpreted by Ernst Cloos (1932, 1935a) to have sent a flat tongue westward over its wallrocks. He (1935a, 1935b) emphasized that the flow-structure domes of some plutons demonstrated the batholith to have spread laterally, pushing its wallrocks aside, across the central Sierra.

Assimilation of mafic metamorphic rocks into mobile granitic magmas has been demonstrated in many places and probably has contributed much to the more mafic and calcic character of the western plutons, for mafic metavolcanic rocks are abundant in the western border belt, and widespread assimilation is shown along contacts. Static granitization has nowhere been found on more than a very small scale. Both regional and contact metamorphosed rocks show metamorphic grade and intensity decreasing systematically away from contacts: the primary source of heat for metamorphism to assemblages of higher temperature facies than greenschist was intruded granitic magma (Bateman and others, 1963; Durrell, 1940; Macdonald, 1941).

The flow structures of most plutons show them to have risen as units past their granitic and metamorphic wallrocks. Injection complexes of gently dipping dikes at some contacts demonstrate vertical stretching of wallrocks (Bateman, 1965, p. 116; Moore, 1963; Sherlock and Hamilton, 1958). Offset belts of metamorphic rocks indicate in some places shouldering aside by rising plutons (Bateman and others, 1963; Moore, 1963; Rinehart and Ross, 1964) but elsewhere wallrock belts are truncated irregularly by plutons. The presence of contact breccias along some contacts indicates that stoping was operative during late stages of intrusion, but the general absence of xenoliths away from contacts seems to be evidence against the process as the dominant mode of emplacement. Much detailed mapping apparently demonstrates that the intrusion of the plutons was dominantly forcible (Bateman, 1965, p. 115–123; Bateman and others, 1963, p. 44).

Evidence for a very shallow depth of crystallization of Cathedral Peak Quartz Monzonite—a large pluton of very coarse grained leucocratic rock—was summarized by Evernden (1965). The relation between variations in potassium-argon ages and in elevation led him to conclude that the pluton was emplaced no deeper than 7 km, and possibly as shallow as 4 km.

Seismic-refraction data indicate that the Mohorovicic discontinuity lies 40 or 45 km below sea level beneath the crestal region of the Sierra Nevada near the 39th parallel (Eaton, 1963) and about 50 km beneath the highest part of the crest farther south (L. C. Pakiser, written commun., 1964). Integration of gravity data with this seismic information requires that the greater part of the crustal thickening represented by the deepening of the Mohorovicic discontinuity is in crustal rocks that are markedly denser than the granitic rocks of the exposed batholith. The gravity data published by Thompson and Talwani (1964), for example, when considered with the seismic model, suggest that the Sierra Nevada batholith near the 39th parallel is a tabular structure with a thickness of at most 8 km.[1] The negative Bouguer gravity anomalies are not sufficiently large to permit both granitic rocks and the entire crust to be thick; as the crust is demonstrably thick, the granitic rocks cannot be.

From his preliminary interpretation of explosion seismic waves passing longitudinally through the eastern Sierra Nevada, Jerry P. Eaton (oral commun., 1965) concluded that velocities increase downward in about the manner indicated schematically in figure 1. We are much indebted to Eaton for permission to incorporate this information. Rocks with a velocity appropriate for silicic granitic rocks such as those dominating the surface exposures apparently extend no deeper than about 10 km. We interpret the velocity increase near this depth to occur beneath the thin batholith, and suggest that the underlying high-velocity (6.4 km per sec) rocks are metasomatized schist and gneiss displaced beneath the plutons of the batholith as they rose toward the surface. The depth-velocity fields of common types of igneous rocks are shown in the figure for comparison.

Other seismologists have drawn different interpretations from seismic data. Mikumo (1965) suggested that low-velocity granitic rocks need extend little deeper than sea level, where they give way downward to denser (Vp=6.3 km per sec) rocks. He achieved close agreement between measured Bouguer gravity and gravity calculated from a model in which the entire crust beneath the Sierra Nevada, from the surface to a Mohorovicic discontinuity reaching 46 km, has a density of 2.80 gm per cm^3. (Surface rocks lighter than this are, however, exposed throughout much of the region.)

Press and Biehler (1964), on the other hand, inferred that there is a velocity inversion within the upper crust and that rocks with a velocity appropriate to granite extend to great depth. They studied arrival times of *P* waves from nuclear explosions in the western Pacific, and found that arrivals at Tinemaha and Reno (both short distances east of the Sierra

[1] Thompson and Talwani assumed that the Mohorovicic discontinuity lies no deeper than 32 km beneath the Sierra crest. This assumption is contradicted by the seismic data; and even so, they could fit no more than about 12 km of granitic rocks to the gravity model.

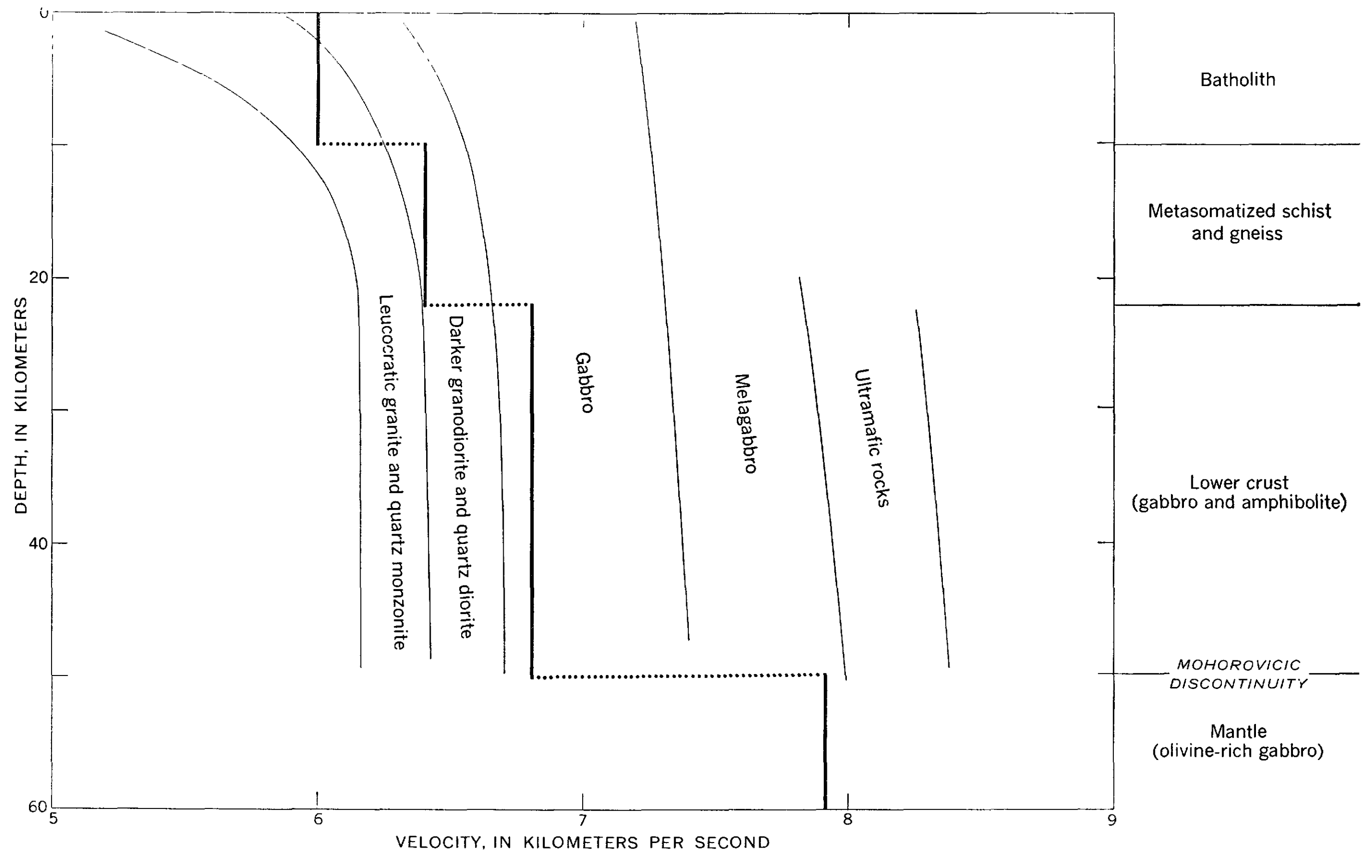

FIGURE 1.—Velocities of compressional waves (heavy lines) in a longitudinal section beneath the eastern Sierra Navada, from an unpublished interpretation by Jerry P. Eaton. Depth-velocity fields of representative types of igneous rocks are generalized from Birch's (1960) data. An interpretation of the velocities is shown at the right.

Nevada mountain block, but near the axis of the gravity minimum associated with the Sierra and its batholith) were about 0.8 sec later than they would have been if the structure of the crust and upper mantle was the same throughout California as it is at Pasadena. If this delay is due (as they assumed) to low-velocity granitic rocks beneath the Sierra Nevada, then the batholith has a deep root, and a thickness of granite and granodiorite of approximately 37 km is indicated (Press and Biehler, 1964, p. 2987–2988).

The interpretation by Press and Biehler is not however in accord with the local seismic-refraction data (such as that illustrated by fig. 1), and the P-wave delays they found can be interpreted alternatively in terms of variations within the upper mantle. To eliminate the possibility that variations in thickness of the lower crust could account for the arrival-time delays, Press and Biehler (1964, equation 10) analyzed the data for fit to these relationships:

$$\langle \Delta g \rangle = 42 \Delta h(\rho - \rho') \approx -12(PD)\alpha\alpha'$$

in which $\langle \Delta g \rangle$ is the difference in slab-equivalent Bouguer gravity, in milligals, between the reference station (Pasadena) and the seismograph station; Δh is the difference, in kilometers, in depth of the Mohorovicic discontinuity; $(\rho - \rho')$ is the difference in density, in grams per cubic centimeter, above and below the discontinuity; (PD) is the wave delay, in seconds; and α and α' are the velocities of P waves above and below the discontinuity. Press and Biehler showed that, given reasonable velocities, the expression on the right yields values of $\langle \Delta g \rangle$ approximately 50 percent too high. The middle expression, however, provides calculated values agreeing with the observed ones: the reasonable figures of $\Delta h = 22$ km and $(\rho - \rho') = 0.3$ gm per cm^3, for example, yield $\langle \Delta g \rangle = 280$ mgal, the same as that observed. The conflict can be resolved by the interpretation that the wave delay is due to lower velocities in the upper mantle beneath the Sierra Nevada than beneath coastal California, and that the batholith is thin, rather than thick as Press and Biehler assumed.

Heat-flow data also indicate the Sierra Nevada batholith to be thin. Thus, granodiorites west of the crest of the central Sierra produce by radioactive decay about 10 microcal per gram per year, yet heat flow in a deep core hole is only 1.3 microcal per sq cm per sec; if the entire flux came from granodiorite like that in the hole, the granodiorite could be only 15 km thick (Lachenbruch and others, 1966; see also Wollenberg and Smith, 1964). The more mafic granitic rocks of the western part of the batholith produce about 2½ microcal per gram per year, and heat flow in a deep core hole there is only 1.3 microcal per sq cm per sec, equivalent to the heat production in 30 km of the local rock (Lachenbruch and others, 1966). As much of the heat must in fact come from greater depths, batholithic rocks like those near the surface must be much thinner than these limits. Five heat-flow determinations farther north in the Sierra batholith average only 0.9 microcal per sq cm per sec, half the value in the Great Basin to the east (Roy and Blackwell, 1966), leading to the same conclusion that the batholith is thin.

We conclude that the crustal root beneath the high part of the Sierra Nevada is largely of rocks markedly more mafic and heavier than the exposed quartz monzonite and leucogranodiorite, and that the batholith is limited to the upper part of the crust.

Figure 2 shows a section through the central Sierra Nevada. The interpretation incorporates seismic and gravity data, and illustrates conclusions developed in this paper.

IDAHO BATHOLITH*

The Idaho batholith of central Idaho has an exposed length of 400 km and a width of 130 km and is surrounded by regionally metamorphosed rocks (Hamilton, 1963a, b; Larsen and Schmidt, 1958; Reid, 1959; Ross, 1963; Schmidt, 1964; and others). Massive granodiorite and quartz monzonite underlie two main regions, one in the southwest part of the granitic terrane and the other in the northeast; elsewhere there is much schist and gneiss interspersed with the granitic rocks of the batholith. Quartz diorite and trondhjemite are widespread in the gneissic western border zone of the batholith. The culminating intrusions of the Idaho batholith occurred about the middle of Cretaceous time according to lead-alpha determinations on zircons (Jaffe and others, 1959).

It is possible that the Idaho batholith formed without a roof, its plutons having reached the surface and erupted a volcanic capping, beneath which magma crystallized more slowly. The Casto Volcanics in the east-central part of the region of the batholith can be interpreted speculatively to be remnants of this volcanic cap. The Casto is undated, variably altered, and contact metamorphosed but generally nonschistose intermediate lavas and pyroclastics intruded by the Idaho batholith (Leonard, 1962; Ross, 1934).

*See note on page C30.

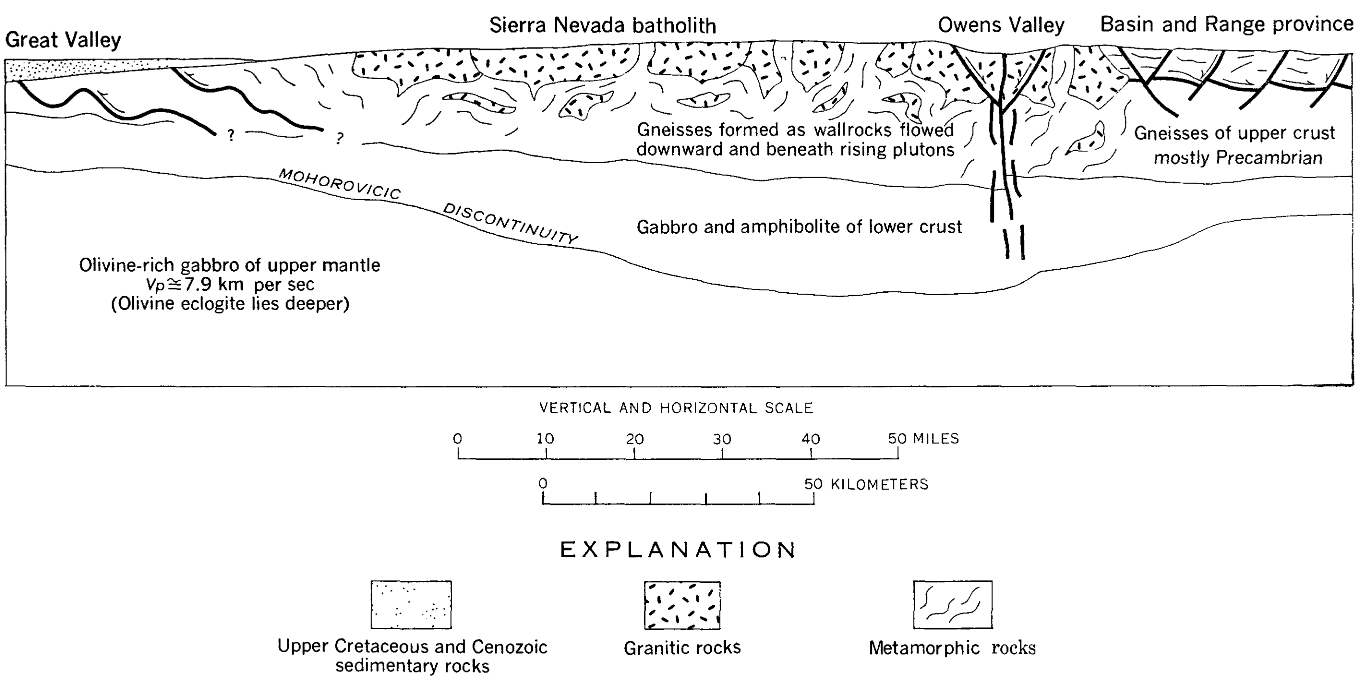

FIGURE 2.—Geologic and crustal section through the Sierra Nevada of California, along the 37th parallel. Adapted from Hamilton and Pakiser (1965). Plutons of granitic magma, melted in upper mantle and lower crust, rose through crust and coalesced at surface to form Sierra Nevada batholith. In the Basin and Range Province, Paleozoic sedimentary rocks moved along bedding-plane thrust faults, then broke into normal-fault blocks.

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Deformation of the volcanics is only moderate. Typical dips are about 25°, and the rocks apparently lie unconformably upon highly deformed metasedimentary and metavolcanic rocks of which some at least are of late Precambrian age. Ross and Leonard both assumed the Casto to be part of the Paleozoic and Mesozoic eugeosynclinal suite, much older than the batholith, and perhaps of Permian age; if this assumption is correct, so is their conclusion that there has been no major deformation or regional metamorphism since pre-Permian time. In the western border zone of the batholith, however, the fossiliferous Upper Triassic Martin Bridge Limestone was extremely deformed, highly metamorphosed, and intruded by granitic rocks which in turn were intruded by the main plutons of the Idaho batholith (Hamilton, 1963a, b), and the geometry of the deformation that accompanied the regional metamorphism indicates a genetic relation to the batholith. The absence of comparable deformation in the Casto Volcanics might indicate that they postdate the emplacement of the batholith—and yet the volcanics are intruded by the batholith. A possible explanation is that the Casto Volcanics formed by extrusion of lava from the Idaho batholith itself; if so, at least part of the batholith formed with no roof other than a crust of its own ejecta.

Leonard (1963) and Leonard and Stern (1966) have, however, presented further evidence which they regarded as indicating that major deformation in central Idaho long predated the Idaho batholith. Highly deformed upper Precambrian metamorphic rocks were intruded there by syenite whose lead-uranium and lead-thorium calculated age is about 600 million years. (One potassium-argon hornblende age of only 93 m.y. apparently represents metamorphism by the Cretaceous batholith.) Leonard considered the syenite to postdate the major folding of the enclosing rocks, and our conjectures in the preceding paragraph are wrong if he is correct. The syenite, however, is grossly concordant to the structures of its Precambrian wallrocks, and has been variably crushed and recrystallized; so it is possible alternatively that the syenite was intruded into the old rocks before, rather than after, their metamorphism and deformation and thus that the syenite provides a maximum rather than minimum age for that event.

BOULDER BATHOLITH

The Late Cretaceous Boulder batholith of southwestern Montana is a composite mass 100 km long and 50 km wide, and consists of plutons of granodiorite, quartz monzonite, and other granitic rocks (Becraft and others, 1963; Klepper, 1950; Klepper and others, 1957; Knopf, 1963, 1964; Ruppel, 1963; Smedes, 1962). The batholith lies far to the east of the general belt of late Mesozoic metamorphism and batholith formation. The batholith appears to be a floored sheet, roofed only by its own volcanic ejecta.

The roof rocks of the batholith are exposed in many areas and consist of dacite, rhyodacite, and quartz latite. (Considerable "andesite" has been reported in the literature also, but published analyses of such rock are clearly of dacite and rhyodacite.) The granitic and volcanic rocks are of the same age within the Late Cretaceous insofar as the radiometric and isotopic age determinations of the granitic rocks can be compared with the paleontological dates of the volcanics. The volcanic rocks have a maximum thickness of 2 or 3 km and are mostly pyroclastic. They are broken only by warps and normal faults in most areas, and are but slightly altered except where converted to hornfels near the contact with the batholith.

Younger plutons within the batholith are in general more silicic and richer in alkalis than are older ones. A similar irregular compositional age progression within the volcanic roof rocks, silica and alkalis increasing upward, has been reported on the basis of field and petrographic studies, although it is not apparent in the few published chemical analyses of the rocks involved. Granitic and volcanic rocks broadly overlap in composition but the granitics are in bulk composition more silicic and more potassic than are the volcanics.

The present erosion surface everywhere is probably within 2 km above or below the position of the original roof of the batholith, for the subhorizontal contact with volcanic rocks is exposed in most regions about the batholith. Presumably the original relief of the roof was still less, as Cenozoic deformation has much affected western Montana, and the initial top of the batholith may have been almost horizontal. Roof-rock lavas lap across the edge of the batholith onto the older wallrocks. A few small masses of contact-metamorphosed sedimentary rocks, probably of prevolcanic Mesozoic units, are present locally in the roof complex and perhaps represent rafts of floor rocks. Lower Eocene quartz latite lies upon eroded granitic rocks (Smedes and Thomas, 1965), and shows that the batholith was exposed by erosion very soon after its formation and that magmatism continued into Tertiary time.

The volcanic rocks also form the wallrocks for most of the east margin of the batholith. The contact is steep and irregular. Deformation of the wallrock volcanics has been more severe than that of the roof-rock ones, and the wallrock volcanics have gentle to moderate dips and are broken by many faults.

The north contact of the batholith is semiconcordant to Paleozoic and prevolcanic Mesozoic strata which dip southward beneath the granitic mass (fig. 3; Knopf, 1963; Smedes, 1962). Dips in the wallrocks tend to steepen toward the contact with the batholith. The contact forms in plan three large cusps, concave toward the batholith; the western and central cusps appear in figure 3. The cusps are synclinal sags and are separated by sharp anticlines. The western cusp is 25 km across, and the batholith is in contact with middle and upper Paleozoic beds. The eastern two cusps are 20 and 10 km wide, and the granitic rocks lie against Cretaceous and Jurassic strata. Relatively dense (specific gravity about 2.8) granodiorite lies along the central and eastern cusps and forms an outcrop belt 0 to 4 km wide, south of which is less dense (about 2.71) quartz monzonite. A subhorizontal sheet of still lighter granophyre lies discordantly above the heavy granodiorite, in the same structural position occupied by nearby remnants of volcanic roof rocks (fig. 3). We interpret the general parallelism of the granodiorite-quartz monzonite contact to the margin of the batholith as suggesting that the granodiorite is part of a sheet which dips southward beneath the quartz monzonite.

The south contact of the batholith is against Precambrian crystalline rocks and Paleozoic and Mesozoic strata. Available data indicate the contact to be complex but are too meager to permit satisfactory generalization beyond the observation that long segments of the contact are semiconcordant to wallrock formations whose tops lie toward the north. The west contact of the batholith is hidden beneath Cenozoic deposits but probably is largely or entirely against Paleozoic and prevolcanic Mesozoic strata.

When it is viewed on a broad scale such as that of the geologic map of Montana (Ross and others, 1955), the Boulder batholith is seen to occupy a structural depression. The batholith is surrounded mostly by Mesozoic and Paleozoic strata, whereas Precambrian rocks are extensively exposed in other parts of the same tectonic province elsewhere in Montana. The batholith fills a basin.

The roof of the west-central part of the batholith is against an almost constant stratigraphic level in the overlying volcanic rocks through a broad area (Ruppel, 1963, p. 37). Contacts between variant granitic rocks in the batholith are partly sharp and partly gradational but are in general subparallel to the roof contact, and successively lower units tend to be succcessively coarser grained; fine-grained quartz monzonite several hundred feet thick typically lies between the volcanic roof rocks and the coarser quartz monzonite of the interior of the batholith (Ruppel, 1963, p. 32, 37).

Lawson (1914) long ago suggested that the Boulder batholith was a floored sheet, intruded between the Cretaceous volcanic rocks and the older rocks beneath. Barrell (1907, p. 166) made similar suggestions still earlier but thought them improbable. Ruppel (1963) suggested that the west-central part of the batholith was a floored sheet because the subhorizontal contacts between plutons are strong evidence for horizontal flow of the intrusive magma. The concordance of the northern contact to the inward-dipping right-side-up Mesozoic and Paleozoic section indicates that sector also to be floored. The near lack of prevolcanic rocks along the east margin of the batholith suggests that there, too, the granitic rocks lie wholly above the prevolcanic section. Prevolcanic rocks are virtually lacking in the roof.

The Boulder batholith is capped almost exclusively by its own volcanic ejecta and is better regarded as an extrusive complex, of which the volcanic rocks form the upper part and the granitic rocks the lower, than as an intrusive complex. The batholith magma flowed, in effect a gigantic mantled lava flow, across a broad basin whose subsidence may have been due to the withdrawal of magma from depth. Presumably a crosscutting batholith (in the customarily limited sense of the term) within the basin, or several stocks and small batholiths, served as magma conduits. The magma formed volcanic rocks where it erupted to the surface and granitic rocks where it crystallized beneath an insulating crust of its own ejecta. The capping crust was thickened by eruption over the top, by injection into it of dikes and sills (which are abundant in the volcanic roof), and by crystallization along the bottom of such rocks as granophyre and fine-grained quartz monzonite.

Granitic rocks intrude the volcanic rocks wherever the two are in contact, but this does not require the granitic rocks to be wholly younger than the volcanic ones: it indicates that granitic rocks formed where magma crystallized beneath insulating cover. The generally offset compositional ranges of volcanic and granitic rocks—the more silicic half of the volcanic rocks having the same general composition as the less silicic half of the granitic rocks—may indicate that the overlap in age consisted largely of the younger half of the volcanic series and the older half of the granitic series. If this is correct, then thick volcanic rocks formed across the entire basin before much magma spread laterally between volcanics and floor; however, the hidden deep part of the batholith may well consist of relatively mafic rocks of the same com-

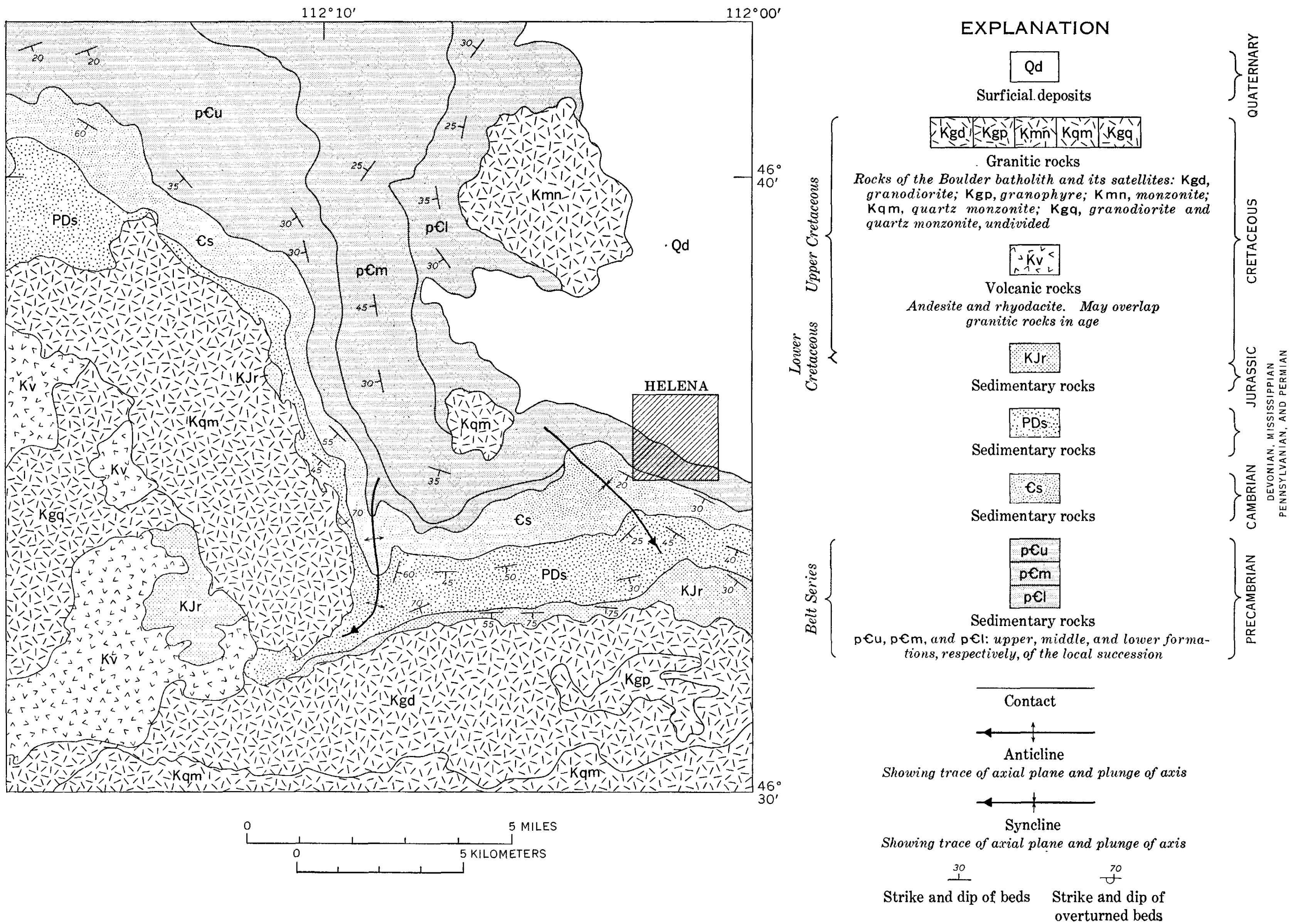

FIGURE 3.—Geologic map of part of the north end of the Boulder batholith, Montana. Generalized from Knopf (1963); nomenclature of granitic rock types is that of this paper rather than of Knopf.

position and age as the more mafic volcanics. A reasonable picture is that intrusion and extrusion largely overlapped in age and that the volcanic crust was thickened and broadened as granitic rocks crystallized beneath it. Repeated injections of magma beneath the volcanic cover are indicated by the many contacts between plutons. In the west-central part of the batholith, successive plutons flowed laterally.

Successive episodes of subsidence are suggested by the geometry along the north margin of the batholith. The granodiorite south of Helena (fig. 3), for example, may have crystallized after its magma filled the cuspate embayment in the wallrocks, and then both granodiorite and walls may have subsided more. The deepened cusp was then filled by the magma which formed the quartz monzonite to the south and above the granodiorite, if, as we infer, dips between the plutonic units decrease into the batholith. The aggregate thickness of the units is probably markedly less than might be inferred from extrapolation of dips of the north contact of the batholith.

A crust of volcanic rocks perhaps 2 km thick floated upon granitic magma over a region of about 7,000 square kilometers. Barrell (1907, p. 166) implied this long ago. Whether or not all the roof floated at any one time is not yet certain. The warping and normal faulting of the volcanic pile may have occurred largely as a result of its vertical and horizontal jostling and stretching while afloat.

Many of the points discussed here and in subsequent sections are illustrated by the diagrammatic section through the Boulder batholith (fig. 4).

A gravity survey across the north end of the Boulder batholith shows no abrupt decrease in gravity corresponding to the margin of the granitic mass, which thus cannot there have great thickness (Renick, 1965). The data indicate that the surveyed part of the batholith is very thin at its north edge and thickens only gradually southward to a thickness of about 5 km, 15 km south of the north margin. (Renick's model shows the batholith thickening to 6 or 7 km, but 5 km provides a better fit with the observed gravity.) An unpublished gravity survey of a larger area of the batholith and surrounding region is said to indicate the batholith to be thinner than 15 km (Biehler, 1966).

The Boulder batholith lies in a miogeosyncline—not in a eugeosyncline. Most Mesozoic batholiths of western North America occur at least partly within eugeosynclines, but such an environment is obviously not necessary for their formation; and an explanation for the origin of batholiths cannot properly apply to a eugeosynclinal setting alone. Such matters are discussed in the subsequent section on "The environment of batholiths."

VOLCANIC ASH

Mesozoic strata of the western interior United States contain 1 or 2 million cubic kilometers of clay altered from volcanic ash blown in from sources nearer the Pacific Ocean. The varicolored siltstones and claystones of the Upper Triassic Chinle Formation of the southern part of the Colorado Plateaus consist largely of montmorillonitic clay derived from latitic or quartz latitic ash and of mixed-layer clay derived from rhyolitic ash, and relic shards and volcanic minerals are common (Schultz, 1963). Clay in the Middle and Upper Jurassic Carmel Formation and Upper Jurassic Morrison Formation of the Colorado Plateaus also is largely of volcanic-ash origin, and relic volcanic textures are widespread (Keller, 1962; Schultz and Wright, 1963). Far more voluminous still are the volcanic-ash clays (montmorillonite and mixed-layer clay) of the marine Upper Cretaceous shales of the Colorado Plateaus and Great Plains (Schultz, 1965; Tourtelot and others, 1960; Leonard G. Schultz, oral commun., 1965). The volume of Upper Cretaceous strata of the western interior is approximately 4 million cu km (Gilluly, 1963); of this total, perhaps one-fourth or even one-half is altered volcanic ash, to judge by detailed studies of the Pierre Shale and reconnaissance studies of other formations (Schultz, oral commun., 1965).

Volcanoes were active throughout much of the Late Triassic and the Jurassic near the Pacific margin of North America, whereas known Cretaceous volcanoes were less widespread and were farther inland (Gilluly, 1965). Granitic rocks of Triassic age are uncommon, those of Late Jurassic and Early Cretaceous age are widespread, and those of Late Cretaceous age are abundant and form large parts of the great Sierra Nevada and Idaho batholiths (Gilluly, 1963, 1965; but see Kistler and others, 1965). Unroofed batholiths may have provided much of the pyroclastic material preserved in the strata of the western interior, and the Late Cretaceous batholiths may have been the source of the greatest part of that material. The numerous middle Cretaceous bentonites of Wyoming came primarily from dacitic and quartz latitic sources farther west, presumably volcanoes above the Idaho batholith (Slaughter and Earley, 1965).

TERTIARY PLUTONS OF CASCADE RANGE

The Cascade Range of Washington is among the regions where middle or late Tertiary plutons are known to have broken through to the surface and produced volcanic piles. Fiske, Hopson, and Waters

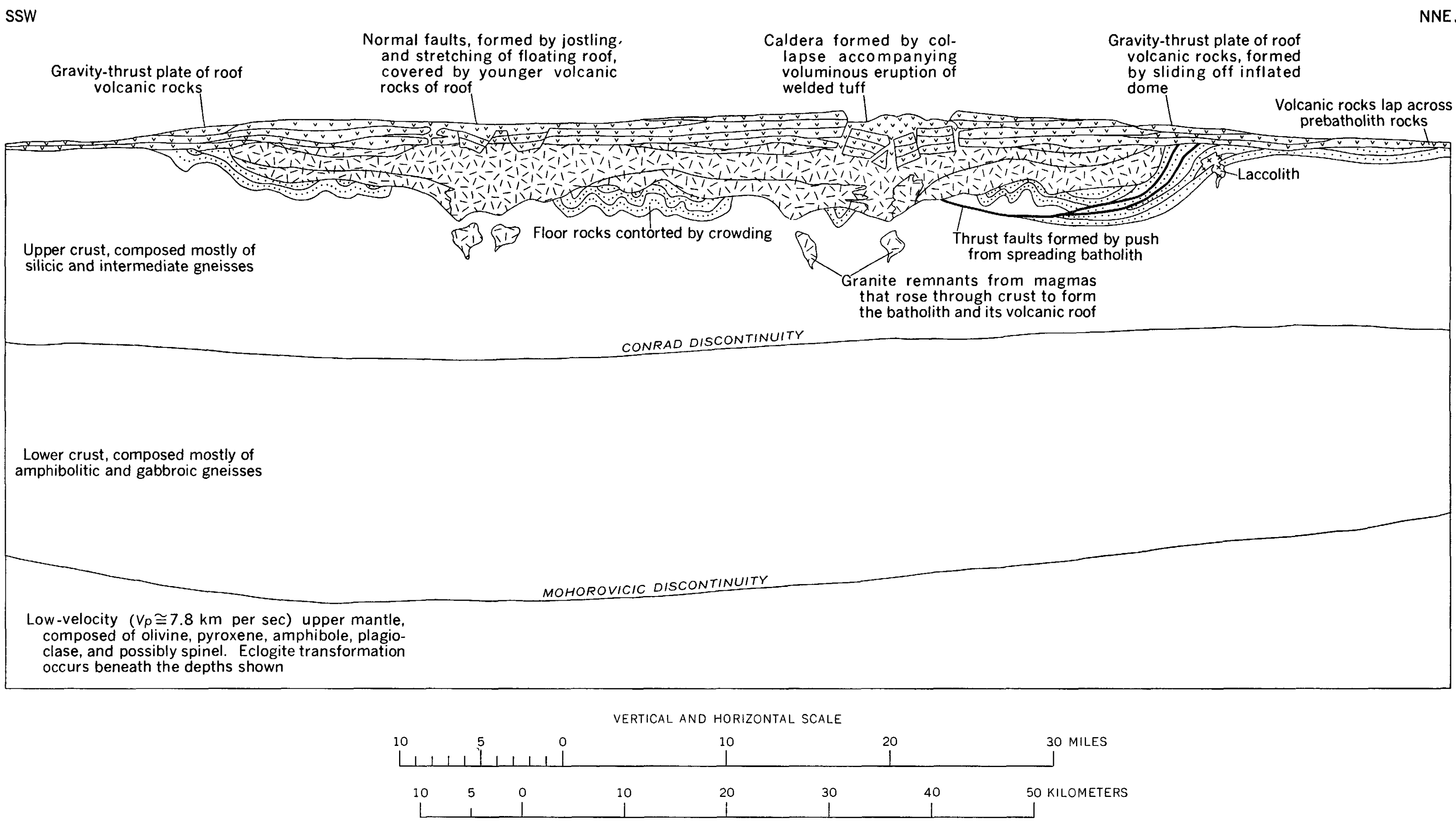

FIGURE 4.—Diagrammatic longitudinal geologic and crustal section through the Boulder batholith, Montana. The batholith and its roof are of Late Cretaceous age, and the stratified rocks range in age from late Precambrian to early Late Cretaceous.

(1963, p. 40–63) showed that the 250-sq-km Tatoosh pluton of granodiorite and allied rocks in the Mount Rainier area produced explosive eruptions through its roof. Rapid loss of heat and volatiles resulted in distinctive crystallization features—vertically alined vesicles, miarolitic vugs, explosively brecciated rocks, local granophyre, aphanite, and vitrophyre, and transitions between pluton and volcanic plugs—in the upper few hundred feet of the pluton in vent areas. Rapid crystallization of the entire pluton, despite its granitic texture, is indicated by the absence of flow banding and lineation: "there was no prolonged interval during which the material moved as a viscous crystal mush" (Fiske and others, 1963, p. 46).

The much larger Snoqualmie Granodiorite batholith, farther north in the Washington Cascades, also deroofed itself explosively (R. E. Fuller, unpub. 1925 thesis, as cited by Fiske and others, 1963, p. 59). Pyroxene diorite in cupolas of the batholith grades upward through volcanic plugs into andesite flows and pyroclastics. Here also, resulting dehydration produced chill effects high in the batholith. Miocene plutons of quartz diorite and granodiorite crop out near five of the northern volcanoes of andesite and dacite in the Cascade Range, and their presence suggests that the volcanic magmas came from the plutons, which remained active at depth (Hopson and others, 1966).

The Tertiary volcanic and intrusive rocks of the Cascades in part were erupted through pre-Cenozoic crystalline complexes but in part have probably formed where only oceanic crust existed in Mesozoic time. In northern Washington and in southern Oregon and northern California the Cascade igneous rocks cut and overlie the granitic and metamorphic rocks of the late Mesozoic orogens. These older rocks strike southeastward in the north and northeastward in the south so that northwestern Oregon and southwestern Washington are on the Pacific side of all exposed and projected pre-Cenozoic rocks (Carey, 1958, fig. 56; King, 1959, p. 161). We infer that there is no pre-Cenozoic continental crust within this tectonic embayment. The central Cascades (including Mount Rainier) and the Coast Ranges of most of Oregon and Washington have apparently been built upon oceanic crust as a Cenozoic addition to North America.

ALEUTIAN ISLANDS

The island arc of the Aleutians consists of young volcanoes built upon platforms of middle Tertiary and older submarine lavas and pyroclastics. Andesite dominates both suites, but basalt and dacite are common. Many of the old rocks are variably spilitized, but their prealteration compositions were identical to the young rocks. (The geology of the islands has been described by various authors, under the general title "Investigations of Alaskan Volcanoes," in the many chapters of U.S. Geological Survey Bulletin 1028.)

The older complexes are intruded by Tertiary granitic rocks on a number of islands. These are limited to stocks and smaller masses except on Unalaska Island, where three small batholiths occur (Drewes and others, 1961, p. 610–634 and table 1). About 70 percent of the batholithic rocks consists of granodiorite. The remainder ranges from quartz gabbro to light-colored quartz monzonite. The granodiorite is chemically the same as andesites and dacites of both older and younger volcanic sequences and presumably is but an intrusive manifestation of the same igneous activity.

The Aleutian Islands extend across the North Pacific. No evidence requires that continental crust existed there before the onset of andesitic island-arc volcanism. The mechanism of generation of the magmas from an oceanic mantle is discussed in the second major section of this paper.

TERTIARY IGNEOUS ROCKS OF COLORADO

Numerous stocks of lower Tertiary granitic rocks dominated by quartz monzonite and granodiorite form a chain trending northeastward across western Colorado. The rocks intrude Precambrian plutonic rocks and the thin overlying Paleozoic and Mesozoic platform sedimentary rocks.

The great middle (and late?) Tertiary volcanic pile of the San Juan Mountains of southwestern Colorado is formed largely of rocks equivalent in composition to quartz diorite and granodiorite. Large calderas and other magmatic collapse structures occur throughout a region of at least 7,000 sq km (Luedke and Burbank, 1963; Steven and Ratté, 1963; Thomas A. Steven, oral commun., 1965); Tertiary batholiths must lie hidden beneath the volcanic cover of this area. No data are available to suggest whether the granitic rocks are mostly in a floored complex above the pre-Tertiary rocks.

Western Colorado was a stable platform region during late Precambrian(?), most of Paleozoic, and early Mesozoic times, but during Cretaceous and Cenozoic times it has been the site of moderate deformation. Western Colorado now has the highest regional elevation in the conterminous United States, and has a thick crust, high heat flow, and low-velocity upper mantle. The heat flow and mantle velocity distinguish the region from the still-stable platform east of the Rocky Mountains, and presumably the abnormal heat flow and mantle velocity, and the crustal thickening, are products of Cretaceous and younger changes and are related genetically to the deformation and igneous activity of western Colorado. A geosynclinal environment is not essential for the generation of granitic magmas: the controlling events occur in the lower crust or upper mantle, not in the upper crust.

TERTIARY IGNEOUS ROCKS OF BASIN AND RANGE PROVINCE

Silicic volcanic rocks (mostly welded tuffs) and intrusive porphyries and granitic stocks of similar compositions formed throughout broad parts of the Basin and Range province of the Western States during middle Tertiary time. Presumably the intrusives fed the extrusives. Blank (1963) found a porphyry in southwestern Utah that broke through its roof after partial crystallization and produced quartz latite welded tuffs. Gilluly (1932, p. 69) found that in the Oquirrh Range of central Utah, silicic stocks and porphyries cut the volcanic pile that probably was erupted from the same magma chambers, and similar relationships have been reported elsewhere; but in general, wallrocks of stocks consist of pre-Tertiary rocks.

The Tertiary silicic volcanic rocks of Nevada and western Utah have a total volume of approximately 120,000 cu km (Mackin, 1960, p. 83). This is equivalent to about half the volume of the Sierra Nevada batholith if we are correct in assigning a thickness of approximately 8 km to the batholith. Most of this Great Basin material was melted beneath a miogeosynclinal terrane, and part of it was melted beneath an unstable platform environment.

ST. FRANCOIS MOUNTAINS BATHOLITH

The St. Francois Mountains of southeastern Missouri expose a Precambrian complex in which a roof of extrusive rhyolite and quartz latite was intruded by leucocratic granite, quartz monzonite, and granophyre (Bridge, 1930, p. 59–64; Dake, 1930, p. 26–44; Haworth, 1895; Hayes, 1961; Robertson, 1966; Snyder and Wagner, 1961). Granophyric and granitic rocks are about 1,300 million years old (Allen and others, 1959; Tilton and others, 1962). According to Anderson (1962; oral commun., 1965), the roof rocks above the plutons of granophyre and granite are chiefly welded tuff and tuffaceous sedimentary rocks and are about 2 km thick. Despite their age and their intrusion by granite, the volcanic rocks have generally low dips, and are broken only by minor normal faults which developed contemporaneously with volcanism. The volcanic rocks have not been metamorphosed regionally, although some have been weakly silicified and albitized (propylitized?), and they contain well-preserved primary volcanic fabrics such as collapsed pumice lapilli and shards. Granophyre tends to lie with gently dipping contacts between volcanic rocks above and coarse granitic rocks beneath, and therefore apparently represents a rapidly crystallized roof facies of the batholith. In bulk composition and compositional variation, granitic and granophyric rocks are similar to the dominant volcanic rocks. Rocks of all three types are mostly red, and are moderately alkalic and low in alumina. Minor sodic dacites present low in the volcanic succession, however, are less silicic than are any of the exposed silicic intrusive rocks (Anderson, 1962). Sheets and irregular intrusive masses of basalt and diabase cut both intrusive and extrusive silicic rocks.

We infer the volcanic rocks to be consanguineous with the intrusive granitic and granophyric rocks. The petrological similarities are great, and the volcanic rocks lack any evidence of regional deformation and metamorphism to suggest that they represent a wallrock terrane intruded by the batholith. If our inference is correct, then the batholith is roofed only by its own ejecta. The volcanic crust solidified from magma erupted from the molten interior of the complex. Granophyre crystallized first beneath the volcanic crust, and its formation thickened the insulating crust beneath which the granitic rocks then solidified. The cover above the coarse plutonic rocks when they crystallized was probably not thicker than 3 km. The earliest volcanic rocks included types less silicic than the magmas which crystallized at the exposed high levels of the intrusive complex.

The red color of most of the granite, granophyre, and rhyolite and their chemical character intermediate between normal calc-alkaline silicic rocks and the moderately alkalic silicic differentiates of lopoliths (Anderson, 1962; compare with Hamilton, 1960) might permit the inference that the St. Francois rocks are the silicic caprocks of a gabbroic lopolith, but gravity surveys indicate that the Precambrian complex in the St. Francois Mountains is batholithic rather than lopolithic. The Duluth and Mellen lopoliths of the Lake Superior region and the Wichita lopolith of Oklahoma (and the probable lopoliths forming a buried chain trending southwestward from Duluth through Minneapolis and Omaha to Abilene) are marked by great positive Bouguer gravity anomalies, with a relief of 100 mgals or more and steep gravity slopes. There is no such anomaly in the St. Francois region. (The lopolithic gravity anomalies are perhaps the most remarkable features shown on the United States gravity map by Woollard and Joesting, 1964.)

NEW ENGLAND APPALACHIANS

The Paleozoic orogenic terrane of the Appalachian system is best known in New England. Granitic rocks are widespread in this region but there are no great batholiths comparable to the late Mesozoic batholiths of western North America. Some reasons are suggested here for the contrasts between the Paleozoic and Mesozoic terranes of opposite sides of the continent.

Successive episodes of geosynclinal sedimentation and volcanism, deformation, metamorphism, and granitic intrusion have been superimposed complexly in New England, and many problems remain unsolved. Despite the overlapping of events, the major episodes of metamorphism and intrusion appear in a general way to be younger in the west than in the east and intermediate in the medial belt (fig. 5). Discussion here is restricted to that medial belt, whose analysis appears particularly relevant to the topic of this report. We are particularly indebted to Wallace M. Cady and Robert H. Moench for discussions clarifying New England geology. Summary published reports include those of Billings (1956); Billings, Rodgers, and Thompson (1952); and Goldsmith (1964).

The medial belt of New England is formed largely of sedimentary rocks and subordinate volcanic rocks, of Silurian and Devonian ages, which were metamorphosed and widely intruded by granitic masses during the Devonian. In New Hampshire, the belt is a sillimanite-grade "plateau" of highly metamorphosed rocks, injected by granitic plutons which generally are concordant in both structure and mineralogy to the wallrocks, although structurally discordant granites also are present in the northeastern part of the belt.

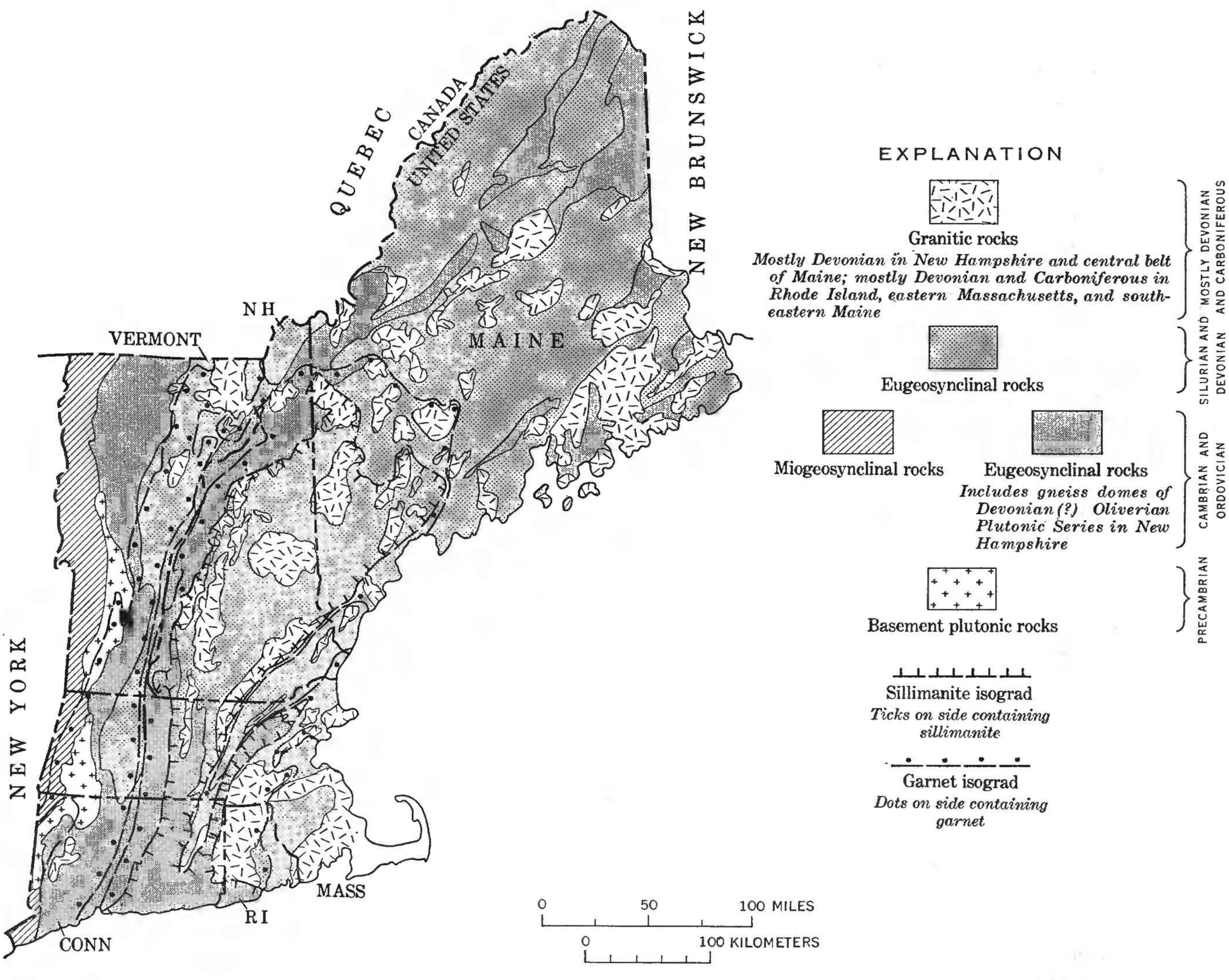

FIGURE 5.—Geologic and metamorphic map of New England. Generalized from Goldsmith (1964).

The high-grade belt is bounded on both sides by steep metamorphic gradients (figs. 5 and 6) which correspond approximately in some places, although not in others, to the contact zones with older (Cambrian and Ordovician) eugeosynclinal rocks. Outside the metamorphic gradients are less metamorphosed rocks whose pattern is much complicated by granitic intrusions of various ages and by superimposed metamorphisms which produced large areas of garnet-grade rocks but only small areas of sillimanite-grade rocks.

The sillimanite plateau of New Hampshire gives way northeastward along the strike in Maine to a terrane of low-grade metamorphosed Silurian and Devonian rocks. These are intruded by crosscutting masses of granitic rocks, mostly Devonian, and middle- and high-grade metamorphism is limited to narrow contact aureoles surrounding the intrusions. The regional biotite isograd, not shown in figure 5, crosses Maine northeast of the garnet isograd shown; most of the medial belt in Maine consists of chlorite-zone rocks.

Structures within the high-grade part of the New England medial belt are extremely complex. Lithologic units swirl, intersect, branch, and pinch out in patterns of great fluidity, and dips typically are gentle to moderate (Billings, 1955; Goldsmith, 1963, 1964). Structures are refolded isoclinally and irregularly (Goldsmith, 1961). Metasomatic changes and granitic injections are widespread. The metamorphic gradients bounding the sillimanitic terrane are locally so steep that biotite and sillimanite isograds are only about 1 km apart (fig. 6; Billings, 1955), although 8 km is a more common distance.

Petrologists in general agree that the biotite isograd represents a temperature of 200° to 300°C, whereas sillimanite forms near 600° to 700°C, or at about the temperature of granitic magma. Biotite and sillimanite isograds thus represent a temperature difference of about 400°C, and the temperature gradient in response to which they formed in New Hampshire was as steep as 300°C per km in the plane of the present ground surface. The common assumption that increasing metamorphism is due to increasing depth (for example, Turner, fig. 77, in Fyfe and others, 1958) is inadequate to explain such gradients (Hamilton, 1963a, p. 90). No model based upon heat conducted

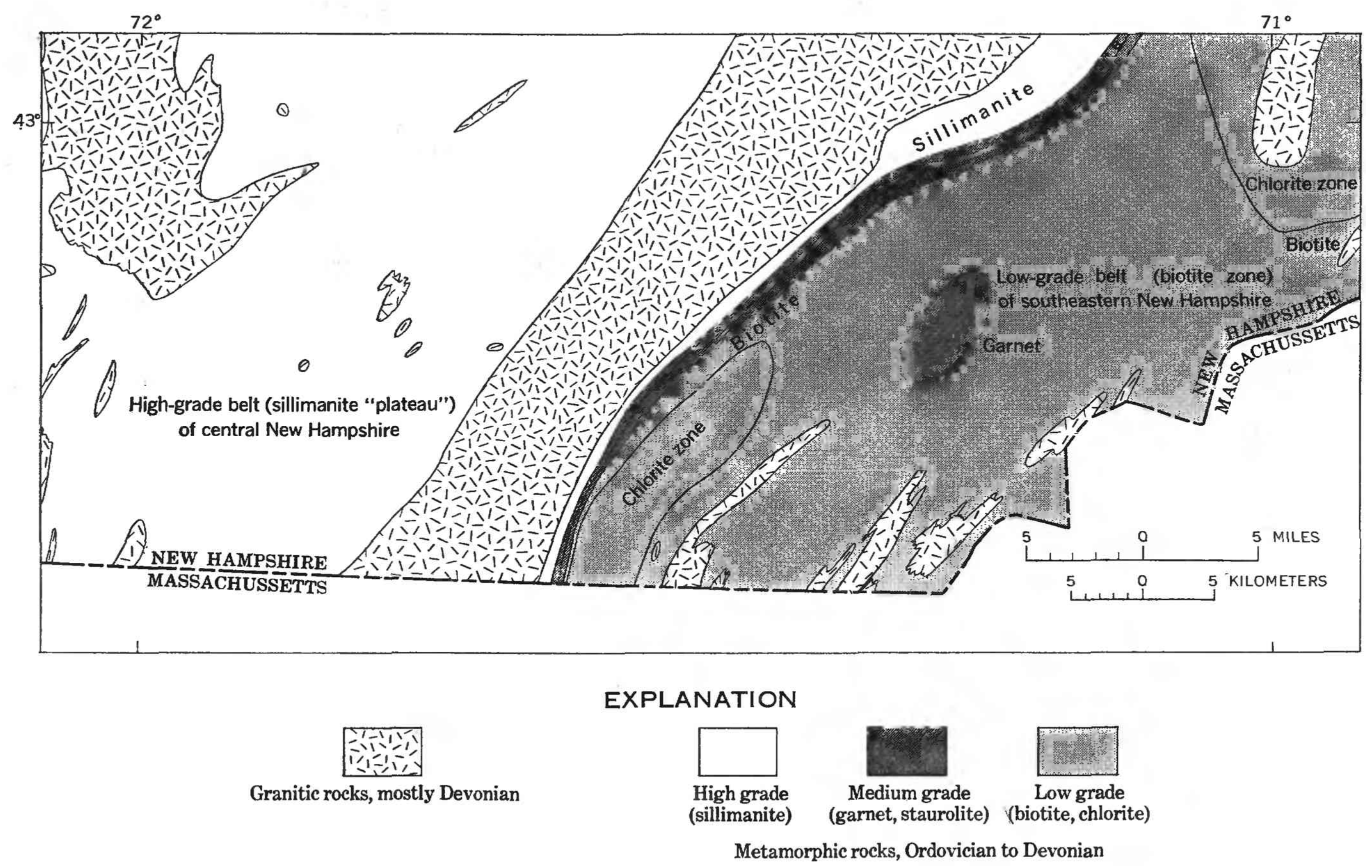

FIGURE 6.—Metamorphic zones and granitic rocks of a part of southeastern New Hampshire. A steep metamorphic gradient, defined by the sillimanite, staurolite, garnet, and, in part, biotite isograds, separates high-grade and low-grade terranes. Adapted from Billings (1955).

from deep levels in a static system can account for more than a small part of such metamorphism.

The apparent alternative is that the heat was introduced in magmas from hotter deeper levels. (Barrell, 1921, reached this conclusion long ago.) This requirement negates the speculation that the gneisses and migmatites of the sillimanite belt represent spontaneous partial melting due to tectonic depression into levels heated to melting temperatures by heat conducted from the base of the crust. The granitic component added to the complex came from markedly greater depths than those exposed.

Exposed granitic masses are distributed irregularly through the gneiss belt. Had they alone carried the heat upward, the metamorphic effects should be related concentrically to them, but this is not the situation. The belt is generally bounded by a straight and steep metamorphic gradient, irrespective of the local distribution of exposed plutons within the belt. Some might interpret these relationships as indicating the presence of a continuous great batholith beneath the exposed gneisses, but so steep are the flanking metamorphic gradients that such a batholith would have to be very near the surface, and it is unreasonable to postulate that the top of the batholith could everywhere be at a uniform shallow depth without being exposed.

The known factors of structure and metamorphism agree with the hypothesis that the gneiss belt of the New England Appalachians formed beneath a batholith analogous to that of the Sierra Nevada. Molten plutons may have risen through the gneisses and coalesced above them in a thin, shallow batholith. Granitic melts are much lighter than metamorphic rocks—even solid granitic rocks are lighter than most high-grade metamorphic rocks—and must rise through them wherever the buoyancy of the plutons exceeds the strength of the metamorphic rocks. The extremely plastic and irregular deformation shown by the gneisses indicates that their structures did not form by simple response to systematic regional compression but rather that the rocks flowed about in complex patterns of rising, sinking, underflowing, and overflowing. This in turn suggests that the dominant causes of flow were gravitational instability and differences in plasticity, as Rosenfeld (1960) concluded on the basis of detailed analysis of microscopic and macroscopic structures. The heating of the metamorphic rocks and their copious injection by granitic fluids caused them to flow readily outward, downward, and beneath the rising plutons. Much of the granitic material was enveloped and trapped in the gneisses, but most of it rose through to coalesce higher into batholiths, which have since been eroded away. The steep metamorphic gradients bounding the gneissic terrane may mark the margins of the region through which the plutons rose.

If this rationale is correct, then the northeastern limit of the sillimanite plateau in the medial belt also represents the limit of initially nearly continuous granitic plutons in the overlying batholithic complex that has since been largely eroded away. Crosscutting granitic masses are more abundant near the along-strike transition from high-grade to low-grade terranes than they are anywhere else in the medial belt (fig. 5); this is consistent with the suggestion that a thin batholith, initially continuous to the southwest but since largely eroded away, here gave way northeastward to a terrane of scattered smaller plutons. The lack of high-grade regional metamorphism in most of the medial belt in Maine is evidence that there is no batholith of regional extent beneath the belt there, and also suggests that there was never a batholith of regional extent above the levels now exposed by erosion.

Four heat-flow measurements in the Devonian granitic rocks indicated fluxes of only 1.2 to 1.7 microcalories per square centimeter per second (Birch and Roy, 1965); considered with the probably high radioactivity of these rocks, this low heat flow indicates the granites to be thin.

The northern Appalachian region was intruded during Mesozoic time by the stock and small batholiths of alkalic rocks of the White Mountains magma series. The relation between observed heat flow and measured radioactivity in these White Mountains rocks is such that rocks of high radioactivity, like those exposed at the surface, can extend only a few kilometers downward (Birch and Roy, 1965).

EROSION INTERVALS

Batholiths have crystallized and then been exposed by erosion within such short intervals of time in a number of places that it is difficult to visualize the granitic rocks as having formed at depths of many kilometers. Only a small fraction of a geologic period separates rocks deformed and intruded by batholiths from beds resting unconformably upon the same batholiths. The Baja California batholith, for example, intruded upper Lower Cretaceous strata and is overlain unconformably by middle Upper Cretaceous beds (Allen and others, 1960). Rocks as young as middle Upper Jurassic were metamorphosed and intruded by stocks and small batholiths in the Klamath Mountains of northwestern California, and the resulting crystalline complex is overlain unconformably by high Upper Jurassic beds (Irwin, 1960).

Zircon of middle and Late Cretaceous age, from batholithic rocks farther west, is abundant in sandstone throughout the Upper Cretaceous sequence of Montana and Wyoming (Houston and Murphy, 1962, 1965), and thus shows regional exposure of the batholiths to erosion within a very short interval after crystallization.

STRONTIUM ISOTOPES

Of the stable isotopes of strontium, some Sr^{87} is formed radiogenically by the decay of Rb^{87}, whereas the remaining Sr^{87} and all Sr^{86} and Sr^{88} are nonradiogenic. The ratio of radiogenic to nonradiogenic strontium in a magma thus provides an index to the ratio of rubidium to strontium in the source rocks from which the magma was derived; and as these two elements behave quite differently chemically—the rubidium being generally associated with potassium, the strontium with calcium—valuable information can be gained from this ratio. The strontium-isotope method of tracing geologic processes was developed by, among others, Hurley and his associates (Hurley and others, 1962), and has been applied by them (Fairbairn and others, 1964a, b; Hurley and others, 1965) and by many more, including Hedge and Walthall (1963) and Zartman (1965). In abundance, Sr^{87} is comparable to Sr^{86}, whereas Sr^{87} is far less abundant than Sr^{88}; hence analyses are generally made for the ratio Sr^{87}/Sr^{86}.

The young basalts of Hawaii and Iceland have Sr^{87}/Sr^{86} ratios ranging only from about 0.7023 to 0.7045 (fig. 7; Hedge and Walthall, 1963; Powell and others, 1965). The mantle sources from which their magmas were melted do not vary much in Sr/Rb ratios. Stony meteorites had an Sr^{87}/Sr^{86} ratio of about 0.698 at the time they formed (Hedge and Walthall, 1963; Pinson and others, 1965). No igneous rocks of any age yet analyzed have initial ratios that

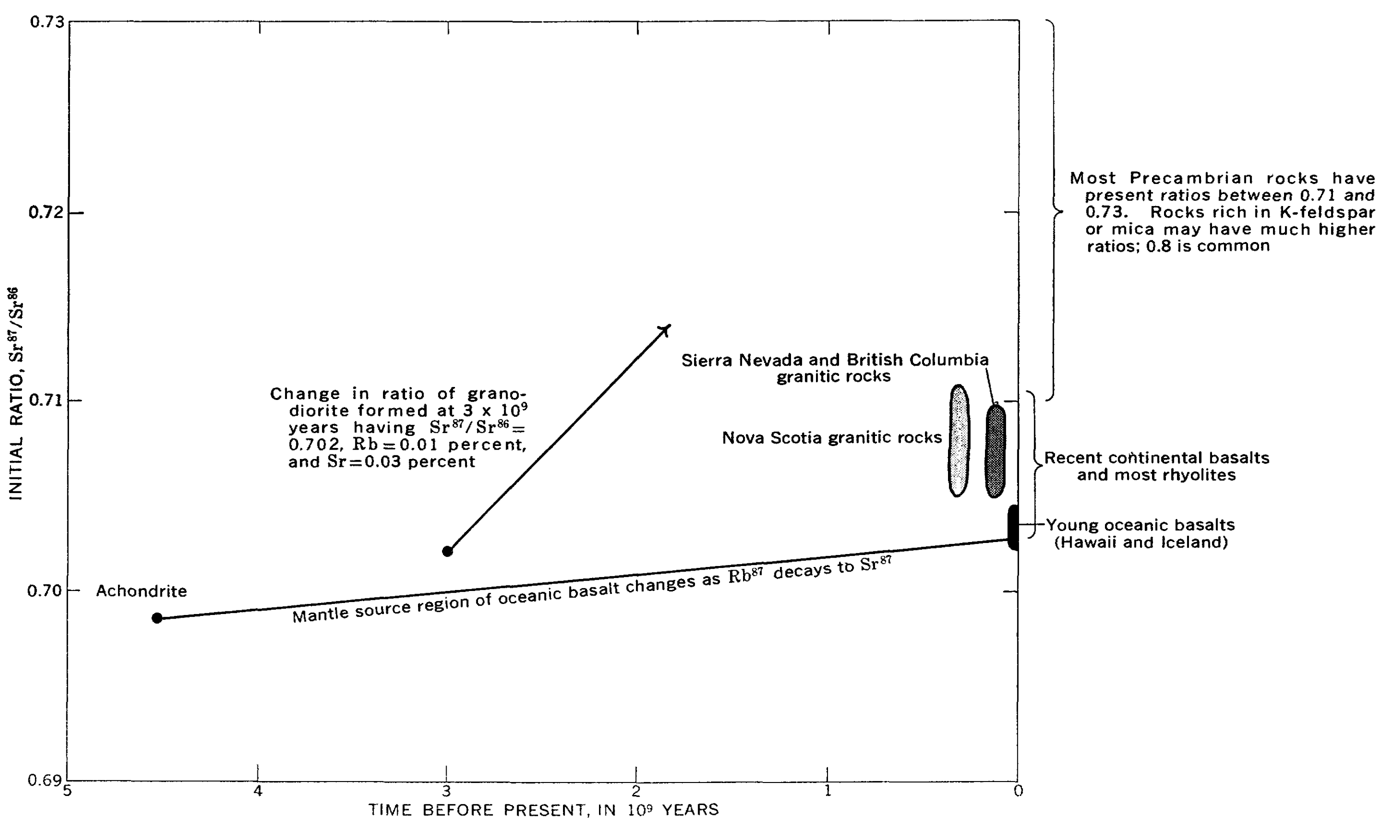

FIGURE 7.—Initial ratio of Sr^{87}/Sr^{86} in Paleozoic and Mesozoic batholithic granitic rocks. Because Sr^{87} includes that strontium produced by radiogenic decay of Rb^{87}, and Sr^{86} and the remaining Sr^{87} are not radiogenic, the ratio Sr^{87}/Sr^{86} is a function of the Rb/Sr ratio in the rocks from which the magmas were melted. The initial Sr^{87}/Sr^{86} ratios of granitic rocks, like those of most continental basalts and rhyolites, are a little higher than those inferred for the mantle source regions of oceanic basalts, but much lower than those of most Precambrian basement rocks. The granitic magmas might represent mantle melts which assimilated crustal rock, or melts of rock whose Sr/Rb ratio was intermediate between that of the mantle and that of Precambrian basement rocks. Data from Fairbairn, Hurley, and Pinson (1964a, b), Hedge and Walthall (1963), Hurley, Bateman, Fairbairn, and Pinson (1965), Pinson, Schretzler, Beiser, Fairbairn, and Hurley (1965), and Powell, Faure, and Hurley (1965).

fall as much as 0.001 below the line between the ratios of meteorites and the ratios of the lower limit of young basalts: initial ratios in all igneous rocks fall along the line or are above it by less than 0.005, or at the most 0.008. The slope of this line indicates an Sr/Rb ratio in the source region of oceanic basalt of about 50:1, and this ratio is the virtual lower limit for the source region of all igneous rocks (Hedge and Walthall, 1963). This Sr/Rb ratio of 50:1 is similar to, or a little lower than, that of tholeiite. Tholeiite must form by the melting of a large proportion of its mantle source rocks, which thus cannot be ultramafic.

The Sr^{87}/Sr^{86} ratios of Recent continental volcanic rocks overlap those of oceanic basalts and range upward only to about 0.71 (Hedge and Walthall, 1963). The ratios in continental rhyolites and basalts overlap throughout their ranges: the source rocks from which the basalt and rhyolite magmas are derived have about the same Sr/Rb ratios, and there is little basis in these data for the common assumption that basalt magmas are melted from deeper and much more mafic rocks than are rhyolite magmas.

Precambrian basement rocks have in general markedly higher ratios of rubidium to strontium than have the mantle source rocks of oceanic basalts, and hence the Sr^{87}/Sr^{86} ratio of average continental rock has increased more rapidly during geologic time than has the ratio in the mantle. (A line in fig. 7 illustrates the changing ratio in a typical Precambrian granodiorite whose Rb^{87} is decaying to Sr^{87}.) Even average Precambrian rocks—markedly more mafic than those most likely to be partially melted to produce granitic magmas—have present Sr^{87}/Sr^{86} ratios of about 0.720 (Faure and others, 1963). If silicic crustal rocks were melted directly to form any magmas, then the resulting igneous rocks should instead have initial Sr^{87}/Sr^{86} ratios plotting high above the achondrite-oceanic basalt zone of figure 7.

Granitic rocks crystallized from magmas whose initial Sr^{87}/Sr^{86} ratios were slightly higher than those of the mantle source regions of oceanic basalts but markedly lower than those of Precambrian plutonic rocks (fig. 7). The initial Sr^{87}/Sr^{86} ratio in the Paleozoic granitic rocks of Nova Scotia averaged 0.708 and only ranged from 0.705 to 0.711 (Fairbairn and others, 1964a), and ratios in the upper Mesozoic granitic rocks of southeastern British Columbia and the Sierra Nevada batholith were mostly from 0.706 to 0.708 (Fairbairn and others, 1964b; Hurley and others, 1965). The average ratio in Precambrian rocks in Paleozoic and Mesozoic time would have been at least 0.715, and many felsic Precambrian rocks would have had ratios of 0.8 to 1.0 and more.

Similar relationships hold for granitic rocks of all ages. Thus granitic rocks, 1,000 to 1,120 m.y. old, of the Llano Uplift of Texas had markedly lower Sr^{87}/Sr^{86} ratios at the time of their crystallization than did their wallrocks (Zartman, 1965). Tertiary granitic rocks of Colorado and Washington had initial Sr^{87}/Sr^{86} ratios of about 0.705, and some Precambrian granitic rocks, 2,400 to 2,800 m.y. old, of the Precambrian shield had ratios of only 0.701 to 0.703 (Hedge and Walthall, 1963). Some small Paleozoic granites also had very low ratios, such as 0.703 (Czamanske, 1965).

As the various authors cited have emphasized, these ratios of radiogenic to nonradiogenic strontium in granitic rocks preclude the possibility that granitic magmas are wholly melted from mantle rock like that which yields oceanic basalts, and also preclude the possibility that they are wholly melted from silicic crustal rocks like those exposed in Precambrian terranes. The source materials for granitic magmas are richer in rubidium (of which the radioactive isotope decays to Sr^{87}) than are oceanic mantle rocks, but they are poorer in rubidium than are exposed crustal rocks. Obviously the magmas either are melted in an environment of intermediate composition, or else are hybrid products of melts combined from contrasting rocks.

The strontium-isotope data considered alone permit the speculations that granitic magmas are derived in the lower (intermediate) continental crust; that they represent mixed mantle and crustal materials; or that they are melted from eugeosynclinal volcanic rocks and volcanogenic sedimentary rocks. The eugeosynclinal-melting hypothesis is attractive for some granitic rocks but obviously is not applicable to the Boulder batholith (which formed in a miogeosyncline) or to the granitic rocks of southeastern California, Colorado, and elsewhere, which were intruded into Precambrian basement complexes.

The ratios between the various stable and radiogenic isotopes of lead and their parental uranium and thorium in granitic rocks provide another isotopic method for placing limits on the composition of the source rocks from which the magmas were melted. The lead relationships are being investigated by Bruce R. Doe (written commun., 1965), who finds that the lead isotopes generally support conclusions similar to those reached by the strontium-isotope researchers.

ORIGIN AND EMPLACEMENT OF BATHOLITHS

The examples cited in this paper are all from the United States. The local interpretations made could be supported further by other North American examples, and further yet by numerous examples from other continents. The explanation of the examples is accordingly integrated here into a general theory of the characteristics of batholiths.

THE ENVIRONMENT OF BATHOLITHS

Most large Phanerozoic batholiths lie at least partly within eugeosynclines, and this has led to the popular tectogene hypothesis, whereby batholiths are visualized as the products of downbuckling and partial melting of the geosynclinal fillings. Batholiths also form, however, in such other environments as miogeosynclines (for example, the Boulder batholith), unstable platforms (as in the San Juan Mountains of Colorado), and oceanic island arcs (including the Aleutian Islands). The Idaho batholith intrudes both eugeosyncline and miogeosyncline. Boundaries between eugeosynclinal and miogeosynclinal suites shift back and forth through time, and in California the boundaries are crossed at high angles by the Sierra Nevada batholith (Gilluly, 1963, 1965).

Precambrian basement rocks also are intruded by batholiths, in geosynclinal as well as nongeosynclinal settings. In southern California in the San Gabriel and Orocopia Mountains (Crowell and Walker, 1962), in the West Riverside Mountains (Warren Hamilton, unpub. data, 1966), and elsewhere, large plutons of upper Mesozoic granitic rocks intrude Precambrian basement plutonic complexes as well as Paleozoic and Mesozoic metasedimentary and metavolcanic rocks.

The common association of batholiths with eugeosynclines seems best interpreted to indicate that the batholiths and the geosynclinal volcanic rocks have related origins, rather than that the batholiths form because the geosynclines are present. Batholiths form wherever temperatures are high enough at depth to melt the needed magmas, and eugeosynclines are only one setting in which these conditions are met.

ORIGIN OF GRANITIC MAGMAS

The discussions in previous sections indicate to us that all or most granitic magmas are generated at depths below any ever exposed by erosion. Strontium-isotope data preclude the possibility that most granitic (and silicic-volcanic) magmas could form by melting of silicic crustal rocks alone. Sources permitted by the isotope data include volcanic rocks and volcanic sedimentary rocks in geosynclines; gabbroic or amphibolitic rocks of the lower continental crust; and mantle rocks, provided some continental rock relatively high in rubidium is assimilated by rising magmas. Geologic, petrologic, and physiochemical reasoning provides much further information on the origin of the magmas.

Some granitic masses are so closely associated with high-alumina volcanic rocks that their magmas must have come from the same source. The granodiorite batholiths of the Cascade and Aleutian provinces share the compositional patterns of the volcanic rocks they intrude and of the volcanic rocks which overlie them, and appear to be but different aspects of continuing magmatism from the mantle. Pregranitic deformation of the older volcanic rocks in these provinces was slight so that major orogeny cannot be postulated to have intervened.

Some granitic magmas may form by melting of eugeosynclinal materials, but many cannot possibly form in this way. In the Boulder batholith and the eastern part of the Idaho batholith, for example, no volcanic piles were available. In other batholiths (for example, in southern California) the magmas intruded subgeosynclinal basement rocks and hence came from still deeper sources. The rocks formed from such demonstrably noneugeosynclinal magmas cannot be distinguished from those (for example, in the Sierra Nevada) for which a eugeosynclinal origin might otherwise be postulated. The hypothesis that granitic magmas are generated by downbuckling and partial melting of eugeosynclinal materials is certainly inapplicable in many places and may be wholly invalid.

If there is a single magma-generating mechanism, it must thus be more general than that of geosynclinal downbuckling. The mechanism must produce granitic magmas beneath diverse geologic environments. Quartz diorite, granodiorite, and quartz monzonite are the dominant products required in Phanerozoic terranes, but their relative abundance differs widely from region to region. (In some Precambrian orogens, potassic granite is dominant or at least abundant, but it is lacking or is much subordinate to less potassic rocks in nearly all Phanerozoic suites.) Quartz diorite and trondhjemite are the dominant types in some regions, and gabbro and diorite are present in many assemblages and are abundant in some.

Compositions inferred for the lower crust and upper mantle must be incorporated into the explanation. Seismic, heat-flow, and petrologic data provide clues. In the uppermost mantle, seismic waves generally travel faster beneath stable regions of the continental crust than beneath orogenically and volcanically active regions (Herrin and Taggart, 1962; Nuttli, 1963; Pakiser, 1963). Heat flow is in general higher in active regions than in stable ones (Lee and Uyeda, 1965). These relations preclude the possibility that peridotite forms the uppermost mantle in both active and stable regions: pressures in the uppermost mantle are too low to invert olivine (Wentorf, 1959), so that density-phase transformations cannot be called upon; and neither can the variable hydration of peridotite to serpentine be postulated, for hydration would be suppressed in the high-temperature regions, and hence would correlate with heat flow in the direction opposite to that required by the seismic data. Pressure and temperature at the base of the continental crust in stable regions, but probably not in active ones, are appropriate for the eclogite transformation of plagioclase and low-alumina pyroxene to the denser phases of garnet and aluminous alkalic pyroxene (Yoder and Tilley, 1962, fig. 43). Velocity data determined experimentally at high pressure (Birch, 1960) permit the suggestion that the Mohorovicic discontinuity at the base of the continental crust represents a compositional change, from basalt above to basalt plus dunite below, and that in the low-velocity mantle and lowermost crust of active regions the basaltic phase is in low-density plagioclase and pyroxene or amphibole. In the high-velocity mantle, and probably also in the deepest crust, of stable regions the basaltic phase is in high-density pyroxene and garnet, but in the low-velocity mantle of tectonically active regions the transformation to those dense eclogitic minerals occurs well within the mantle (Pakiser, 1965). Another possibility is that high-velocity mantle contains more olivine than does low-velocity mantle. Such a contrast might develop as basalt is wrung out of the upper mantle, residual materials being progressively more peridotitic and dunitic.

Kimberlite and its dense inclusions provide direct evidence for the existence of mantle rocks of basalt-plus-olivine composition. Kimberlite consists of large corroded crystals of high-pressure minerals—forsterite, jadeitic clinopyroxene and aluminous orthopyroxene, pyrope and almandine, phlogopite, diamond, magnesioilmenite—in a matrix of serpentine, chlorite, olivine, calcite, and phlogopite. Most kimberlite contains inclusions of dense rocks composed of varying combinations and proportions of heavy minerals like those of the corroded crystals (Holmes, 1937; O'Hara and Mercy, 1963; Wagner, 1914; Williams, 1932). Kimberlite has a peculiar bimodal composition of ultramafic components on the one hand and alkalic and volatile components on the other, and is probably a mixture of mantle rock—like the dense inclusions—with mantle-derived water, carbon dioxide, and alkalies. The average composition of the dense inclusions, and of kimberlite itself minus the 15 or 20 percent of apparently added volatile and alkalic components, is approximately equal to a mixture of equal parts of tholeiitic basalt (in the mineralogic form of high-pressure pyroxenes and garnet) and magnesian olivine.

If these inferences are correct, then the upper mantle and lower crust differ primarily in olivine content. As olivine is more refractory than any other major mineral in the rocks, it would not be melted appreciably; therefore, only the melting of the basaltic phase need be considered here, and such discussion may be equally applicable to upper mantle and lower crust.

High-pressure laboratory data show that the dense alkali-bearing aluminous pyroxenes and the garnet of basalt-composition pyroxenite and eclogite melt together over a rather narrow temperature range under both anhydrous and hydrous conditions at pressures appropriate to the upper mantle or deep continental crust (fig. 8; Cohen and others, 1966; Yoder and Tilley, 1962, tables 43–47). Tholeiite (basalt low in alkalies, moderately low in alumina, and saturated in silica) and olivine-tholeiite magma could originate by the nearly complete fusion, within such a narrow temperature range, of the pyroxenitic or eclogitic fraction of mantle rock. Partial melting under such conditions to produce granitic magmas is not however possible. No substantial quantity of silicic magma can be generated by direct melting at depths greater than that of the gabbro-to-eclogite pressure transformation if the meltable part of the mantle is of basaltic composition.

At pressures corresponding to the uppermost few miles of the crust, pyroxene and plagioclase melt or crystallize together over a limited temperature range

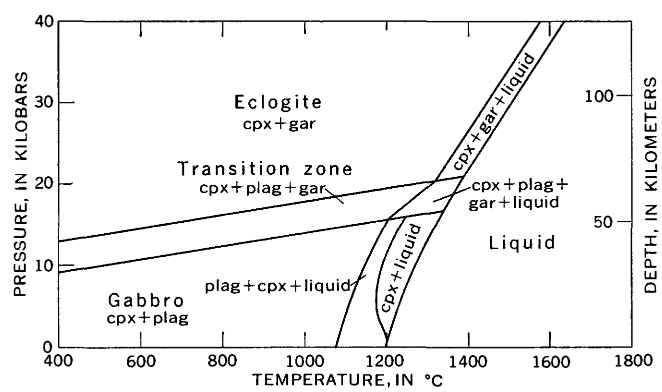

FIGURE 8.—Pressure-temperature phase relationships in material of anhydrous basaltic composition. The data plotted are those of Cohen, Ito, and Kennedy (1966), with additions consistent with the data of Yoder and Tilley (1962). Olivine or orthopyroxene or both would be present in rocks of appropriate compositions—the liquidus and solidus curves are in fact those of olivine tholeiite—but are not illustrated here. Cpx, clinopyroxene; gar, garnet; plag, plagioclase.

(Yoder and Tilley, 1962, figs. 6, 27–30). Crustal temperatures, however, preclude melting at such shallow depths.

The melting behavior of basaltic rock at pressures appropriate to the lower crust or uppermost mantle, but shallower than the gabbro-eclogite transition, is quite different. Plagioclase at these intermediate pressures is a markedly lower temperature mineral than is either pyroxene or amphibole (Cohen and others, 1966; Yoder and Tilley, 1962, figs. 27–30). The temperature range of crystallization or melting is broad and increases with pressure. Progressive melting under these shallower-than-eclogite conditions will thus produce highly feldspathic magmas progressively richer in calcic plagioclase and in mafic minerals. Such melting could produce the high-alumina basalt, andesite, and dacite which typify island arcs and eugeosynclines (Hamilton, 1964).

This mechanism might provide an explanation for the melting of intermediate and silicic magmas from rock of basaltic composition and thus from the lower crust or mantle. Mantle rocks of the same composition that yield tholeiitic basalt magma by partial melting beneath the eclogite transformation could yield high-alumina basalt, andesite, and dacite magma by less complete melting above it. (High-alumina basalt differs from tholeiite primarily in being richer in the components of calcic plagioclase.) The intimate association of tholeiite and high-alumina rocks in some island arcs and also in some continental environments can be explained easily in this way. The contrast between midoceanic volcanism (tholeiite on the sea floor, tholeiite plus alkaline olivine basalt on islands) and island-arc volcanism (tholeiite plus high-alumina volcanic rocks) can then be explained in terms of variations in pressure-temperature relations in the mantle beneath the contrasted provinces, without any need for variations in mantle compositions. The curves of actual temperature and of the melting temperature of the basaltic component are probably close together and subparallel within a considerable thickness of the upper mantle (MacDonald, 1964, fig. 15), so that such variations in conditions need not be great.

Differentiation by partial crystallization of a magma within the crust or upper mantle, but above the eclogite transformation boundary, would produce liquids of the same compositions as would partial melting there, as O'Hara (1965) emphasized. The composition of a magma provides clues as to the depth of its last equilibration with crystals, but not as to whether this equilibration was achieved by partial melting or by partial crystallization.

These mechanisms of deep derivation of magmas by varying degrees of melting satisfy some of the conditions but not all imposed by the data. How, for example, do voluminous granitic magmas form that are almost devoid of mafic components? Why are silicic magmas nearly lacking in the ocean basins except in those places—as, island arcs, and Iceland—where abnormally thick volcanic crust is present? Regional relationships yield further data to be considered.

Available data suggest parallel regional variations in compositions of eugeosynclinal volcanic rocks and the granitic rocks which intrude them. Magmatism of the type that produced the volcanic piles could also have produced the batholiths. The eastern part of the Sierra Nevada batholith consists largely of light-colored quartz monzonite and granodiorite, and Mesozoic metavolcanic rocks are mostly dacite and quartz latite (Rinehart and Ross, 1964, p. 30–38; Bateman and others, 1963, p. 6). The western part of the batholith consists in general of more mafic and calcic rocks, and the Mesozoic volcanic rocks are correspondingly also more mafic and calcic on the average (Bateman and others, 1963, p. 6). Similarly, metavolcanic rocks along the west side of the Idaho batholith are basalt, andesite, and dacite, all low in potassium, and the granitic rocks intrusive into them are quartz diorite and trondhjemite (Hamilton, 1963a, table 5 and pls. 1, 2). Metavolcanic rocks are lacking in the eastern part of the batholith, which is of granodiorite and quartz monzonite.

The western part of the belt of late Mesozoic batholiths of western North America is dominated by quartz diorite, and the central and eastern parts are dominated by granodiorite and quartz monzonite (Moore, 1959, fig. 2). Silica and potassium are higher inland, whereas calcium, iron, and magnesium are higher oceanward. One possible explanation of this difference is that when the batholiths were forming the Mohorovicic discontinuity lay deeper beneath the inland region than beneath the coastal one and that the inland magmas were melted largely within the lower crust whereas the coastal ones came from the upper mantle. An inference proceeding from this postulate and from factors noted previously is that although the lower crust and the meltable part of the upper mantle are both grossly basaltic in composition, the lower crust is richer in potassium and is less mafic.

A similar postulate can be made for the melting of magmas during middle and late Cenozoic time throughout the Western United States. The ratio of potassium to sodium in high-alumina rocks of any given silica content tends to increase with altitude and hence with crustal thickness. This is demonstrated by the data plotted by Moore (1962, figs. 2, 3): the K/Na ratio tends to increase as Bouguer gravity becomes increasingly negative.[2] (Bouguer gravity in general correlates broadly with regional surface altitude and with crustal thickness.) The bulk composition of the high-alumina volcanic suites also tends to become more silicic with increasing altitude. This can be seen by considering the many provincial plots of Moore's figure 1 as frequency-distribution diagrams. These relationships can be explained if the magmas are melted entirely in the mantle where the Mohorovicic discontinuity is relatively shallow and partially or entirely within the lower crust where it is deep. Gradations in magma types reflect gradations in crustal thickness and hence might be due either to melting together of both crustal and mantle materials in various proportions or to assimilation of deep-crust material in magmas rising from the mantle. Gilluly (1965, p. 28) made similar suggestions.

A mechanism which would permit generation of a magma in the mantle, followed by great assimilation of crustal materials in the rising magma, and which is capable of producing wholly leucocratic rocks, would appear to satisfy all the requirements set forth. Such a mechanism is available if we combine Dickson's concept of zone melting with that of partial melting within the upper mantle and lower crust.

Part of the rise of any magma must be accomplished by zone melting (Dickson, 1958), and this could result in profound modification of the original magma by contamination. The pressure gradient to which any magma of appreciable vertical extent is subjected must cause migration of the most volatile components toward the top of the chamber. This results in a lowering of the melting temperature at the top and a raising of the melting temperature at the bottom, so that roof rocks are melted and incorporated into the melt, whereas the basal magma is forced to crystallize. The rising magma becomes progressively enriched in the lowest melting components of both the initial magma and the rocks through which it passes. The energy needed to melt the roof comes from crystallization at the base. The energy loss as cold rocks are heated is partly compensated for by the lowering of temperatures of fusion as pressure decreases with rise of magma to higher levels, and by the rise of magma through a graded or layered crust which becomes less refractory upward.

The final magma resulting from such a rise by zone melting could contain only a trifling quantity of material present in the initial melt: an energy envelope has risen through the crust, but only the most volatile of the initial components are present in the final magma. Magma can be mobilized beneath the crust, and yet the final high-level igneous rock can consist largely of components derived from the crust. The correlations between compositions of granitic rocks and the compositions of the columns through which they rose can thus be easily explained, as can the leucocratic character of many granitic rocks and the lead and strontium isotopic relationships in them.

The effects of differentiation, both by fractional crystallization and by upward migration of the less refractory components, must further complicate the evolution of the magmas. Zone melting of course represents a combination of assimilation and differentiation.

The mechanisms suggested here require that the crust be continually growing as material is added to it from the mantle. The general restriction of potassic batholiths to terranes more than 1,000 m.y. old perhaps indicates that the mantle and crust became by that time so differentiated that potassium-rich magmas could no longer be generated in great volume from the mantle.

Silicic and high-alumina magmas form primarily in long belts, initially either oceanic or continental, whose tectonic mechanisms are not yet apparent; but the magmas can form anywhere in the high-heat-flow regions of the continents.

EMPLACEMENT OF BATHOLITHS

Batholith magmas are melted in the lower crust and upper mantle. The buoyant magma masses rise and probably become completely detached from their zones of melting. Overlying rocks are heated and displaced outward, then sink and flow beneath the rising plutons, becoming intensely metasomatized and injected by granitic material. Much additional granitic material is enveloped by the flowing gneisses, joins in their irregular motion, and crystallizes in concordant foliated sheets.

Such interpretations have been made by other geologists. Hans Cloos (1923) concluded that many batholiths are connected with deep sources by dikelike channels rather than by full-size chambers and that concordance and discordance are determined by the structure of the country rock. Chamberlin and Link (1927) suggested that batholiths are shaped like sheets, tongues, and mushrooms, which have narrow

[2] Moore plotted high-alumina and other provinces together. The relation between the K/Na ratio and gravity anomalies becomes more regular when only high-alumina provinces are considered, and so provinces of tholeiite, olivine basalt, basalt and rhyolite, and highly alkaline rocks—which all may originate beneath the eclogite boundary—are logically omitted.

feeders and spread out high in the crust. Lane (1931, p. 823) wrote of batholiths as "intrusions on the surface"—flat mushrooms, reaching the surface in some places and spreading out between basement and overlying sediments in other places, fed by relatively narrow conduits. Bott and Smithson (1966) concluded from gravity analyses that granitic plutons extend typically to depths of only about 10 km.

The plutons of magma rise until crystallization, forced by loss of heat and volatiles, prohibits further flow, or until they reach the surface. As the plutons stiffen, heating of the wallrocks decreases, metasomatism virtually stops, and thin aureoles of thermal metamorphism result. Wallrock screens are dragged upward and pushed outward by rising and expanding plutons, and roofs are raised. The deformation of the metamorphic rocks may be due largely to emplacement of the plutons.

The larger and more numerous the plutons, the higher they should rise. Many must reach the surface and produce voluminous volcanic eruptions. Any pluton that reaches the surface over a wide area is roofed only by its own volcanic ejecta, and granitic textures develop as the magma crystallizes beneath its insulating volcanic cover.

Large batholiths are formed by the coalescence of many plutons and may be less than 10 km thick and may be unroofed over broad areas. The plutons rise a minimum of perhaps 40 km before reaching their site of final crystallization, which is generally within a few kilometers of the surface. The loss of volatiles by eruption through roof rocks must stop the rise of many plutons, but it is difficult to visualize any mechanism which could prevent the broad surface breakthrough of many other large masses of light magma.

The exposed parts of the Boulder and St. Francois batholiths are covered almost entirely by consanguineous volcanic rocks, and such a relationship may be common. Volcanic crusts may float on broad plutons of batholithic magma.

The Boulder batholith probably is floored by a downbowed surface of prevolcanic rocks. This shape is indicated by the way the flanking rocks dip right side up beneath the batholith on the north and probably on the south; the shape is also indicated by the horizontal flow shown by internal structure in the west part of the batholith, although the west contact is hidden. The sagging presumably accompanied the withdrawal of magma at depth and its eruption at the surface. Near both east and west margins of the batholith, the volcanic roof rocks dip monoclinally outward, perhaps in response to lifting by the growing batholith. The final major eruptions of magma through the preigneous rocks apparently were primarily intrusive; the first eruptions were certainly volcanic. Extrusion through the volcanic crust and intrusion beneath it probably occurred simultaneously during much of the history of the complex: new injections of magma beneath the crust fed eruptions at the surface so that crust and batholith thickened simultaneously, although the volcanic rocks spread far beyond the batholith. The sharp contacts between some plutons in the batholith show sporadic intrusion and crystallization and it cannot yet be said how much of the batholith was molten at any one time. The size of the vents through the prevolcanic rocks, which fed the shallow batholith and its volcanic cover, is not known, but the great area of the complex suggests that they were large.

The central part of the Sierra Nevada batholith is flanked by steeply dipping metamorphic rocks whose tops face inward toward the batholith. If this structure is related to the batholith, rather than being an old structure fortuitously intruded by the granitic magmas, then conceivably it has an explanation similar to the structure flanking the Boulder batholith. The Sierran plutons may also have reached the surface and spread laterally as the wallrocks were pushed outward and downward. If this is so, the volcanic cover once above the plutons has been removed by erosion. The overlapping ages of granitic rocks and wallrocks complicate any interpretation, and some of the metavolcanic wallrocks may have formed as roof rocks of early plutons.

The largest Phanerozoic batholiths now exposed are of late Mesozoic age. This fact suggests either that this was a period unique in the earth's history or else that batholiths are features of the uppermost crust and are so thin that 200 million years' erosion commonly suffices to remove them. We prefer the latter alternative. Most gneiss terranes may have formed beneath batholiths since removed by erosion, and the steep metamorphic gradients bounding some gneiss terranes may mark the limits of the belts through which the plutons rose. There are, however, many large Paleozoic and Precambrian batholiths, so either such batholiths were less eroded than their largest correlatives, or else they extended to greater depths than we infer.

The rise of plutons can be likened to that of salt domes. Magma and salt both rise because they are lighter than the rocks above them, although the density differential is probably in general greater for magma and metamorphosed wallrocks than for salt and indurated sedimentary rocks. Isolated salt plugs form where the supply of salt is small, whereas salt megadikes form where the supply is great. The varied salt structures demonstrated by drilling in northern Germany and described by Trusheim (1960) are strikingly similar geometrically to various Late Cretaceous and early Tertiary igneous complexes in western Montana. An inference to be drawn from the salt-dome analogy is that plutons of magma do not mark the locations of hot spots in the mantle, but rather that magma is generated over broad regions at depth and coalesces into masses whose spacing is controlled by the supply of magma and whose position and shape are controlled by structural features of the crust. Batholiths form where the supply of magma is so great that the masses coalesce and rise toward the surface as large plugs and megadikes.

BATHOLITHS AND METAMORPHISM

The deformation and heating shown by metamorphic rocks may be largely products rather than causes of magmatism. The steep metamorphic gradients of many terranes of regionally metamorphosed rocks cannot be quantitatively explained as due to conducted geothermal heat, but can be easily understood as produced by contact metamorphism on a regional scale. Gneisses and migmatites, ascribed by many writers to anatexis (partial melting in place due to ultrametamorphism), may instead be produced largely by metasomatism and injection by rising plutons. The highly plastic flow patterns of gneisses may form while wallrocks sink and close beneath rising plutons. The contortions in low-grade metamorphic rocks flanking high-level batholiths also may have been produced largely by rising plutons.

Similar suggestions have been made by many authors. Sederholm (1919, p. 250) wrote of "Kontaktmetamorphose regionaler." Barrell (1921, p. 255) concluded that "batholithic invasion is * * * one of two major factors in dynamometamorphism." Buddington and Chapin (1929, p. 50) used the phrase "contact metamorphism on a regional scale."

The metamorphic rocks of the east-central Sierra Nevada include mineral assemblages typical of the contact-metamorphic hornblende hornfels facies and the regional-metamorphic almandine amphibolite facies of Turner (in Fyfe and others, 1958). Bateman, Clark, Huber, Moore, and Rinehart (1963, p. D 11) assumed that this combination indicates that the rocks formed at a depth whereat the conditions of contact metamorphism were giving way downward, because of increasing pressure, to conditions of regional metamorphism; and they accepted Turner's (fig. 107 *in* Fyfe and others, 1958) guess that this transition might occur at a pressure corresponding to a depth of about 20 km and specified this as the depth of crystallization of the Sierran plutons. (Interpretations based on granitic rock compositions led Bateman and others to infer a water pressure of 5 kbars, equivalent to about the same depth.) The contrasting metamorphic mineral assemblages, however, occur in rocks of different compositions, and we regard this as evidence that the mineralogical contrasts reflect compositional differences rather than some critical pressure and that the distinction made between contact and regional metamorphism has no meaning here. The metamorphic grade and degree of recrystallization of the wallrocks decrease away from granitic contacts (Bateman and others, 1963, p. D 11), and the regional-type metamorphic rocks clearly owe the heat of their metamorphism to the nearby intrusive plutons. The pressure assigned the almandine amphibolite facies by Turner was based on the assumption that rocks of that facies were heated by conducted geothermal heat. If such an assumption is accepted, great depth must be postulated even if a steep thermal gradient is also assumed; but as this heat demonstrably came from intruded magmas, and not from deep burial, the assumption is not valid. Hamilton (1963a, p. 89–93) raised numerous objections to the depth-zone correlations and to the oversimplified applications of facies concepts that are still prevalent in the literature of metamorphic petrology.

Not all medium- and high-grade metamorphism can be explained as regional effects of batholithic intrusion, however. The Blue Ridge province of the southern Appalachians, for example, comprises a metamorphic plateau of general staurolite- or kyanite-zone rocks with local highs of sillimanite-zone rocks (for example, Bryant, 1962, p. 20–23). Upper Precambrian rocks were metamorphosed progressively to these grades during early or middle Paleozoic time, when the older Precambrian basement rocks (which make up most of the Blue Ridge) were metamorphosed retrogressively to the same grades. The metamorphic slope bounding the plateau is rather gentle: the biotite and kyanite isograds are separated by an average distance of 20 km in the Great Smoky Mountains (Hadley and Goldsmith, 1963, pl. 3). Known Paleozoic igneous activity in the province is limited to small dikes and to large and small pegmatites in scattered areas. Inasmuch as effects of metasomatism are limited to these areas, passage of plutons upward through the entire exposed terrane cannot reasonably be postulated, nor is there evidence to suggest the presence of a hidden batholith beneath the entire province. Severe deformation accompanied the metamorphism, and it is likely that the temperatures of deep burial were much increased by deformational heating.

BATHOLITHS AND THRUST FAULTING

Many thrust faults are temporally and spatially related to batholiths. The relationship might be ascribed to various factors, but one that must be considered in each such relationship is that the intrusion of a batholith is responsible for the thrusting. Keith (1923, p. 365–375), among others, has proposed such mechanisms, which are worthy of serious evaluation even though they are not currently popular.

The belt of late Mesozoic batholiths of the Western United States is flanked on both sides by broad terranes deformed by thrust faults during late Mesozoic and very early Cenozoic time. On the east is the Laramide thrust belt, whose east margin is coincident with that of geosynclinal strata and trends south-southeastward through western Montana, southward across westernmost Wyoming, south-southwestward across western Utah and southern Nevada, and intersects the crystalline terrane in southeastern California. The belt widens again in southern Arizona and trends into Mexico. The distance of the east margin from the Idaho, Sierra Nevada, and related batholiths ranges from 50 km in southeastern California to as much as 500 km across Nevada and western Utah. (The maximum width of the belt may have been less than 300 km when it formed, however, for the wide sector may have undergone great extension during the formation of basin-and-range structures and of volcanic piles during Cenozoic time.) The geosynclinal assemblage is shingled by thrust faults eastward from the edge of the batholiths (although interpretation is clouded by the widespread late Paleozoic thrust faults present at least in central Nevada). Major thrust faults generally are nearly parallel to bedding of the rocks. Faults in the west and central parts of the belt tend to be subhorizontal or gently folded, and many carry younger strata over older strata, whereas major faults in the east generally dip westward and carry older rocks over younger. Thrusting probably preceded folding in most regions, and most of the folding is better regarded as an effect of the sliding that produced the thrusting than as a cause of thrusting. At least in western Wyoming and southeastern Idaho, thrusts in general become younger eastward within the belt (Rubey and Hubbert, 1959, p. 188). The thrust-belt rocks are not metamorphosed except close to the batholiths on the west. Individual faults have horizontal displacements reaching at least 40 km, as proved by older-over-younger exposures in southeasternmost British Columbia, and the aggregate displacement across the belt may reach several times this amount. Basement rocks do not appear in the overthrust sheets.

The lack of basement rocks in the overthrust sheets, despite great displacements, requires that the exposed faults are confined to the geosynclinal fill and do not reach the basement. It has been inferred from this relationship, in the Western States and elsewhere, that the strata slid off some highland arch—but the thrust-faulted terrane extends westward to the batholithic belt, so such a hypothetical highland cannot be placed anywhere east of the batholiths. Possible explanations are that the blanket slid eastward off the batholiths themselves or that the rising and spreading batholiths shouldered the sedimentary rocks eastward. The lack of basement rocks in the thrust sheets would require that, in the latter explanation, the batholiths expanded eastward above the basement rocks.

The thrust belt is particularly irregular in the region of the Boulder batholith and neighboring small batholiths which lie within the belt in southwestern Montana. (See the geologic map of Montana: Ross and others, 1955.) The east margin of the belt swings from a southwestward trend to a southeastward trend to define a curve rudely concentric to the Boulder batholith and 50 km or less from it. Trends of thrust faults within the belt are very irregular because early structures have been deformed by later ones which have a tendency toward concentricity about the batholith. Jumbled thrusts trend northward a similar distance west of the batholith. One possible explanation is that shouldering by the batholith produced much of the thrusting. As the batholith thickened, rocks of its inward-sloping floor would be forced plastically outward. Another alternative is that once formed, the stiff plate of the batholith greatly influenced the response of the sedimentary terrane to thrusting due to other causes. Both processes may have operated together.

We interpret published geologic maps of southwestern Montana as indicating that a relatively simple system of east-directed thrust faults had been formed, perhaps by the shouldering action of the Idaho batholith, before the eruption of the Boulder batholith and its neighbors at the end of Cretaceous time. The younger batholiths superimposed complex local patterns of outward thrusting upon this eastward-thrust terrane. We interpret the structure of the northern Flint Creek Range (McGill, 1965), for instance, as recording the superposition of generations of folds and thrusts in this sequence: (1) Eastward thrusting from the direction of the Idaho batholith to the west,

(2) westward thrusting from the Boulder batholith to the east, and (3) northward crowding by the small Philipsburg and Royal batholiths to the south (see Mutch and McGill, 1962).

Thrusting west of the batholiths is of comparable extent but different type. Great Late Cretaceous overthrusts carry variably metamorphosed eugeosynclinal materials, intruded by stocks and small batholiths during Cretaceous(?) and Jurassic, and possibly older, orogenies, westward over similar rocks and over little-metamorphosed Upper Jurassic and Cretaceous deposits. Such overthrusts occur throughout the Klamath Mountains of northwestern California and southwestern Oregon (Irwin, 1964) and in western Idaho (Hamilton, 1963b), and may be present in northwestern Washington (Hamilton, 1963b; Misch, 1952, and Barksdale, 1960, make other interpretations). South of the Klamath Mountains, throughout the length of the Coast Ranges of California, an eastern facies of miogeosynclinal Cretaceous and Upper Jurassic strata has apparently been thrust westward over the eugeosynclinal Franciscan facies of rocks of the same age, the fault extending into the basement rocks beneath the eastern facies (Brown, 1964; Irwin, 1964). Displacements on single faults may exceed 100 km in the Coast Ranges and in Washington. Sheets of serpentine and peridotite lie along many of the faults.

These western thrust faults broke the basement rocks and tapped sources of ultramafic material, and therefore are very different from the thrusts east of the batholiths. The age of major thrusting approximately coincides with the age of major plutonism, so the two processes may have been related genetically, although it is difficult to visualize a satisfactory mechanism linking them.

There was much thrusting of sedimentary rocks at least in Nevada during late Paleozoic and early Mesozoic intervals. No batholiths of corresponding age are known in the region (Gilluly, 1965, p. 23). Great thrust faults can form without assistance from batholiths.

BATHOLITHS AND YOUNGER STRUCTURE

Batholiths, once formed, influence subsequent deformation because of their mechanical strength. The largest unbroken mountain masses in the Western States—the Sierra Nevada and the mountains of central Idaho—are carved from those parts of the batholithic belt which contain the highest proportion of granitic rocks and the smallest proportion of metamorphic rocks. The fault zone bounding the Sierra Nevada block on the east cuts obliquely through the batholithic belt, which curves northeastward from the northern Sierra and southeastward from the southern Sierra, but the fault zone approximates the eastern limit of almost continuous granitic rocks within the belt. The crystalline belt contains a higher proportion of metamorphic rocks along strike to both the northeast and southeast on the east side of the fault zone, where it is broken into numerous large and small basin-and-range blocks. Similarly Baja California and the Peninsular Ranges of southwestern California are composed of almost continuous batholithic rocks, which here form the west part of the belt; the east part, in southeastern California and western mainland Mexico, contains abundant metamorphic rocks, and is broken into many fault blocks. The large basin-and-range fault blocks striking northwestward into central Idaho abruptly lose structural relief and vanish as they intersect the Idaho batholith. Other fault blocks lie north, west, and east of the batholith, but no large ones break it (Hamilton, 1962). The Coast Range of British Columbia is another massive mountain block consisting largely of batholithic rocks.

The history of batholithic mountains indicates that the presence of batholiths markedly influences uplift and erosion. Consider the Sierra Nevada batholith, which was sufficiently raised within Late Cretaceous time that its roof rocks (whether mostly volcanic, as we suggest, or metamorphic, as most geologists would infer) were eroded away before Eocene time. Mountains were still present in early Tertiary time, for middle Eocene strata in the foothills lap upon a bedrock surface whose local relief was at least 350 m (Allen, 1929, p. 382), and very coarse Eocene and Oligocene river boulder gravels are preserved in fossil valleys throughout the northern Sierra (for example, Lindgren, 1911, pls. 4, 21). The west base of the mountain block has remained near sea level throughout Tertiary time, and marine and nonmarine deposits of various ages lap onto it. The rest of the block has apparently undergone both subsidence and uplift within the Tertiary but not profound erosion. The widespread preservation of lower Tertiary deposits in the northern part of the Sierra shows that ridge surfaces there are not far from the early Tertiary surface, although there has been subsequent uplift and tilting. Lower Tertiary deposits are lacking in the central and southern Sierra, but presumably there also middle and late Cenozoic erosion has been limited largely to incision. During middle Tertiary time the Sierra Nevada was relatively low, and Miocene floras of the northern Sierra crest region indicate an altitude much lower than the present one (Axelrod, 1957). By the end of the Pliocene, local relief in the crest region reached at least 2,000 m (Dalrymple, 1964), so its altitude must have been as high or nearly as high as it is now. Much of the faulting that outlines the present Sierra block has occurred within late Pliocene and Pleistocene time (Dalrymple, 1964), and thus may have consisted primarily of the downdropping of blocks (such as Lake Tahoe and Owens Valley) on the east rather than of the upraising of the Sierra block itself.

The late Cenozoic uplift of the Sierra Nevada must be related to the deep downbulge in the Mohorovicic discontinuity beneath it. The range is high because the root is deep. Had this root formed at the same time as the batholith, the range would long ago have reached maximum altitude by rising isostatically; instead, therefore, much of the downbulge must be a late Cenozoic development. (Christensen, 1966, reached the same conclusion.) Sierran silicic batholithic rocks have a high rate of production of radioactive heat, which increases markedly with potassium content (Wollenberg and Smith, 1964), and perhaps this property is responsible for the root growth and uplift. The highest part of the 650-km-long Sierra Nevada is the 300-km segment, between Sonora Pass and Olancha, in which quartz monzonite and silicic granodiorite form most of the eastern part of the block. The lower crest north and south of this segment, and the middle-altitude region west of the high segment of the crest, consist largely of granodiorite, and the western foothills consist mostly of quartz diorite and metamorphic rocks. Radioactive decay would retard cooling, particularly of the crestal silicic granodiorite and quartz monzonite; thus a partial barrier to conduction of heat from the mantle would be formed, and heating of the lower crust and upper mantle would result. Movement of crust or mantle materials, change in pressure-phase mineralogy or hydration state, or partial melting might have resulted from this deep heating and so enlarged the Sierra root.

Other late Mesozoic batholiths also tend to form mountain masses now standing higher than the surrounding nonbatholithic terranes, and the highest regions within the batholithic mountains tend to be those with the most silicic and potassic rocks. In the Coast Range of British Columbia, for example, the inland belt of quartz monzonite and granodiorite stands on the average about 1 km higher than the coastal belt of quartz diorite.

Regional or global events unrelated to the batholiths may of course have caused such observed relationships. It does appear, however, that batholiths are uplifted selectively two or more times, and if so they must also be eroded selectively more than neighboring terranes. Batholiths perhaps cause their own destruction. The general scarcity of great pre-Mesozoic batholiths is explicable in such terms. So are such along-strike transitions as that from the gneiss terrane of New Hampshire to the low-grade complex of Maine: the batholith that formed above the gneisses caused greater uplift and erosion there than in Maine, where the plutons were not numerous enough to have such an effect.

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NOTE

The discussion of the Idaho batholith in this report (p. C5–C6) was based in part upon the interpretation by Ross (1934) that the Casto Volcanics was intruded by the Cretaceous batholith. This interpretation was disproved during the 1966 field season, when all rocks assigned by Ross to the Casto west of the Middle Fork of the Salmon River were mapped by Frederick W. Cater, Warren Hamilton, Benjamin F. Leonard 3d, Raymond L. Parker, and Edwin V. Post. These rocks were found to be that part of the Challis Volcanics (of Eocene age as dated by Axelrod †) which has been altered or converted to hornfels by contact metamorphism by a Tertiary batholith of granite and quartz monzonite. (Ross recognized that the young batholith was Tertiary, but he did not see its contacts with the volcanic rocks and did not recognize that fresh Challis grades into the altered rocks which he called Casto.) The "Casto Volcanics" apparently does not exist in the sense intended by Ross. This finding negates the argument in this paper that the Casto might be part of the extrusive cover beneath which the Idaho batholith crystallized—but it is wholly in accord with the general concepts developed here, for the quartz latite welded tuffs of the Challis form the roof beneath which the Tertiary batholith crystallized. The tuffs presumably formed largely as the ejecta of that batholith.

† Axelrod, D. L., 1966, Potassium-argon ages of some western Tertiary floras: Am. Jour. Sci., v. 264, p. 497–506.