The Sierra Nevada Batholith A Synthesis of Recent Work Across the Central Part

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SHORTER CONTRIBUTIONS TO GENERAL GEOLOGY

GEOLOGICAL SURVEY PROFESSIONAL PAPER 414–D

Prepared in cooperation with the State of California, Division of Mines and Geology

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON : 1963

UNITED STATES DEPARTMENT OF THE INTERIOR

STEWART L. UDALL, *Secretary*

GEOLOGICAL SURVEY

Thomas B. Nolan, *Director*

For sale by the Superintendent of Documents, U.S. Government Printing Office
Washington, D.C., 20402

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ABSTRACT

Maps prepared by the Geological Survey since 1945, in conjunction with other published and unpublished maps available to us, cover more than half the area of a wide strip across the central Sierra Nevada between 36°45′ and 38° N. lat. This report is an attempt to synthesize ideas acquired by us during studies of different areas within this strip, and to relate them to some of the larger problems of the batholith. Emphasis is on the identification of problems and the suggestion of hypotheses for future testing rather than the presentation and defense of unique hypotheses.

Attitudes of stratigraphic sequences in the wallrocks and in roof pendants indicate that the batholith occupies the axial part of a complex north- to northwest-trending syclinorium. Probably the synclinorium began to form in Late Permian or Triassic time, but parallelism of Upper Jurassic strata with older strata indicate that the principal folding took place during the Late Jurassic.

The Paleozoic rocks on the east side of the batholith are miogeosynclinal and transitional sedimentary rocks, whereas those on the west side are eugeosynclinal and include volcanic as well as sedimentary rocks. The Mesozoic rocks on both sides of the batholith include eugeosynclinal sedimentary rocks and volcanic flows, tuffs, and breccias, but some sedimentary remnants within the batholith are miogeosynclinal. After or concurrent with the late stages of folding, a system of steeply dipping faults, along which the movement may have been dominantly horizontal, was formed. One or more later periods of deformation, probably contemporaneous in part with the faulting, locally produced cleavages and steep lineations.

In latest Jurassic or earliest Cretaceous time two groups of plutonic rocks were emplaced in what now is the west wall of the batholith: an older ultramafic group and a younger granitoid group. The ultramafic rocks were localized principally along shear zones, whereas some granitoid rocks cut across major shear zones. Nevertheless, both groups are locally sheared and thus were probably involved in late stages of folding and faulting.

The batholith is composed chiefly of quartz-bearing granitic rocks which range in composition from quartz diorite to alaskite, but it includes smaller scattered masses of darker and older plutonic rocks and remnants of metamorphosed sedimentary and volcanic rocks. The granitic rocks are in discrete plutons, which are either in sharp, generally steeply dipping, contact with one another or are separated by thin septa of metamorphic or mafic igneous rocks or later aplitic dikes. The batholith consists of a few very large plutons elongated northwestward, parallel with the long direction of the batholith, and many smaller plutons that lie between the larger ones.

In general, the major plutons in the western part of the batholith are older, more mafic, and of higher specific gravity than those in the eastern part; in the Yosemite region the larger plutons were emplaced as a west-to-east succession. Nevertheless, some older plutons lie along the east side of the batholith, and the sequence of intrusion is probably more correctly described asymmetric than as one sided.

Most of the granitic rocks are medium grained and hypidiomorphic-granular, but both seriate and porphyritic textures are common. The phenocrysts of porphyritic rocks commonly are K-feldspar and were formed during the later stages of crystallization. Porphyritic rocks generally are calcic quartz monzonite or sodic granodiorite, and it is believed that rocks of this compositional range may be especially favorable to the formation of phenocrysts because of characteristics in their mode of crystallization.

Mafic inclusions are abundant only in granitic rocks that contain hornblende; that is, in quartz diorite, granodiorite, and calcic quartz monzonite: and their presence very likely is determined by the stability of hornblende in the enclosing magma. Probably they represent refractory rocks that were picked up at various levels and carried upward by the granitic magma. Preferred orientation of lens-shaped mafic inclusions provides the most conspicuous expression of primary foliation, but foliation is also shown by tabular and prismatic minerals, especially biotite, hornblende, and phenocrysts of K-feldspar. Locally, elongate inclusions define lineation, but this is generally best shown by the alinement of hornblende prisms, Progressively flatter shapes of mafic inclusions toward contacts are common and are probably caused by stretching in the partly crystallized margins of plutons as a result of expansion during emplacement.

Some plutons are of about the same composition and texture throughout, but many are compositionally zoned. Zoning is either concentric or lateral. Lateral zoning is usually evident in larger plutons and is shown by progressive change in composition from one end or one side to the other. Concentrically zoned plutons are progressively more felsic inward from their margins. The cause of most zoning is believed to be differentiation during crystallization, although contamination of some granitic rocks by older mafic igneous rock is evident in places. Movements of a still-liquid core magma probably explain internal contacts in concentrically zoned plutons. The Tuolumne intrusive series of Yosemite National Park consists of the Sentinel granodiorite, Half Dome quartz monzonite, Cathedral Peak granite, and Johnson granite porphyry, and may be an example of a very large concentrically zoned pluton in which repeated movements of the core magma took place during cooling and crystallization.

Field observations and comparison of compositional trends with experimental data show that most of the granitic rocks crystallized from magma. They were forcibly emplaced, and several plutons have pushed their wallrocks aside and upward. Forcible intrusion produced broad flexures in the wallrocks around steep axes; these flexures generally can be distinguished from north- to northwest-trending horizontal or gently plunging regional folds. Lateral separations within individual pendants of as much as 3 miles are attributed to forcible intrusions of magma. A separation of 8 miles between the south end of the Mount Morrison pendant and a discontinuous septum that extends north from the Pine Creek pendant probably was caused by emplacement of a pluton of Round Valley Peak granodiorite or a younger pluton of quartz monzonite similar to the Cathedral Peak granite.

In the areas mapped thus far stoping appears to have been quantitatively unimportant, but the role of stoping has been very incompletely evaluated. Granitization and assimilation effects are conspicuous on a small scale where granitic magma came into contact with mafic metavolcanic or plutonic rocks or amphibolite, and may have been important in terranes composed largely of those rocks. The reactions involved are in accord with Bowen's reaction series as it applies to granitic magma and crystalline materials composed of minerals that form earlier in the reaction series.

INTRODUCTION

Since 1945 the U.S. Geological Survey in cooperation with the California Division of Mines and Geology has carried on a program of geologic study of the tungsten-bearing districts in the east-central part of the Sierra Nevada batholith. As part of this program seven 15-minute quadrangles have been mapped, and the mapping of several others is in progress. During the same period studies have been carried on in the metamorphic terrane of the western foothills of the Sierra Nevada, which includes the Mother Lode gold belt and the Foothills copper belt. The geologic maps prepared as part of the tungsten and western foothills studies, together with published maps by Calkins (1930, pl 51), Macdonald (1941), Krauskopf (1953), Hamilton (1956), and Sherlock and Hamilton (1958), and unpublished maps available to us cover more than half the area of a wide strip across the central part of the batholith. This strip, which lies between 36°45′ and 38°N. lat, is shown in figure 1 and on plate 1, and is referred to in this report as the central Sierra Nevada.

The purpose of this paper is to synthesize ideas acquired by each of us during detailed studies in different areas and with somewhat different emphasis, and to relate them to some of the larger problems of the batholith. It is a progress report, and almost all of the conclusions in it are tentative. At this stage we are particularly concerned with identifying problems and suggesting possible alternative solutions for future testing. Agreement among us is unanimous on most, but not all, points; where opinions differ, the conflicting views are generally presented together. We are more familiar with the eastern than with the western part of the batholith, and final judgment of the hypotheses and tentative conclusions advanced here must await more work in the western part. Because the emphasis is on the batholith, the discussion of the prebatholithic history of the region is brief; if some statements appear dogmatic in their brevity, they are not so intended.

In preparing this report we have each assumed responsibility for parts that are closest to our experience. Clark has studied the metamorphic belt in the western foothills, and is responsible for the parts dealing with the rocks of that area. Rinehart and Huber have studied areas on the east side of the batholith that include large roof pendants, and they are chiefly responsible for the discussion of the rocks in pendants. Bateman and Moore are chiefly responsible for the parts dealing with the batholith itself.

GENERAL GEOLOGIC RELATIONS

The Sierra Nevada is almost 400 miles long and 60 to 80 miles wide, and extends along the eastern half of California between the Mojave Desert on the south and the Modoc Plateau on the north. It is a physiographic feature that cuts obliquely across the Sierra Nevada batholith and earlier structural trends. Structurally it is a westerly tilted Cenozoic fault block that has been locally modified by internal Cenozoic faults, especially at the north end. It consists largely of pre-Cenozoic granitic and metamorphic rocks, which are overlapped on the west by the Tertiary deposits of the Great Valley and are overlain on the north by Tertiary and Quaternary unmetamorphosed volcanic and sedimentary rocks.

The Sierra Nevada batholith, on the other hand, is part of a more or less continuous belt of plutonic rocks that extends from Lower California on the south, northward through the Peninsular Ranges and Mojave Desert, through the Sierra Nevada, and into western and northwestern Nevada. A single name, such as Cordilleran batholith, could be applied to the entire batholith, but different parts are customarily referred to by different names. In this report, the part of the belt of plutonic rocks that lies within the Sierra Nevada is considered to be the Sierra Nevada batholith. By this definition the batholith occupies all of the range except the belt underlain by metamorphic rocks and isolated plutons along the west side of the north half (fig. 1).

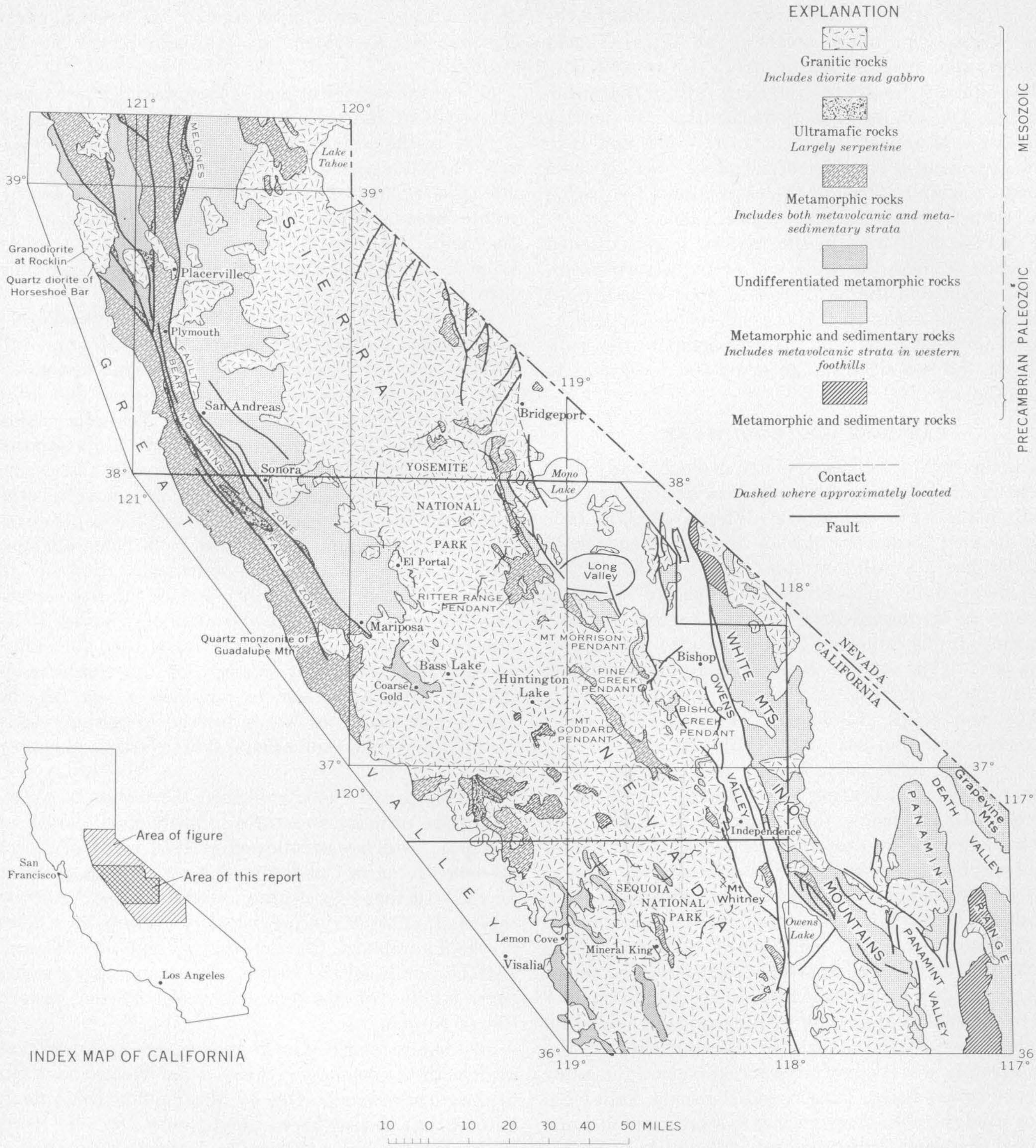

FIGURE 1.—Generalized geologic map showing distribution of pre-Cenozoic rocks in part of eastern California and location of area discussed in this report.

Exposures of the wallrocks on the west side of the batholith are continuous north of lat 37°15′ N., but farther south wallrock and roofrock are present only as sporadically distributed remnants within the batholith (fig. 1). On the east, prebatholithic sedimentary rocks are continuously exposed in the White and Inyo Mountains, and a belt of discontinuous wall and roof remnants extends through the east side of the batholith between Owens Lake and Lake Tahoe.

The area dealt with in this report is well situated because it includes the south end of the western metamorphic belt and the north end of good exposures of prebatholithic rocks on the east side of the batholith; thus it provides a suitable place to compare the wallrocks on the two sides and to study their relations to the batholith.

PREVIOUS GEOLOGIC WORK

The Sierra Nevada has attracted geologic study since the early 1860's, when the Geological Survey of California made its investigations (Whitney, 1865). It is only recently, however, that an area large enough to be considered a valid sample of the batholith has been mapped in sufficient detail to permit a serious attempt to synthesize its history and gross geologic structure. In the following brief summary of the literature, most of the reports cited deal with areas included within the boundaries of the map on plate 1.

The early folios (Lindgren, 1894, 1896, 1897, 1900; Lindgren and Turner, 1894, 1895; Ransome, 1900; Turner, 1894, 1897, 1898; Turner and Ransome, 1897, 1898) of the gold-bearing areas along the west side of the batholith show the lithologic distribution of the metamorphic and plutonic rocks with considerable accuracy, but few aspects of the stratigraphic succession and structure were worked out. This may seem curious to some, but not to those who have attempted to determine even the simplest of relations in the strongly deformed and sparsely fossiliferous strata of this region. It was not until 1943, when Taliaferro published the results of a study of the manganese deposits of the Sierra Nevada, that the interpretation of the stratigraphy was successfully carried beyond the point reached by Lindgren, Turner, and Ransome. Taliaferro recognized the volcanic sequences as stratigraphic units, showed that some beds previously thought to be Paleozoic are of Jurassic age, and worked out interpretations of broader structures based in part on determinations of top directions. Knopf (1929) recognized a great throughgoing fault zone in the Mother Lode belt, and Ferguson and Gannett (1932) mapped this fault zone to the north through the Alleghany district. Knopf (1929, p. 18–19) also made the suggestion that some granitic and dioritic intrusives in the western metamorphic belt are older than the main Sierra Nevada batholith.

One of the earliest attempts to represent the separate intrusives of the batholith on a map was by Knopf (1918) in the eastern Sierra Nevada. In a single season (1913), assisted by F. H. Lahee, he mapped a 100-mile segment of the eastern Sierra Nevada escarpment. Because of the small amount of time spent in the field, the map is oversimplified, but it clearly pointed the way for future mapping within the batholith.

Even earlier, probably prior to 1910, H. W. Turner had mapped parts of the Yosemite and Mount Lyell 30-minute quadrangles, which together span the batholith. Turner's maps were never published, but they served as the base for a map of the Yosemite region prepared by Calkins between 1913 and 1915 (Calkins, 1930, pl. 51). Calkins' map shows clearly and accurately the separate intrusions which constitute the batholith; the accompanying text of nine pages gives their relative ages, as determined from field relations. Calkins' report established a solid foundation for an understanding of the batholith and remains the single most important step toward an understanding of its history and structure. It served as a base for Ernst Cloos' (1936) well-known study of the structure of the batholith, and recently provided a test for the age determination of rocks by the potassium-argon method (Evernden and others, 1957; Curtis and others, 1958).

In 1932, Ernst Cloos published the results of a survey of the primary structures in granodiorite south of Mariposa, which was followed in 1936 by "Der Sierra Nevada Pluton in Californien," his well-known report on the structure of the granitic rocks in the Yosemite region. In 1934 Mayo published the results of studies in the Laurel and Convict basins, and in 1937 and 1941 he published the results of reconnaissance structural studies of the granitic rocks of the eastern Sierra Nevada.

Just before World War II, two reports appeared that dealt with the geology of areas in the western foothills of the Sierra Nevada. One by Macdonald (1941) deals with an area east of Fresno, and one by Durrell (1940) deals with an area a few miles farther south, outside the area of the present report. These reports are more concerned with the structure and metamorphism of the metamorphic rocks than with the constitution of the batholith, but both reports include maps on which some of the different granitic rocks are distinguished.

During World War II, geologists of the Geological Survey investigated most of the tungsten deposits in the Sierra Nevada. In connection with these investigations Krauskopf (1953) mapped a large area in the west-central part of the batholith. Most of the studies made since World War II are an outgrowth of the wartime studies.

ACKNOWLEDGMENTS

We are deeply indebted to our associates within the Geological Survey for frequent helpful informal critiques, and also to many others outside the Survey, particularly to members of the departments of geology at the University of California and at Stanford University, with whom we have had many fruitful discussions. The California Division of Mines, in addition to financial support, made available its file of unpublished maps. D. B. Stewart assisted with the parts of the report dealing with experimental work on melts of granitic composition.

WALLROCKS AND ROOF ROCKS

The walls of the batholith and roof remnants within the batholith are composed chiefly of metamorphosed sedimentary and volcanic rocks of Paleozoic and Mesozoic age. The east contact of the batholith is hidden beneath surficial deposits of Cenozoic age, but the composition, structure, and age of the wallrocks are shown in exposures in the White and Inyo Mountains, and in a narrow belt of roof pendants that extends southeast from Bridgeport to Independence. The west contact of the batholith is exposed for about 200 miles in the northern Sierra Nevada, but is concealed in the southern part. The Paleozoic rocks indicate a transition from a miogeosynclinal environment east of the batholith to a eugeosynclinal environment west of the batholith. Most of the Mesozoic rocks indicate a eugeosynclinal environment on both sides of the batholith, but a belt of roof pendants that extends southwest from Huntington Lake indicates a miogeosynclinal environment. The relations of the rocks of these two environments to one another is uncertain, but there is a good possibility that the eugeosynclinal rocks are Jurassic and the miogeosynclinal rocks Triassic.

PALEOZOIC ROCKS

All Paleozoic systems are represented in the White and Inyo Mountains, where the stratigraphic section totals about 36,000 feet. The lowest 13,000 feet of strata are unfossiliferous and may be Precambrian, but structurally are essentially conformable with the overlying fossiliferous strata, and are here included as part of the Paleozoic sequence.

Paleozoic rocks in the roof pendants of the eastern belt contain much less carbonate rock than those of the White and Inyo Mountains. They consist largely of fine-grained, thin-bedded siliceous hornfels derived from siltstone, mudstone, and shale. Interbedded with the siliceous hornfels are subordinate amounts of (a) metamorphosed limestone, (b) orthoquartzite that is in part calcareous, (c) chert, and (d) calcareous or dolomitic siltstone. The most complete stratigraphic section is in the Mount Morrison pendant and is more than 30,000 feet thick (Rinehart and others, 1959). The lower part of the section contains fossils of Early to Middle or Late Ordovician age and consists of 14,700 feet of alternating thin-bedded silicious and pelitic hornfels, metachert, calcareous orthoquartzite, and marble. The middle part is unfossiliferous and consists of 10,500 feet of alternating siliceous hornfels and calcareous orthoquartzite. The upper part contains fossils of Pennsylvanian and Permian(?) ages and consists of about 7,200 feet of siliceous hornfels and some limestone.

In the west wall of the batholith, within the area shown on plate 1, rocks of Paleozoic age are confined to the area between the batholith and the Melones fault zone, and are all included in the Calaveras formation. Turner (1893, p. 309) reported finding *Fusulina*, which he considered to be of Carboniferous age, near El Portal, but to our knowledge no other Paleozoic fossils have been reported from the part of the western metamorphic belt within the map area (pl. 1). In the northwestern Sierra Nevada fossils ranging in age from probable Devonian to Permian have been found in a section that is generally similar to that near the Merced River except that it contains more mafic volcanic rocks, and a thick section of sandstone or felsic tuff that is at least in part of Silurian age.

Near the Merced River the lower part of the section, about 10,000 feet thick, consists chiefly of dark-gray phyllite and siltstone, but contains some graywacke, chert, and pyroclastic rocks. This sequence is overlain by a unit about 5,000 feet thick composed of mafic pyroclastic rocks. Next overlying is a unit more than 5,000 feet thick composed predominantly of dark-gray siltstone and phyllite, but which includes some thin-bedded chert, scattered lenses of limestone, and minor intraformational conglomerate. The youngest unit, also more than 5,000 feet thick, is composed mostly of thinly interbedded chert, very fine grained quartzite, and dark-gray phyllite; it contains small limestone lenses throughout and a large limestone body in the north part of the area (pl. 1). Near the east edge of the metamorphic belt are a few hundred feet of quartz sandstone and arkose interlayered with mafic volcanic rocks; the stratigraphic relations of these strata are unknown, and they are arbitrarily assigned to the phyllite-chert sequence.

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MESOZOIC ROCKS

Rocks of Mesozoic age crop out in two nearly parallel northwest-trending belts: one in the west side of the western metamorphic belt and one in the west side of the belt of roof pendants that extends through the east side of the batholith. In addition a group of pendants extending southeast from Huntington Lake through the central part of the batholith is composed of strata of probable Mesozoic age. The Mesozoic strata in the belt of pendants through the east side of the batholith consist chiefly of metavolcanic rocks and graywacke-type sediments derived chiefly from volcanic rocks. These rocks weather gray and contrast strongly with the nearby Paleozoic hornfelsic metasediments, which weather reddish brown. In the western metamorphic belt both the Paleozoic and Mesozoic rocks contain volcanics, and the differences between the rocks of the two eras are much less conspicuous. The Mesozoic rocks in the pendants southeast of Huntington Lake resemble lower Paleozoic strata along the east side of the batholith and are unlike Mesozoic strata elsewhere in the central Sierra Nevada.

In the Mesozoic rocks of the eastern belt of roof pendants, pyroclastic rocks of felsic to intermediate composition are the most common types and are interlayered with less common mafic flows and hypabyssal intrusives. Thin beds of epiclastic rocks are sporadically scattered throughout the metavolcanic sequence. The thickest section of Mesozoic strata in the eastern belt is exposed in the Ritter Range roof pandant (pl. 1), where about 30,000 feet of metamorphosed pyroclastic rocks of intermediate to felsic composition stratigraphically overlie rocks of Paleozoic age. Marine fossils of Early Jurassic age occur in a thin sedimentary unit 10,000 feet above the base of the Mesozoic section, and provide the only age data in the pendant. Marine fossils were also found 5,000 feet higher in the section but are too poorly preserved for their age to be determined. (Rinehart and others, 1959).

The eastern belt of Mesozoic rocks extends discontinuously southeastward to the Inyo Mountains northeast of Owens Lake, where about 1,800 feet of fossiliferous marine limestone and shale of Early and Middle Triassic age are overlain (possibly unconformably) by 4,200 feet of unfossiliferous interbedded volcanic and continental sedimentary strata (Merriam, 1963).

Metavolcanic rocks of the western belt are chiefly andesite but include rocks that range in composition from rhyolite to basalt and constitute about three-fourths of the Mesozoic section; most are pyroclastic, but lava, showing pillow structure in places, is interlayered throughout all the metavolcanic formations. Most of the metasedimentary rocks were derived from siltstone and shale, but metamorphosed graywacke and conglomerate occur in all formations and are widespread geographically. The conglomerates contain pebbles of chert, vein quartz, felsite tuff, orthoquartzite, diorite, coarse quartz-mica schist, and undeformed Permian limestone. Similar chert, felsite tuff, and quartzite form parts of the Paleozoic section east of the Melones fault zone in the Sierra Nevada, and may also constitute parts of the concealed basement of the Great Valley of California.

The remnants of metasedimentary rocks extending southeast from near Huntington Lake consist chiefly of crossbedded feldspathic quartzite, micaceous schist, and subordinate marble, calc-hornfels, and metavolcanic rocks. Although these rocks lithologically resemble the Paleozoic sections in the eastern belt of roof pendants more closely than definitely dated Mesozoic sections, they have tentatively been assigned a Mesozoic age on the basis of fossils recently discovered near the confluence of the Middle and South Forks of the Kings River (Moore and Dodge, 1962). Fossils of Late Triassic age previously reported from the Mineral King pendant south of the map area (Durrell, 1940, p. 17) further support this designation. The possibility that these rocks are Triassic and that the strata in the predominantly volcanic belts on the two sides of the batholith are Jurassic merits serious consideration.

STRUCTURE

The batholith probably occupies the axial region of a faulted synclinorium. The existence of this synclinorium is shown by the predominence of opposing, inward-facing top directions in the strata on the two sides of the batholith. The synclinorium is not readily apparent in the map pattern (pl. 1) chiefly because strike faults of large displacement interrupt the gross sequence of strata in the western metamorphic belt. The axis of the synclinorium trends about N. 40° W. in the central Sierra Nevada, but trends northward in the northern Sierra Nevada. Axes of major folds in both walls and in roof pendants generally plunge northwest and southeast at angles of less than 30°. Axial planes of these folds generally dip steeply.

The eastern limit of the synclinorium is probably marked by a belt of Precambrian and Cambrian rocks that extends from the White Mountains southeastward into the Death Valley region. The western limit presumably is beneath the Cretaceous and Tertiary strata of the Great Valley (fig. 1).

From the belt of older rocks that extends from the White Mountains to Death Valley, the strata in the east wall are successively younger westward. The range-front faults that bound Owens Valley strike obliquely across the major structures in the Paleozoic and Mesozoic strata, and as a consequence successively younger strata are exposed toward the south on the western flanks of the White and Inyo Mountains. The strata in the White and Inyo Mountains and in many of the metamorphic remnants within the batholith are strongly folded and faulted, causing many repetitions of formations. In the Mount Morrison and Ritter Range pendants, however, the gross structure is homoclinal across many thousands of feet of strata ranging in age from Ordovician to Jurassic, although minor folds within formations are common (Rinehart and others, 1959). Folds in the western part of the Ritter Range pendant may be related to the axial region of the synclinorium.

In the western metamorphic belt the gross distribution of strata resulting from development of the synclinorium has been reversed by movement along steeply dipping fault zones, and Paleozoic strata lie east of Mesozoic strata (Clark, 1960a). The internal structure of individual fault blocks is, in general, homoclinal, and most tops are to the east; beds dip eastward more than 60°. The homoclinal structure is interrupted in parts of the belt by both isoclinal and open folds, but even in such places the older strata in a fault block generally are exposed near its west side and the younger strata near its east side.

Cleavage is common, especially in the western metamorphic belt and in metamorphic remnants of Mesozoic rock, but is also present locally in the White and Inyo Mountains. Most cleavage is approximately parallel to the axial planes of folds, but bedding plane schistosity is present locally in the Paleozioc rocks of the west wall; in the White Mountains a second cleavage locally parallels thrust faults. In general, only one cleavage, which parallels the axial planes of the major folds, was recognized in the White Mountains and in the eastern pendants. Farther west other cleavages related to younger, generally steeply plunging, folds are also present. In the Goddard pendant steeply dipping cleavage parallel to the axial planes of steeply plunging folds cuts across vertical or slightly overturned beds, and in the pendant 8 miles east of Shaver Lake several cleavages and related folds appear to be present. In the western meamorphic belt one or more cleavages can be identified in most outcrops, but in many places the movements that led to the development of the last cleavage destroyed the earlier ones.

Accompanying the cleavages are many linear elements such as minor fold axes; flattened elongate fragments of cataclastic, pyroclastic and epiclastic origin; elongate pods of chlorite and mica; parallel amphibole crystals; intersections of bedding and cleavage; and intersections of two cleavages. Generally, all of these elements are parallel to minor folds, and in places where two or more sets of minor folds are present the other linear elements also are in two or more orientations.

Steeply plunging cataclastic linations are found in granitic as well as in metamorphic rocks. Some lineations in granitic rocks undoubtedly were caused by regional deformation, but others may have resulted from protoclasis. Vertical lineations marked by parallel amphibole prisms in the sheared western part of the pluton that abuts the Melones fault zone at Plymouth (fig. 1), and steep lineations in some older sheared plutons within the batholith (for example, plutons intruded into the Goddard pendant) undoubtedly were caused by regional deformation, because the lineated zones continue into the wallrocks.

On the other hand, lineations in granitic rocks south of Kerchoff Lake and at Courtright Reservoir could have been caused either by protoclasis or as a result of regional deformation. The zone south of Kerchoff Lake lies approximately on the projected trend of the Melones fault zone, and the Courtright Lake zone coincides approximately with the boundary between Mesozoic pendants of dominantly volcanic lithology to the east and ones of dominantly sedimentary lithology to the west. At both Kerchoff Lake and Courtright Reservoir the lineated zones are cut by unlineated younger granitic intrusives, showing that the lineations were produced during the general period of magma emplacement.

The major fault zones in the western metamorphic belt strike nearly parallel with the regional strike of the metamorphic strata, and dip from 70° eastward to vertical. The Melones fault zone is continuous throughout the metamorphic belt, and other fault zones form a continuous braided system from lat 37° 30′ to lat 40° N. The Melones fault zone appears to have been the zone of greatest movement, for along it Mesozoic rocks are faulted in a stratigraphic position beneath Paleozoic rocks for a distance of more than 90 miles. The stratigraphic separation along this fault zone doubtless is measurable in miles, although the thickness of strata involved are not known well enough to give a precise figure. Because the angle between the fault zone and bedding, both in strike and dip, is small, the net separation along the fault zone must be several times greater than the stratigraphic separation. The stratigraphic separation along the Bear Mountains fault zone southwest of Plymouth exceeds 3 miles, and the net slip along the fault zone must also be much greater.

Dominant strike-slip movement along these faults is suggested by steep linear elements within the fault zones, which have been interpreted to be b-lineations (Clark 1960b, p. 491–493; 1961), but there is as yet no direct evidence of the direction or amount of net slip. Alternative interpretations are that the major movement along these faults was dip-slip reverse (Knopf, 1929, p. 45–46), or that the faults are low-angle thrusts along which movement began early in the development of the synclinorium, and continued as they were folded along with the adjacent strata to their present steeply dipping positions. Both of the latter interpretations require that the steep lineations be superimposed on the fault zones during a later stage of deformation not related to that during which the chief movements took place.

EVOLUTION OF THE GROSS STRUCTURE

Although the faulted synclinorium into which the batholith is intruded probably did not begin to take form before late Permian or Triassic time, and the period of principal folding was Late Jurassic, the synclinorium is apparently localized along a transitional zone in the Paleozoic strata, which separates unlike facies of different depositional and tectonic histories. In the Inyo Mountains and for hundreds of miles eastward, the Paleozoic rocks are miogeosynclinal sedimentary rocks, whereas those west of the transitional zone are eugeosynclinal sedimentary and volcanic rocks. Subsidence during the Paleozoic was greater where the Paleozoic rocks of the roof pendants accumulated than farther east, for the partial section of transitional Paleozoic rocks (Early Ordovician to Permian?) in the roof pendants is thicker by more than 10,000 feet than a correlative section of miogeosynclinal rocks in the Inyo and White Mountains. Whether subsidence was still greater where the Paleozoic rocks of the western wall accumulated is not established, but seems likely.

The Paleozoic rocks are, in general, more deformed than the Mesozoic rocks and, according to Knopf (1929, p. 9), metamorphosed differently. In the pendants of the eastern Sierra Nevada the lower Paleozoic strata and fold axes appear to trend more northward than the upper Paleozoic strata and folds. Possibly the more northerly trends in the lower Paleozoic mark the middle-Paleozoic Antler orogeny recognized in Nevada (Roberts and others, 1958). A projection of the Antler orogenic belt to the southwest would intersect the Sierra Nevada near Bishop.

The paucity of Triassic fossils and the general absence of Middle Jurassic fossils in the central Sierra Nevada makes it difficult to synthesize a regional tectonic picture for the Mesozoic prior to the middle Late Jurassic (Kimeridgian) except in a broad and speculative way. Evidence of local uplift in early Mesozoic time is found in both the east and west walls. In the west wall, uplift at this time is suggested by widely distributed rounded pebbles of Paleozoic rocks in Mesozoic conglomerates, and, in the northern Sierra Nevada, by an angular unconformity between probable Silurian and Upper Triassic(?) rocks (Clark and others, 1962, p. B18–B19). In the eastern belt of roof pendants conglomerates composed of pebbles of Paleozoic rocks are present at several places along the contact of Paleozoic metasedimentary rocks with Mesozoic metavolcanic rocks. If the belt of pendants extending southeast from Huntington Lake are Triassic, the Triassic may be missing along the east side of the batholith except in the southern part of the Inyo Range where sedimentary strata of Early and early Middle Triassic age underlie (probably unconformably) unfossiliferous volcanic strata. Although the rocks of the western wall record volcanism through much of Paleozoic time, extensive volcanism probably did not begin in the area of the eastern belt of roof pendants until Middle Triassic time at the earliest and possibly not until Early Jurassic time.

Accelerated subsidence during the Mesozoic is indicated by the compositions and textures of the rocks and by comparison of the thicknesses of Mesozoic sections with thicknesses of Paleozoic sections. Although the thicknesses of partial Mesozoic sections are no greater than the Paleozoic sections, they accumulated in shorter intervals of time. The time relation of subsidence to deposition has not been determined, but the two are believed to have been coincident.

Although folding and subsidence took place during the Paleozoic and early Mesozoic, steep eastward dips of Late Jurassic rocks and faulting of Paleozoic rocks over Late Jurassic rocks in the western metamorphic belt indicate that most of the folding related to the present form of the synclinorium took place during the later part of the Late Jurassic. The youngest folded and faulted beds that have been dated are of middle Jurassic (early Kimeridgian) age. During or after the major episode of folding, the block of Paleozoic rocks in the east side of the western metamorphic belt was faulted into place. Both the folds and some major faults in the western metamorphic belt are truncated by the granodiorite at Rocklin, the quartz diorite of Horseshoe Bar, and the quartz monzonite of Guadalupe Mountain (fig. 1), which are of latest Jurassic or earliest Cretaceous age according to determinations by the potassium-argon method; but the Melones fault zone truncates an undated pluton north of Plymouth.

One or more deformations occurred during and after the emplacement of some plutons in the western metamorphic belt and probably also the large mass of hornblende-biotite grandiorite of "Dinkey Creek" type south of Huntington Lake but before most plutons in the east half of the batholith were emplaced. Steep cataclastic foliations and lineations found locally in several of the older granitic rocks show that they were involved in one or more later periods of deformation. Doubtless many of the steeply dipping systems of minor folds and associated cleavages were formed at this time; continued or renewed movement probably also took place along the major faults in the western metamorphic belt.

METAMORPHISM

Metamorphism in the Sierra Nevada can be conveniently considered in terms of three successive episodes characterized by three conditions: the first, kinematic; the second, overlapping kinematic and thermal; and the third, thermal. The first episode was related to the principal period of regional folding and resulted in the formation of slates and phyllite, but not rocks of high thermal grade. The second episode was related to synkinematic intrusion of the early plutonic rocks during a time when regional deformation continued or was renewed. During this episode many minor steeply plunging folds, secondary cleavages, and steep lineations were formed; adjacent to intrusive magmas, doubtless directional fabrics were impressed upon rocks of high thermal grade. The last episode resulted in the development of hornfelses by postkinematic intrusions. The effects of the first and third episodes can be readily distinguished, but in hornfelsed rocks it is difficult to distinguish the effects of the second episode from those of the third, which are superimposed on those of the first.

At the present time our knowledge warrants further discussion of the metamorphism only in the belt of discontinuous roof pendants along the eastern margin of the batholith. Most pendants exhibit elements of both kinematic and thermal metamorphism, but the effects of thermal metamorphism are dominant almost everywhere. The most common metamorphic texture is granoblastic, and is best shown by siliceous and impure calcareous rocks close to plutonic contacts. Directional fabric, chiefly planar but locally linear, is common in pelitic rocks and in tuffs, but the degree of development is varied. Phyllites, schists, and gneisses are generally found only in smaller pendants and in thin septa.

Because thermal effects are more conspicuous than kinematic effects in the metamorphic rocks, generally the metamorphic grade is best expressed in terms of Turner's (*in* Fyfe and others, 1958, p. 199–239) facies of contact metamorphism, but in some his regional facies, which he believes reflect higher H_2O pressures, seem to apply. Most mineral assemblages are compatible with his hornblende hornfels facies; both amphibole and intermediate to calcic plagioclase are stable in most rocks. Other of Turner's facies represented include the pyroxene hornfels and possibly the albite-epidote hornfels facies of contact metamorphism and the almandine amphibolite and greenschist facies of regional metamorphism. According to Turner the mineral assemblages in his albite-epidote hornfels facies are identical with those of his greenschist facies, and recognition of the albite-epidote hornfels facies is dependent on association in the field with the hornblende hornfels facies and with intrusive rocks.

The metamorphic grade is probably shown best in the impure carbonate rocks. Bowen has distinguished 13 steps in the progressive thermal metamorphism of siliceous dolomite (1940, p. 225–274). The steps are marked by the upper limits of stability of various mineral assemblages as follows:

Stable below step 1	Dolomite and quartz
2	Dolomite and tremolite
3	Calcite, tremolite, and quartz
4	Calcite and tremolite
5	Dolomite
6	Calcite and quartz
7	Calcite, forsterite, and diopside
8	Calcite and diopside
9	Calcite and forsterite
10	Calcite and wollastonite
11	Calcite and akermanite
12	Spurrite and wollastonite
13	Spurrite and akermanite

Recently, Harker and Tuttle (1956, p. 239–256) have shown that at pressures less than 40,000 pounds per square inch (equivalent to the weight of about 35,000 ft. of rock), and probably at any pressure of CO_2, steps 5 and 6 should be in reverse order; wollastonite will form from quartz and calcite at a lower temperature than calcite and periclase will for from dolomite.

It appears that step 3 was attained everywhere in the pendants, step 4 was reached in most places, steps 5 and 6 were reached locally, and step 8 was nowhere attained. Forsterite was not found among the metamorphic minerals; hence, it is not known whether step 7 was attained. Both diopside and tremolite (commonly actinolite) are present in rocks of appropriate composition, but whereas diopside is common in assemblages that include calcite, tremolite has been found rarely in assemblages that contain calcite and has not been observed in assemblages that contain both calcite and quartz, which indicates that Bowen's step 3 was attained, but step 8 was not. Dolomite was not identified in any of the metamorphic remnants, although its former existence is shown by diopside, tremolite, and, in a few places, brucite after periclase. The presence of brucite indicates that step 5 (step 6 according to Harker and Tuttle) was exceeded, but it canot be assumed that it was attained everywhere, because the paucity of brucite and the abundance of diopside and tremolite show that clean dolomite, necessary to record the step, was scarce. Bowen's step 6 (step 5 according to Harker and Tuttle), the reaction of calcite and quartz to form wollastonite, was observed at many localities, particularly along the margins of pendants or along fissures that provided avenues for the escape of CO_2. Step 6, however, was not attained everywhere, for quartz and calcite occur together in the interiors of some pendants.

Turner (*in* Fyfe and others, 1958, p. 208) includes the assemblages formed above steps 4 and 6 in his hornblende hornfels facies. However, he considers periclase, formed above step 5 (step 6 according to Harker and Tuttle), to belong to his pyroxene hornfels facies.

Among the pelitic and quartzofeldspathic rocks, quartz, andalusite, muscovite, biotite and K–feldspar are common and occur in several combinations. Almandine was found in schistose rocks at a few localities where directional fabric is especially conspicuous. In general, the mineral assemblages in these schistose rocks also are typical of Turner's hornblende hornfels facies, but almandine belongs to his almandine amphibolite facies. The assemblage andalusite-K-feldspar, which Turner (*in* Fyfe and others, 1958, p. 206) considered indicative of the pyroxene hornfels facies, is common in rocks that also contain hornblende. This association led Rose (1958, p. 1703) to suggest that the lower limit of stability of the assemblage andalusite-K-Feldspar is below the upper limit of the hornblende hornfels facies. Occurrence of andalusite and K-feldspar together does indicate, however, that the associated rocks are in the upper rather than the lower part of the hornblende hornfels facies. Sillimanite in association with cordierite or with andalusite, observed at several localities west of Independence, also indicates that in places the grade of metamorphism is near the upper limit of the hornblende hornfels facies or transitional to the pyroxene hornfels facies.

Data from the mafic metavolcanic rocks are less definitive than from the impure carbonates or from pelitic rocks, but the common presence of amphibole and intermediate plagioclase shows that most of these metavolcanic rocks are in the hornblende hornfels facies. The amphibole in these rocks generally occurs as deep-green equant crystals that appear to be pseudomorphs of pyroxene, thick prisms with ragged terminations, or small scattered grains. Partial alteration to biotite is common, and exsolved magnetite occurs in and marginal to some crystals. The amphibole appears to be common hornblende; it is darker green and has a much larger extinction angle than the fibrous actinolite of calc-silicate assemblages. Plagioclase commonly ranges in composition from oligoclase to calcic andesine, and a few flows contain relict plagioclase crystals that exhibit oscillatory zoning and grade from a labradorite core to an andesine rim. Most plagioclase crystals, however, are unzoned, and the composition is about the average composition of the relict zoned crystals. Concretions composed chiefly of intermediate to calcic plagioclase and epidote are common in the mafic metavolcanic rock; this assemblage is considered to indicate the almandine amphibolite facies. In a few rocks the plagioclase associated with epidote is albite, indicating the albite-epidote or greenschist facies.

The degree of recrystallization is not everywhere the same; in general, it is roughly correlative with the metamorphic grade, and is greater in rocks of higher grade and less in rocks of lower grade. In small remnants and in the margins of large ones recrystallization is more advanced than in the interiors of large remnants. In general, all the rocks in remnants less than a mile across are completely recrystallized, whereas in larger remnants only the outermost margins and narrow zones parallel to fissures are as strongly metamorphosed. Hence, the average grain size in small metamorphic remnants is generally in the range of 0.1 to 0.5 mm, whereas the grain size in the interiors of large remnants is generally a tenth as large. In large remnants the granoblastic texture is less well developed away from the margins, and in the interior may be subordinate to primary textures.

Although several metamorphic facies are represented, the ranges of temperature and pressure during metamorphism were probably not great. Turner (*in* Fyfe and others, 1958, fig. 107, p. 237) has constructed a schematic representation of his metamorphic facies showing their probable ranges of temperature and H_2O pressure during formation. Turner makes clear that the diagram is highly speculative, but that some such relations must exist between the facies. Even though the temperatures and H_2O pressures shown may be off by considerable amounts, they are probably of about the right magnitude. On this diagram a field representing the Sierran metamorphic rocks in remnants would correspond to H_2O pressures close to 6,000 bars and temperatures ranging from about 400° to 600° C. The H_2O pressure is in good agreement with our estimate of 5,000 bars H_2O pressure in the granitic magmas during the later stages of differentiation (p. 103). If P_{H_2O} is equated with P_{load}, the depth of burial during metamorphism would have been about 11 miles. At 6,000 bars of CO_2 pressure, wollastonite would be stable only above approximately 840°C, according to Turner's diagram. Doubtless the explanation for this seeming anomoly is that wollastonite formed at CO_2 pressures considerably lower than the H_2O or load pressures; wollastonite is most common along intrusive contacts and along fractures in metamorphic rocks, which could have acted as avenues of escape for CO_2.

AGE OF THE BATHOLITH

The Sierra Nevada batholith has been variously referred to the Jurassic or the Cretaceous; the geologic relations provide no basis for discrimination. In the western foothills granitic rocks intrude Upper Jurassic strata (Mariposa slate), and the metamorphic terrane which the granitic rocks intrude is unconformably overlain by Upper Cretaceous strata (Chico formation). In the Devils Postpile area in the eastern part of the batholith, granitic rocks intrude fossiliferous Lower Jurassic strata (Rinehart and others, 1959), and in the Inyo Mountains granitic rocks intrude Middle Triassic strata.

Hinds (1934) demonstrated that the Shasta Bally batholith, a granitic mass near Redding, Calif., which lies along the projected trend of the Sierra Nevada but is separated from the Sierra Nevada batholith by many miles of younger volcanic and alluvial deposits, is of Late Jurassic age, and suggested that the Sierra Nevada batholith is of the same age. Although this suggestion was accepted by many, the reported presence in Lower California of fossiliferous Early Cretaceous and early Late Cretaceous strata in a terrane intruded by granitic stocks and batholiths (Woodward and Harriss, 1938, p. 1330; and Bose and Wittich, 1913, p. 394) raised some doubt.

Zircons from eight plutons in the eastern Sierra Nevada near Bishop have been determined by E. S. Larsen, Jr., and others (1958, table 9, p. 52), to have lead-alpha ages ranging from 88 to 116 million years (table 1). The lead-alpha ages show little relation to the sequence of intrusion as observed in the field. Larsen and others (1958, p. 57) believed that the error inherent in the lead-alpha method is somewhat more than 10 percent within the probable age range of the batholith, and this is not accurate enough to permit comparing the age of one pluton with another.

TABLE 1.—*Lead-alpha ages of batholithic rocks from the eastern Sierra Nevada near Bishop, California*

[After Larsen and others (1958, table 9, p. 52)]

Intrusive rock	Mass	Location	Probable order of emplacement	Mineral	Activity (mg/hr)	Lead (ppm)	Calculated age (10^6 years)
Rocks similar to the Cathedral Peak granite.	Mount Alice	Big Pine quadrangle. SW¼ sec. 26, T. 9 S., R. 32 E. At end of Big Pine Creek Road.	6	Zircon	618	26	105
Tungsten Hills quartz monzonite.	Morgan Creek.	Mount Tom quadrangle. West of surface workings Pine Creek mine.	5	do	796	37	116
Do	Bishop Creek	Mount Goddard quadrangle. NE¼ sec. 20, T. 8 S., R. 31 E. Along Bishop Creek Road.	5	do	792	35	110
Round Valley Peak granodiorite.	Round Valley Peak.	Mount Tom quadrangle. A quarter of a mile northeast of Rock Creek Lake.	4	do	396	12, 13, 14, 16, [1] (13.8)	88
Lamarck granodiorite	Lamarck	Mount Goddard quadrangle. NW¼ sec. 14, T. 9 S., R. 31 E. Northeast side of South Lake.	4	do	400	15	93 } [1] 91
				Thorite	4,670	205	88
Grandodiorite of McMurry Meadows.		Big Pine quadrangle. NE¼ sec. 9, T. 10 S., R. 33 E.	3	Monazite	4,897	234, 238, [1] (236)	100
Tinemaha granodiorite		Big Pine quadrangle. NW¼ sec. 14, T. 10 S., R. 33 E.	2	Zircon	331	15, 16, [1] (15.5)	116
Inconsolable granodiorite.		Big Pine quadrangle. SE. corner sec. 33, T. 9 S., R. 33 E. A quarter of a mile south of Third Lake.	1	do	221	10	112

[1] Average.

More recently the isotopic ages of biotite from a group of intrusives in Yosemite National Park and from four plutons in the metamorphic belt along the west side of the batholith have been determined by the potassium-argon method (Curtis and others, 1958). The calculated ages of the Yosemite biotites range from 82.4 to 95.3 million years (table 2) and show

TABLE 2.—*Potassium-argon ages of batholithic rocks from Yosemite National Park and of plutonic rocks of the west wall*

[After Curtis and others, 1958]

Intrusive	Formation	Location	Probable order of emplacement	Calculated age (10^6 years)
	Johnson granite porphyry.	Approx. sec. 4, T. 1 S., R. 24 E., along Tioga Pass Road (State Highway 120).[1]	7	82.4
	Cathedral Peak granite.	Approx. sec. 2, T. 1 S., R. 24 E., near Tuolumne Ranger Station, along Tioga Pass Road (State Highway 120).[1]	6	83.7
	Half Dome quartz monzonite.	Along trail to Vernal Falls from Yosemite Valley, below Falls.[1]	5	84.1
	Sentinel granodiorite.	Yosemite Valley at north base of Sentinel Dome.[1]	4	88.4
	El Capitan granite.	Just north of State Highway 140 along Tamarack Creek in SE¼ sec. 25, T. 2 S., R. 20 E.[1]	3	92.2
	Granodiorite of the Gateway.	Along State Highway 140 in NE. corner sec. 6, T. 3 S., R. 20 E.[1]	2	92.9
	Biotite granite of Arch Rock.	Along State Highway 140 in SW¼ sec. 36, T. 2 S., R. 20 E., Yosemite National Park.[1]	1	95.3
Granodiorite near Belden.	----------------	Feather River Highway, 2 miles west of Belden, Pulga quadrangle.	---------	135.7
Granodiorite at Rocklin.	----------------	Quarry at Rocklin, Placer County (sec. 19, T. 11 N., R. 7 E.) Rocklin 7½-minute quadrangle.	---------	131
Quartz diorite of Horseshow Bar.	----------------	Quarry in NE¼ sec. 18, T. 11 N., R. 8 E., 3 miles southeast of Loomis. Pilot Hill 7½-minute quadrangle.	---------	142.9
Quartz monzonite of Guadaloupe Mountain.	----------------	Road site in NE¼ sec. 24, T. 6 S., R. 17 E., Indian Gulch quadrangle.[1]	---------	142.9

[1] Locality shown on plate 1.

remarkable agreement with the order of intrusion as established in the field by Calkins (1930). The ages of biotite from the four intrusives west of the batholith range from 130 to 143 million years, and are distinctly older than the Yosemite biotites. These ages are in the same range as biotite from the Shasta Bally batholith, which was determined by the potassium-argon method to have an isotopic age of 134 million years.

The Cretaceous period lasted from 135 to 70 million years ago according to Holmes' recent revision of his time scale (Holmes, 1960) and from 133 to 58 million years ago according to Curtis and others (1958, p. 11). On these scales both the lead-alpha determinations of zircons from the east-central Sierra Nevada near Bishop and the potassium-argon determinations of Yosemite biotites fall close to the middle of the Cretaceous. Nevertheless, much additional work is needed before either the lead-alpha or potassium-argon dates can be accepted without reservation to represent the time of emplacement of the batholith.

OLDER PLUTONIC ROCKS OF THE WEST WALL

Two groups of plutonic rocks, an older ultramafic group and a younger group that ranges in composition from gabbro to grandiorite, are represented among the plutons that intrude the western metamorphic belt. Both groups are of latest Jurassic or earliest Cretaceous age; plutons of both groups intrude Upper Jurassic (Kimeridgian) strata, and several plutons of the younger group have been dated radiometrically (table 2).

The rocks of the ultramafic group are chiefly serpentine, but include some peridotite, dunite, and saxonite. They form elongate bodies that are localized along fault zones, although some penetrate beyond the limits of the zones. Later movements along the fault zones sheared the ultramafic rocks (Ferguson and Gannett, 1932, p. 21–22; Clark, 1960a, p. 488).

Some plutons of the younger group are of very nearly the same composition throughout, but others, particularly smaller plutons, are compositionally heterogeneous and range from gabbro to quartz diorite or granodiorite. Two plutons of this group have been described in detail by Hietanen (1951) and Compton (1955). Where the plutons of the younger group are in contact with ultramafic plutons, the contact relations show that the ultramafic plutons are the older. Some plutons of the younger group truncate fault zones, but others are sheared or cut off at fault zones. Probably these relations reflect ranges both in the ages of the plutons and the times of last movement along the fault zones. Most of the younger group of plutons are clearly intrusive into older rocks, and are magmatic, but some bodies of gabbro may have been formed by the recrystallization of mafic volcanic rock. Durrell (1940) suggests that some gabbro crystallized from a differentiate of ultramafic magma.

CONSTITUTION OF THE BATHOLITH

The batholith is composed chiefly of quartz-bearing granitic rocks ranging in composition from quartz diorite to alaskite, but includes scattered smaller masses of darker and older plutonic rocks and remnants of metamorphosed sedimentary and volcanic rocks (pl. 1). Rocks in the compositional range of quartz monzonite and granodiorite predominate and are about equally abundant. The granitic rocks are in discrete masses or plutons, which are in sharp contact with one another or are separated by thin septa of metamorphic or mafic igneous rock or by late aplitic dikes that were intruded along contacts between plutons.

Individual plutons vary greatly in area; they range from less than a square mile to 200 or more square miles. The limits of many large plutons have not been delineated. On the whole the batholith appears to consist of a few large plutons and a great many smaller ones, which are grouped between the larger plutons. All of the larger and some of the smaller plutons are elongate in a northwesterly direction, parallel with the long direction of the batholith, but many of the smaller plutons are elongate in other directions or are either irregularly shaped or rounded.

The larger bodies of metamorphic rocks are called roof pendants, or simply pendants, and the smaller ones are called inclusions, following the assumption that the larger masses were more likely to have had direct connection with the main mass of roof rocks before erosion. Most pendants are in contact with more than one intrusive, whereas inclusions are enclosed within a single intrusive. Masses of metamorphic or mafic igneous rock that lie along contacts between different plutons, usually in thin discontinuous lenses, are called septa or screens. Some larger septa may have been continuous with the roof, and therefore are also pendants.

LITHOLOGIC GROUPINGS OF THE PLUTONIC ROCKS

The granitic rocks in the Sierra Nevada have been assigned to three kinds of lithologic units on the basis of composition, texture, and intrusive relations or other evidence of relative age (Bateman, 1961). The three kinds of units are: formations with formal names, informal units that include the rocks in several plutons, and informal units that include the rocks in single plutons. Where two or more plutons are assigned to a named formation they contain rock that is similar compositionally and texturally, and are assumed to have been emplaced at approximately the same time. One pluton assigned to each of the formally named formations contains the type locality. Most of the named formations are in the Yosemite region (Calkins, 1930) and in the east-central Sierra Nevade near Bishop (Bateman, 1961) but one formation, the Mount Givens granodiorite, is in the Huntington Lake area (Sherlock and Hamilton, 1958).

Where two or more plutons have been assigned to an informal lithologic unit, the rocks in the assigned plutons are similar compositionally and texturally, but the assigned plutons are not necessarily temporal equivalents. Generally the rocks in smaller plutons not grouped with other plutons are designated informally.

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COMPOSITIONAL CLASSIFICATION

The compositional classification that we have used (fig. 2) is modal and is similar to several published classifications. Although it has shortcomings (and a better classification could easily be made on the basis of the ratio of plagioclase (or of anorthite where known) to total felsics), it has the advantage of being in accord with common usage. According to the classification the granitic rocks must contain at least 10 percent of quartz. Boundaries between the fields of the different granitic rocks are in terms of the ratio of K-feldspar (including perthite) to total feldspar as follows: quartz diorite, 0–10 percent; granodiorite, 10–35 percent; quartz monzonite, 35–65 percent; granite, more than 65 percent. Few Sierran rocks fall in the granite field as defined; however, granitic rocks in which the modal plagioclase is albite (An_{10}) are classed as granite by many geologists even though they plot in the quartz monzonite field as defined in this report. Certain rocks in which the plagioclase is albite contain only a small percentage of mafic minerals and are called alaskite rather than granite.

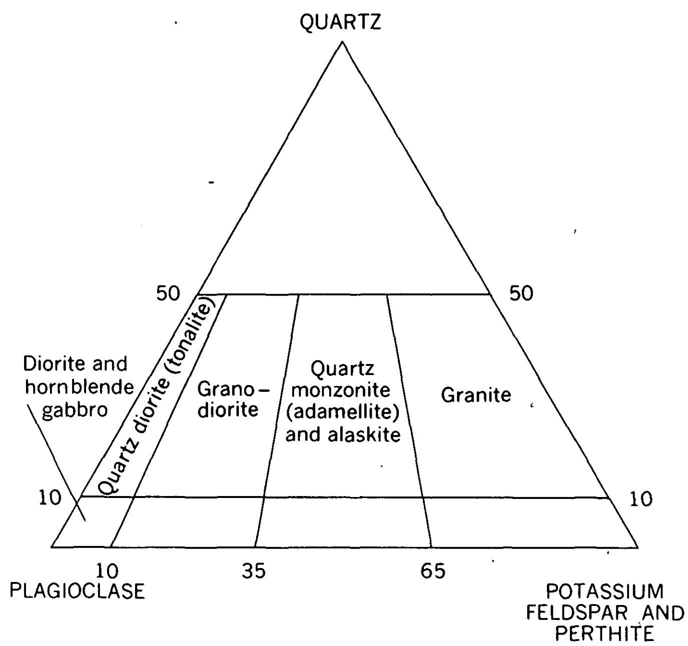


FIGURE 2.—System of classification of the granitic rocks.

Near the plagioclase corner of the diagram quartz diorite passes into diorite where the quartz content is less than 10 percent of the total content of the felsic minerals, or into hornblende gabbro where the anorthite content of the plagioclase is more than 50 percent. The Sierran intrusive rocks are quite uniform in mineral composition, and with rare exceptions mafic minerals are more abundant than felsic minerals only in hornblende gabbro. Most of the relatively small masses of older mafic rock in the Sierra Nevada are either hornblende gabbro or quartz diorite, although some are dark-colored granodiorite, and a few are diorite. Some of these rocks contain augite in addition to hornblende.

The composition indicated by the name of a lithologic unit is the average composition of the unit. Many plutons are compositionally zoned and include two or more compositional variants. Certain intrusive bodies in the Yosemite region named by Calkins (1930) do not conform to this clasification. Thus both the Cathedral Peak granite and the El Capitan granite are probably quartz monzonite according to our classification.

MAFIC PLUTONIC ROCKS

The small bodies of diorite, quartz diorite, and hornblende gabbro have been aptly called by Mayo (1941, p. 1010) "basic forerunners" or simply as "forerunners." Their distribution is reminiscent of the metamorphic rocks; they occur as inclusions or small pendants within individual plutons of more silicic rock, or as septa between plutons. Commonly they are associated with metamorphic rocks, and many are crowded with inclusions. The cause of this intimate association with the metamorphic rocks probably is that the mafic plutonic rocks were the first to be emplaced, and consequently they came in contact with metamorphic rocks on all sides. The original sizes and shapes of most masses were destroyed by later granitic intrusives, which tore them apart and recrystallized, granitized, and assimilated their fragments. In this report the term "assimilation" is used to describe the incorporation of solid rock in magma and "granitization" is used to describe the conversion, essentially in the solid state, of nongranitic rock to granitic rock. Partly as a result of original differences and partly as a result of subsequent modification, the mafic plutonic rocks are heterogeneous in composition and texture; very likely they include rocks of diverse origin.

The darker intrusive rocks (content of mafic minerals 40-60 percent) generally contain plagioclase more calcic than An_{50} and hornblende or uralitic amphibole in substantially greater abundance than augite, and are therefore classed as hornblende gabbro. Most of the hornblende gabbro ranges from 1 to 5 mm in average grain size and is medium grained, but these limits are wide enough to permit very great differences in the appearance of different rocks. Many plutons exhibit panidiomorphic-granular texture, deep zoning of plagioclase, layered facies, and dikes that cut metamorphic rocks; these factors indicate that the plutons crystallized from magma and are truly igneous. Unusual fabrics are common, however. Locally, coarse-grained, almost pegmatitic rock contains euhedral prisms of hornblende an inch or more long and a quarter to half an inch across. In several places, hornblende gabbro contains almost equidimensional crystals of plagioclase. Rocks in which elongate hornblende crystals lie in a well-defined foliation plane, but which are randomly oriented within that plane, occur in thin tubular masses bordered by younger silicic granitic rocks; this relationship suggests that the fabric may be metamorphic.

Rocks of quartz dioritic or granodioritic composition are generally lighter colored than the hornblende gabbro but are darker than the granodiorite in larger plutons. In both texture and composition the quartz diorites and dark-colored granodiorites are transitional between hornblende gabbro and the larger and younger plutons of granodiorite and quartz monzonite. The mafic mineral content of quartz diorite ranges from 20 to 40 percent, plagioclase is less calcic than An_{50} (and commonly less calcic than An_{40}), and the rock generally contains 10 percent or more of quartz. The grain size ranges from 1 to 5 mm, but quartz diorite is typically finer grained than hornblende gabbro. Most quartz diorite is hypidiomorphic-granular, and probably crystallized directly from a magma.

Many bodies of the dark-colored granodiorite and some of quartz diorite may be hybrids of more silicic granitic rocks and diorite, hornblende gabbro, or mafic volcanic rocks. The fabric is generally irregular; some rocks are very coarse grained and in places contain poikilitic hornblende crystals an inch or more long.

GRANITIC ROCKS

MINERALS

The common minerals of the Sierran granitic rocks—quartz, K-feldspar, plagioclase, biotitle, and hornblende—are common minerals in granitic rocks everywhere, and only their more significant characteristics need be described here. Quartz is in anhedral grains, many of which form mosaics of diversely oriented components. Where the grains do not form mosaics, either extinction is undulatory or sharply defined rhombic areas are visible near extinction. In a few rocks linear patterns of extinction can be seen. The association of mosaic and twinned patterns in quartz with mortar structure and throughgoing shears probably represents greater strain than is represented by undulatory extinction.

K-feldspar is white or pinkish in the hand specimen. Much of it, especially in the more felsic rocks, is perthitic and exhibits grid twinning. X-ray patterns show that both orthoclase and microcline are present.

K-feldspar is anhedral except where it forms euhedral to subhedral phencrysts, which are distinctive features of many intrusives of quartz monzonitic and potassic granodioritic composition. Most phenocrysts show Carlsbad twins. In many phenocrysts compositional zoning is revealed by the use of laboratory etch-stain techniques. Much K-feldspar is poikilitic and encloses grains of plagioclase, hornblende, biotite, minor accessory minerals, and rarely, blebs of quartz. In phenocrysts, some of the included minerals are oriented parallel with growth lines in the feldspar. Commonly the cores of phenocrysts are clean, and the included minerals are confined to a broad outer zone. Peripheral concentration of mafic minerals accentuates many phenocrysts.

Plagioclase is in relatively small, white to light-grey, subhedral grains. In finer grained dikes and apophyses plagioclase grains of about the same size and habit as those in the parent intrusive are phenocrysts. Almost all plagioclase grains are compositionally zoned and polysynthetically twinned according to the albite law, and many are also twinned according to the Carlsbad and pericline laws. Generally the overall pattern of zoning is progressive from calcic core to sodic rim, but superimposed oscillatory zones are very common. In most grains an area in the core is unzoned and contains abundant sericite; under certain conditions the cores have weathered out. Some grains also have unzoned rims of more sodic plagioclase, especially adjacent to K–feldspar. The average composition of plagioclase is albite in granite and alaskite, oligoclase in quartz monzonite, and andesine in granodiorite.

Biotite is present in all the granitic rocks, but increases in abundance with decreasing silica content. In some rocks biotite is characteristically in discrete euhedral plates, and in others it is in small irregular-shaped grains which in many rocks form clusters with hornblende and the accessory minerals. As a rule, biotite exceeds hornblende by a small percentage.

Hornblende also increases in abundance with decreasing silica content, and in the eastern Sierra Nevada is present in sodic quartz monzonite and in granite in only trace amounts. Like biotite, it is characteristically in euhedral crystals in some rocks and in ragged grains in others. Augite cores are present in the hornblende of some calcic granodiorites.

Common accessory minerals in the granitic rocks include sphene, magnetite, ilmenite, apatite, zircon, and allanite. Less common accessories are monazite and thorite. The most common secondary minerals are sericite, epidote, chlorite, and hematite. Muscovite is rare except in some alaskites.

TEXTURES

Each of the granitic rocks is characterized by a "typical appearance" which is determined largely by grain size, content of dark minerals, and texture. Except for uncommon local variants the granitic rocks are: medium grained; hypidiomorphic-granular; equigranular, seriate, or porphyritic; and contain 2 to 20 percent dark minerals.

The average grain size of most nonporphyritic rocks and of the groundmass of the commoner porphyritic ones is from 1 to 5 mm. Phenocrysts range widely in size; some K–feldspar phenocrysts are several inches long. Although correlation between plutons assigned to the same formation is based partly on grain size, a common grain size may not be entirely a consequence of belonging to the same formation. Grain size appears to be, at least in part, a function of the size and shape of the granitic mass; the rock of some formations occurs characteristically in masses of larger size than the rock of other formations; the average grain size of larger rock bodies is greater than that of smaller masses. The rock in narrow masses commonly is finer grained than rock of the same composition in wider ones. The grains are of fairly uniform size within most plutons but are commonly finer near the margins and in small apophyses.

Equigranular rocks are commoner than porphyritic or seriate ones, but all three textures are well represented. Generally one texture or the other prevails within a single pluton, but it is not uncommon for porphyritic rock to grade to equigranular rock (Knopf, 1918, p. 66). Gradation from porphyritic to equigranular can be seen in many masses of quartz monzonite similar to the Cathedral Peak granite, in the Mount Givens and Lamarck granodiorites, and in the Half Dome and Wheeler Crest quartz monzonites.

The porphyritic rocks can be conveniently placed in two groups, those with phenocrysts of K–feldspar, and those with phenocrysts of other minerals. The phenocrysts of larger intrusives generally are K–feldspar, whereas those in finer grained dikes, marginal rocks and apophyses are generally minerals that crystallized early, such as hornblende or plagioclase, although in some finer grained rocks all of the essential and varietal minerals occur in two generations. Commonly the groundmass of the finer grained rocks is allotriomorphic-granular. These relations can be explained by the mechanism that is usually assumed for porphyritic rocks: the phenocrysts are minerals that crystallized early and were suspended in the magma when its movement to a new environment resulted in more rapid crystallization.

Phenocrysts of K–feldspar are largely restricted to rock of intermediate to calcic quartz monzonitic or potassic granodioritic composition. They cannot be explained in the same way as phenocrysts of other minerals because their many inclusions of other minerals show that they formed during the later rather than the earlier stages of crystallization. In some rocks, notably the Cathedral Peak granite and its probable correlatives, and the Wheeler Crest quartz monzonite, the phenocrysts are euhedral; in most other rocks they are subhedral or even anhedral. Regardless of external form, the phenocrysts almost always enclose grains of other minerals commonly arranged in geometric zones paralled to K–feldspar faces; some phenocrysts are accentuated by peripheral concentrations of mafic minerals. The faces of the phenocrysts are irregular in detail and show interference between the phenocrysts and bordering grains. The peripheral zone of mafic minerals is reminiscent of the dark carbonaceous zones marginal to porphyroblasts of chiastolite and suggests, by analogy, that the dark minerals were pushed aside and expelled by the growing crystals.

The presence of K–feldspar porphyroblasts in certain wallrocks and inclusions that appear to be exactly like K–feldspar phenocrysts in contiguous granitic rock indicates pressure-temperature conditions like those in the magma, but we do not believe that the large K–feldspar phenocrysts in granitic rock are porphyroblasts. Dimensional orientation of K-feldspar phenocrysts near external contacts and in dikes shows that they were formed before complete consolidation of the intrusive.

Jahns (1962) has postulated concomitant crystallization from magma and a coexisting aqueous gas phase to explain giant textured pegmatite and much finer grained aggregate. Since an aqueous gas phase can separate from a cooling magma because of crystallization and consequent oversaturation of the magma with water, this mechanism might also account for large K–feldspar phenocrysts in plutonic rocks. It is difficult, however, to explain why rocks in the compositional range of calcic quartz monzonite to sodic granodiorite are more favorable for K–feldspar phenocrysts than rocks of other compositions, especially ones of granitic or alaskitic composition, which, in the Sierra Nevada, generally are equigranular.

MAFIC INCLUSIONS

Inclusions of fine-grained mafic material, here referred to as mafic inclusions, are the most abundant included material in the granitic rocks. These or similar inclusions have been called basic segregations (Knopf and Thelan, 1905, p. 239), autoliths (Holland, 1900, p. 212–219; Pabst, 1928, p. 325–386), basic concretions (Grubenmann, 1896, p. 340), and inclusions (Hurlbut, 1935, p. 614). The mafic inclusions have regular shapes and sizes; they cannot properly be described as schlieren, which generally are conceived as streaky masses of irregular form.

The inclusions are variable in size, shape, texture, and mineral content, but the range of variation is rather narrow. Commonly the size, texture, shape, and abundance of the inclusions is characteristic of the intrusive mass in which they occur. They range in size from less than an inch to several feet across, and commonly are oblate spheroids or discoids in shape; among other shapes observed, prolate spheroids are less common and spindles are rare (Balk, 1948, p. 12 and pl. 2, fig. 2, and pl. 3, fig. 1; Pabst, 1928, p. 334). Some discoidal inclusions less than an inch in thickness are many feet in outcrop length.

Minerals in the mafic inclusions are similar to those in the enclosing granite but are in very different proportions. Plagioclase generally makes up 40 to 60 percent of the inclusions, biotite 5 to 20 percent, hornblende 20 to 50 percent, and quartz 5 to 15 percent. K-feldspar is scarce, and apatite and the opaque minerals are relatively abundant. The composition of the plagioclase appears to be about the same as in the enclosing granitic rock, but in some inclusions it may be a trifle more calcic.

The common texture is granoblastic or allotriomorphic-granular; some inclusions are structureless, but many tubular ones have a foliation that is caused by the planar orientation of platy and elongate minerals. Porphyroblasts of plagioclase, hornblende, and biotite are common, although in larger inclusions they are confined to the margins. These porophyroblasts are of the same size and habit as the same minerals in the surrounding granitic rock. Locally, especially near the margins of inclusions, aggregates of porphyroblasts are mineralogically and texturally indistinguishable from the surrounding granitic rock, and make the contact difficult or impossible to identify.

Mafic inclusions are found in almost all of the granitic rocks, but are common only in granodiorites and quartz diorites. Quartz monzonites and granites generally contain inclusions only in the immediate vicinity of mafic igneous and amphibolitic wall rock. Mafic inclusions of relatively uniform size are generally rather evenly distributed in granodiorite and quartz diorite plutons of constant composition throughout, although they may be slightly more abundant near the margins. Many plutons, however, are compositionally zoned (discussion on p. 30–32), and in these plutons the number and size of mafic inclusions is correlative with the composition of the enclosing rock. For example, the number of mafic inclusions in the strongly zoned granodiorite of Cartridge Pass (fig. 3) decreases sharply inward from the margins and few inclusions are found more than three-quarters of a mile from the intrusive contact. The inclusions in this pluton are relatively uniform in size and average about 4 inches in greatest dimension. Comparison of figure 3 with two of the diagrams in figure 16 shows close correlation between the number of inclusions per unit area and the percent of dark minerals and the specific gravity of the enclosing rock. In a general way the number and average size of mafic inclusions in the Sierran granitic rock vary with the mafic mineral content, and especially with the hornblende content, of the enclosing granitic rock. Rocks with less than 5 percent dark minerals generally contain little hornblende and few inclusions.

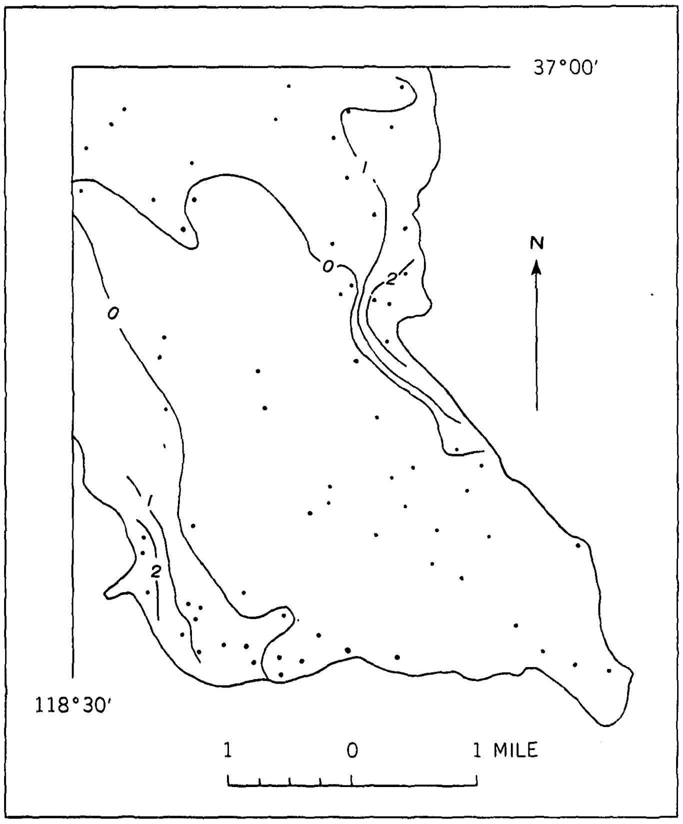


FIGURE 3.—Number of mafic inclusions per square yard of horizontal outcrop within the granodiorite of Cartridge Pass. Dots indicate points of field measurement. Compare with figure 16, especially with the diagrams showing the percent of dark minerals and specific gravity.

The inclusions are also progressively more flattened toward the margins of many plutons. At the contact, inclusions an inch or less thick and several feet in outcrop length are common. Inward from the contact the inclusions appear thicker and shorter, and beyond about a mile from the contact inclusions are generally spheroidal or irregularly rounded. Discoidal inclusions are generally parallel with the nearest contact of the enclosing pluton, and as most intrusive contacts are steep, most discoidal inclusions dip steeply. Recognition of elongation within the plane of flattening is difficult unless the elongation is extreme, which is not common.

The origin of the mafic inclusions has never been completely explained, but three alternative hypotheses merit consideration: the inclusions are (a) clots of early formed minerals; (b) refractory material that was not melted when the magma was formed; (c) fragments of wallrocks of appropriate composition possibly including early crystallizing mafic border facies of the pluton itself. The first and second alternatives are difficult to evaluate. How early crystallized minerals would collect into clots of roughly the same size is unknown. The second alternative is attractive because it explains what happens to material that is left over after selective fusion of part of the earth's crust, and because it provides an explanation for the even distribution of mafic inclusions of the same size throughout an intrusive mass, especially where wallrock of appropriate composition to form mafic inclusions is not exposed. Unfortunately, it is an alternative that may not be possible to test in the Sierra Nevada. Deeper level migmatitic terranes where magmas are presumed to have formed are probably the best places for testing this hypothesis.

The third alternative, that the mafic inclusions are truly inclusions of wallrock and roof rock, has been partly investigated, and it can be shown that some mafic inclusions are of this origin. Field evidence of different stages in the breaking up of amphibolitic wallrock into inclusions was observed at several localities. The amphibolite is intricately penetrated with anastomosing dikes of granitic rock, which separate or nearly separate individual masses of amphibolite (fig. 4).
These dikes appear to be chiefly granitization effects and to have formed by reaction of magma with the amphibolite in accordance with principles laid down by Bowen (1928, p. 197–198). Fractures presumably served as avenues for the movement of the magma. In a few places along the margins of the amphibolite, individual rounded remnants of amphibolite appear to have been caught in the act of drifting away from the main mass. This relation suggests that the granitic rock formed by reaction with the amphibolite was not completely rigid, possibly because it was close to its melting point.

FIGURE 4.—Amphibolite penetrated by anastomosing Wheeler Crest quartz monzonite.

By recrystallization any rock of appropriate chemical composition can be converted into amphibolite like that in the mafic inclusions. The formation of mineral assemblages such as those found in mafic inclusions is assured because the common metamorphic grade is in the hornblende hornfels facies. Mafic volcanic rocks can be made over by recrystallization, and coarser grained mafic plutonic rocks can be converted by reduction of grain size, possibly followed or accompanied by recrystallization (Joplin, 1935). Hurlbut (1935) has presented a strong case for the making over of gabbro fragments into mafic inclusions in the batholith of southern California by a process that involved both recrystallization and minor exchange of material with the enclosing magma. Probably some exchange of material is always involved.

Rocks such as marble and calc-hornfels can also be converted to amphibolite by a process that requires metasomatic interchange of larger amounts of material. Plagioclase-diopside hornfels partly converted to amphibolite was observed at several places, but not in large volume. Such conversion requires the addition of small amounts of H_2O, Fe, Mg, and possibly Al, and the subtraction of Ca. In a few places field evidence indicates that clean marble also was converted to amphibolite—a process that requires metasomatic exchange of much larger amounts of materials than is required for the conversion of calc-hornfels.

The presence of mafic inclusions in granodiorite and quartz diorite and their scarcity in quartz monzonite and granite can be explained in two ways, assuming that the inclusions are wall rock xenoliths. The first is that the mafic inclusions were all picked up during intrusion, and that the granodiorites, having been generally emplaced earlier than the quartz monzonites and granites, had greater opportunity to come in contact with wallrocks of suitable composition to form amphibolite. The alternate explanation, which seems to fit better the observed spatial association of mafic inclusions with rock of granodioritic or quartz dioritic composition, is that mafic inclusions were semistable in magma of granodioritic composition, but were unstable in more felsic magmas. Granitic or quartz monzonitic magma could be expected to react with mafic inclusions in accordance with Bowen's reaction principle more readily than granodioritic magma, which is closer to amphibolite in composition. It may be significant that mafic inclusions are present in the hornblende-bearing rocks and absent in the hornblende-free rocks. Tuttle and Bowen (1958, p. 91–93) suggest that some amphiboles may be unstable in the presence of moderate or high water-vapor pressure. Very likely the content of water is increased in the later differentiates of magma.

With respect to the batholith of southern California, Larsen (1948, p. 162) states that the inclusions were in almost perfect equilibrium with the granitic magma and probably would have persisted in a stagnant magma for a long time without mixing to give a homogeneous rock. The magma to which he chiefly refers is tonalitic.

Progressive decrease in the abundance of mafic inclusions away from the margins of many intrusives (and the fewer inclusions in some large intrusions) can be explained in terms of change of composition of the residual magma and a progressively longer period during which inclusions can be digested. Generally there is some evidence of compositional zoning in intrusives in which mafic inclusions become less abundant away from the margins. A likely explanation for the fact that residual magma becomes more felsic although it is digesting mafic inclusions is that the effect of differentiation outweighs that of contamination.

Reesor (1958, p. 47–48) postulates that the inclusion-bearing marginal ganodioritic shell of the strongly zoned White Creek batholith in British Columbia is simply contaminated quartz monzonite. Two objections can be made to Reesor's hypothesis as applied to Sierra Nevada plutons. First, demonstrable contamination of Sierran granitic rocks is spatially related to specific rock types, notably mafic ones, and the contamination pattern is not concentric, especially where several different kinds of rocks are marginal to the pluton. Second, the hypothesis implies that granodiorite containing mafic inclusions, whether zoned or not, is contaminated rock that originally was more felsic. The compositional range of plagioclase in granodiorite, however, is systematically more calcic than in quartz monzonite, and except in obviously contaminated zones, granodiorite contains no relict picked-up crystals of calcic plagioclase. These relations suggest strongly that the granodiorite crystallized at higher temperatures than the quartz monzonite and was not derived from it by contamination.

PRIMARY FOLIATION AND LINEATION AND FLATTENING OF MAFIC INCLUSIONS

Primary foliation and lineation in granitic rocks are shown by the preferred orientation of such inequidimensional minerals at biotite, feldspar, and hornblende, and of metamorphic inclusions, but are shown most conspicuously by mafic inclusions. Because mafic inclusions are scarce in most bodies of granite and quartz monzonite, primary structures are usually obscure in these rocks; they are most conspicuous in granodiorite and quartz diorite.

Minerals and mafic inclusions are brought to a state of preferred orientation in somewhat different ways, although where both are present in the same outcrop they are generally approximately parallel. Minerals are oriented by external rotation, whereas mafic inclusions can be oriented either by external rotation or by physical flattening. In many plutons the shapes of mafic inclusions change progressively from irregularly rounded in the interior of the pluton to flattened near the margins. The flattening of the inclusions probably took place when they were not much stiffer than the enclosing magma because of melting of their lowest melting constituents.

Mackin (1947, p. 26–32) has suggested that in a radially spreading intrusive body the direction of flow is toward the external contacts and that the stretching of inclusions or other enclosed bodies would be normal to that direction. If a pluton was hemispherical, the inclusions within the stiffer marginal parts of the pluton would be stretched equally in all directions parallel to the external surface of the intrusion. The inclusions nearest the margins would be stretched and flattened most and those farther from the external contact would be progressively less flattened. However, any departure from the hemispherical form would cause elongation of the inclusions in one direction or another within the plane of flattening, resulting in a lineation. This stretching process would, of course also orient the platy and elongate minerals. In the field, lineation is best shown by the alinement of hornblende crystals, but even where hornblende crystals are well alined it is commonly difficult to determine whether or not the mafic inclusions are triaxial.

Growth of a pluton by expansion requires space, and unless large amounts of wallrock and roof rock were in some way incorporated in the magma or stoped and sunk, the walls must have been crowded upward and outward. The mafic inclusions themselves may represent wallrock material, and it is likely that some additional material has been digested by the magma. Nevertheless, most intrusives appear to have incorporated far too little of the exposed wallrock to account for the needed space. Deformation of the wallrocks is evident in many places, but only in a few places where the walls are earlier granitic intrusive rocks. A possible explanation for the apparent absence of deformation in granitic rocks that have been intruded by later plutons is that the earlier rocks were softened by heat from the later magmas. It is also likely that in some places the evidence of deformation is subtle and as a consequence was overlooked.

LAYERED FACIES AND SCHLIEREN

Zones of streaked or layered rock are common along contacts between plutons of markedly different compositions (fig. 5), in inclusions of more mafic granitic rocks enclosed in more felsic granitic rocks (figs. 6*A* and 6*B*), and in felsic dikes contaminated by mafic material (fig. 6*C*). Cloos (1936, p. 280–286) included these zones among features he classed as schlieren. According to him, schilieren indicate zones of intense motion, and may result from differential movement within the magma. Some zones, however, especially ones in which the streaks or layers dip gently, contain graded layers suggestive of magmatic sedimentation.

Graded layers result from progressive upward decrease in the dark mineral content (fig. 6*B*). Gilbert (1906, p. 323-324) and recently Sherlock and Hamilton (1958, p. 1258–1259) suggested that the layers originated through a process of sedimentation from a flowing magma. Gilbert's observations were of a nearly horizontal sequence in which several apparent unconformities were present. He postulated that the unconformities were not only apparent but actual and that the layers were deposited from and partly eroded by liquid magma in motion. Sherlock and Hamilton observed similar gently dipping sequences in which the concentration of dark minerals is graded within layers which also exhibit apparent unconformities. A photograph of the base of the sequence (Sherlock and Hamilton, 1958, pl. 7, fig. 2) shows gently dipping layers resting discordantly on quartz monzonite which has a steep primary foliation shown by discoidal mafic inclusions. These relations suggest that the layering is in a gently dipping dike somewhat thicker than that shown in figure 6*C*. Sherlock and Hamilton state (1958, p. 1259) "Numerous low-angle 'unconformities' suggest that each layer formed a discrete internally flowing unit rather than as a static layer."

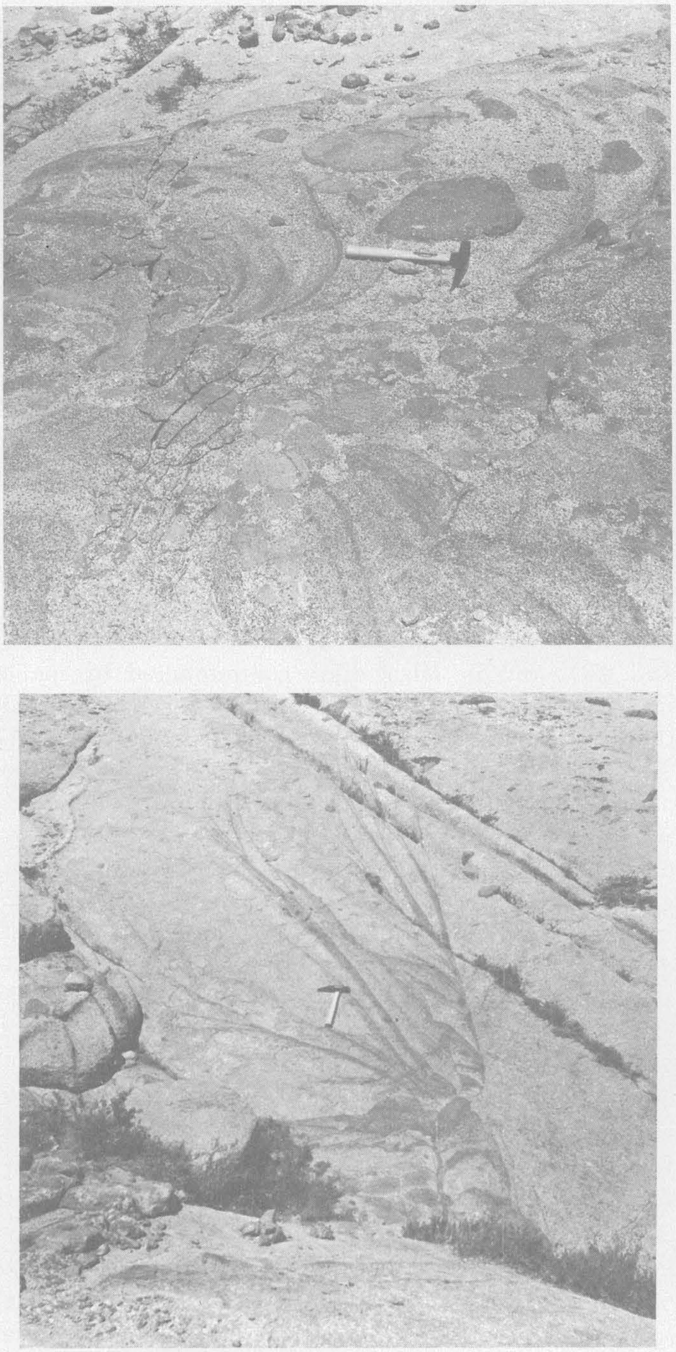


FIGURE 5.—Two views in margin of Mount Givens granodiorite at Courtright Reservoir showing the association of mafic inclusions with schlieren.

It is by no means certain that all the layered zones were formed in the same way. In some exposures the close similarity between the structures observed and structures in sedimentary rocks, such as graded bedding, crossbedding, unconformities, and channel-structures, suggests magmatic sedimentation at the interface between largely liquid magma and essentially solid rock. A mechanism that might explain grading in vertical layered sequences is preferential segregation of crystals into the slower moving side of the layer, which normally would be the side closest to solid rock due to friction and heat loss. Neither of these concepts explains graded layering in inclusions or in marginal parts of the older of two contiguous plutons, where the layering generally is gently dipping. Doubtless the layered rocks were softened when the layers formed. A possible explanation is that differential movements along shear planes separating layers was accompanied by settling of the heavier minerals within each layer, but not across the planes between layers.

SEQUENCE OF INTRUSION

The major plutons in the western part of the batholith are generally older than those in the eastern part. Nevertheless, some older plutons lie along the east side of the batholith, and the sequence of intrusion is probably more correctly described as assymetric than as one sided. The positions of the boxes in the explanation to the geologic map (pl. 1) represent our best guesses as to the relative ages of the different intrusives. These guesses are compatible with the known intrusive relations, but age relations between many intrusives are unknown. The batholith has been bridged by geologic mapping at two places: through the Yosemite region, and 100 miles farther south through the Huntington Lake region.

YOSEMITE SEQUENCE

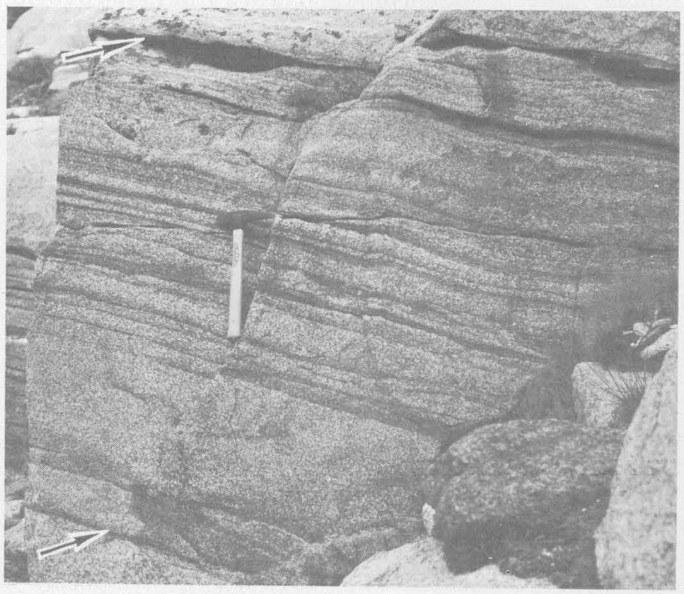
The sequence of intrusion in the Yosemite region was established by Calkins (1930). Except for small bodies of gabbro, diorite, and mafic quartz diorite, the oldest intrusive, by contact relations, is the granodiorite of the Gateway. However, the biotite granite of Arch Rock, which is enclosed in the granodiorite of the Gateway, has an older potassium-argon age (table 2). The granodiorite of the Gateway was intruded by the El Capitan granite, which was subsequently intruded by two granites and a quartz monzonite of small extent, and by the Sentinel granodiorite, the earliest member of the Tuolumne intrusive series. Of the smaller intrusives, the Taft granite is clearly older and the Bridalveil granite is probably younger, than the Sentinel granodiorite; the age of the Leaning Tower quartz monzonite (too small to show on pl. 1) has not been determined. Emplacement of the Sentinel granodiorite was followed by the emplacement of other members of the Tuolumne intrusive series—the Half Dome quartz monzonite, the Cathedral Peak granite, and the Johnson granite porphyry, in that order.

The members of the Tuolumne intrusive series form a concentric pattern, suggesting that they crystallized successively from a single mass of intrusive magma that cooled inward, and that the contacts between different formations represent upward movements of the core magma. Thus, the Tuolmne intrusive series could be considered a single enormous compositionally and texturally zoned pluton which has several internal contacts.

EASTERN SIERRA NEVADA-HUNTINGTON LAKE SEQUENCE

The continuous belt of metamorphic rocks along the west side of the batholith does not extend far south of Yosemite, and the batholith is nearly twice as wide at Huntington Lake as in the Yosemite region. Except for small bodies of gabbro, diorite, and mafic quartz diorite, and the older sheared granitic rocks in the Goddart pendant and the Mount Pinchot quadrangle, the oldest intrusive rock is hornblende-biotite granodiorite of "Dinkey Creek" type (possibly quartz diorite according to our classification) which lies along the west side of the batholith. As mapped by Krauskopf (1953), Macdonald (1941), and Hamilton (1956), this intrusive is very large and continues both to the northwest and southeast beyond the area that has been mapped. However, the rock is variable in composition and texture, and it is possible that detailed mapping will show more than one intrusive body.

The hornblende-biotite granodiorite of "Dinkey Creek" type is intruded on the east by several small intrusives of granite and quartz monzonite, and the extensive Mount Givens granodiorite. Some small bodies of granite and quartz monzonite that intrude the granodiorite of "Dinkey Creek" type also intrude the Mount Givens granodiorite, and hence are younger than the larger units. The Mount Givens granodiorite is lithologically similar to the Lamarck granodiorite and the Half Dome quartz monzonite. The Round Valley Peak granodiorite, which comprises two plutons in the eastern Sierra Nevada, is also possibly correlative, with the Mount Givens.



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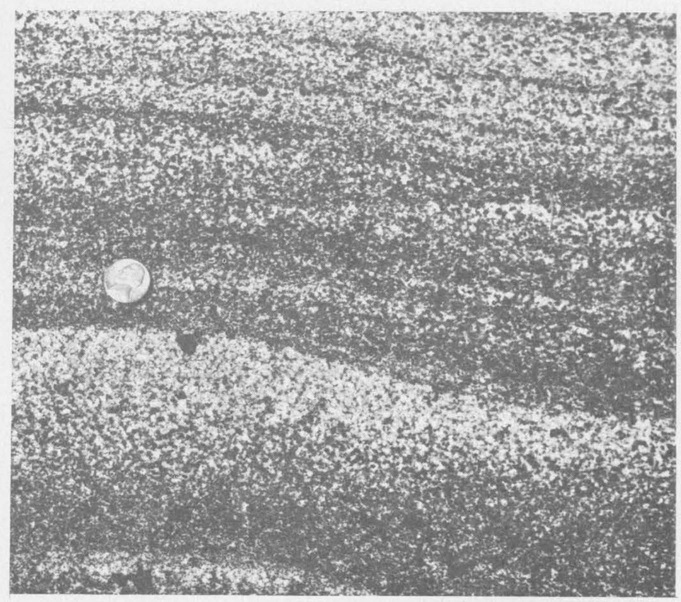

B

C

FIGURE 6.—Gravity layered schieren in inclusions and in a dike. *A*. Layering in an inclusion of Inconsolable granodiorite in quartz monzonite similar to the Cathedral Peak granite. Arrows show upper and lower contacts of inclusion. Locality is along trail along Big Pine Creek, about 5 miles above road's end. *B*. Close-up of gravity layering in an inclusion of quartz diorite in Half Dome(?) quartz monzonite. Locality is near junction of North and Middle Forks of San Joaquin River. *C*. Layering in a felsic dike cutting across foliation in hornblende-biotite granodiorite, "Dinkey Creek" type. Dike has been contaminated by mafic material present along its lower adge. Locality is a quarry at the north end of Shaver Lake.

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The Lamarck granodiorite was intruded on the east by the Tungsten Hills quartz monzonite and a quartz monzonite similar to the Cathedral Peak granite, which also intruded the Tungsten Hills quartz monzonite. The rock designated quartz monzonite similar to Cathedral Peak granite is identical in many places with typical porphyritic Cathedral Peak grantite from Yosemite, but grades to nonporphyritic rock. Alaskite similar to Cathedral Peak granite is equigrandular, but along Big Pine Creek grade to nonporphyritic quartz monzonite. Very likely the alaskite represents a somewhat later stage of differentiation than the quartz monzonite.

Some of the plutons grouped under "finer-grained quartz monzonites" on the map explanation may be younger than the rocks similar to the Cathedral Peak granite and about the same age as the Johnson granite porphyry, but the intrusive relations are obscure and it is possible that rocks of different ages are included under "finer grained quartz monzonites." The youngest intrusive in the eastern Sierra Nevada, by observed intrusive relations, is the granodiorite of Cartridge Pass, but the grandiorite of Coyote Flat may be of about the same age.

Older intrusives in the eastern Sierra Nevada are the Inconsolable and Tinemaha granodiorites, and the Wheeler Crest quartz monzonite. The Inconsolable and Tinemaha granodiorites and the granodiorite of McMurry Meadows have several mineralogic and textural characteristics in common, and probably are closely related. The granodiorite of McMurry Meadows is enclosed in the Tinemaha granodiorite and is strongly zoned, and probably crystallized from the same body of magma. The sharp contact between the two intrusives may reflect movement of the core magma after marginal crystallization.

CONTACT RELATIONS

CONTACTS BETWEEN DIFFERENT GRANITIC ROCKS

Most observed contacts between the different granitic rocks are sharp and steep or vertical; plutonic breccias occur locally along only a few contacts. Probably most contacts dip steeply in the direction of the older rock. Generally the traces of contacts between granitic rocks are flowing curves, but in places straight-line segments meet at sharply angled corners. Dikes of later rocks in earlier ones are present near contacts, but are rarely abundant. At many contacts, adjacent granitic rocks are separated by thin, discontinuous septa of metamorphic or mafic rock or by later aplitic dikes. A notable feature along all contacts between granitic rocks is the absence of any evidence of chemical reaction between the rocks in contact.

Typical sharp contacts are shown in figure 7. The contacts shown were followed in the field for many miles. The contact between quartz monzonite similar to the Cathedral Peak granite and the Inconsolable granodiorite is similar everywhere it was examined; the contact between the Inconsolable and Tinemaha granodiorites, on the other hand, is discontinuously occupied by thin metamorphic septa. A typical septum that extends northward from the Bishop Creek pendant across Bishop Creek (fig. 8) separates Tungsten Hills quartz monzonite from Lamarck granodiorite.

Zones of plutonic breccia are scarce, but a moderately extensive zone is exposed between the Tungsten Hills quartz monzonite and the Lamarck granodiorite in the vicinity of Piute Pass in the Mt. Goddard quadrangle. Much metamorphic and dioritic or mafic hybrid rock is present along the contact, and the metamorphic and mafic rock and the Lamarck granodiorite are intricately penetrated in the breccia zone by anastomosing dikes and apophyses of Tungsten Hills quartz monzonite. An unusual type of breccia is present along Big Pine Creek, where Inconsolable granodiorite adjacent to quartz monzonite similar to the Cathedral Peak granite is shattered, and the interstices between fragments are filled with felsic rock and milky quartz (fig. 9). In addition local small-scale segregations of the light and dark minerals have formed in the Inconsolable granodiorite. Some of the segregations are linear and follow lines of fracture, and some are ovoid. The shattered zone coincides with a bend in the contact; the quartz monzonite lies on the convex side, and the shattering was probably caused by movements within the quartz monzonite during the later stages of its consolidation.

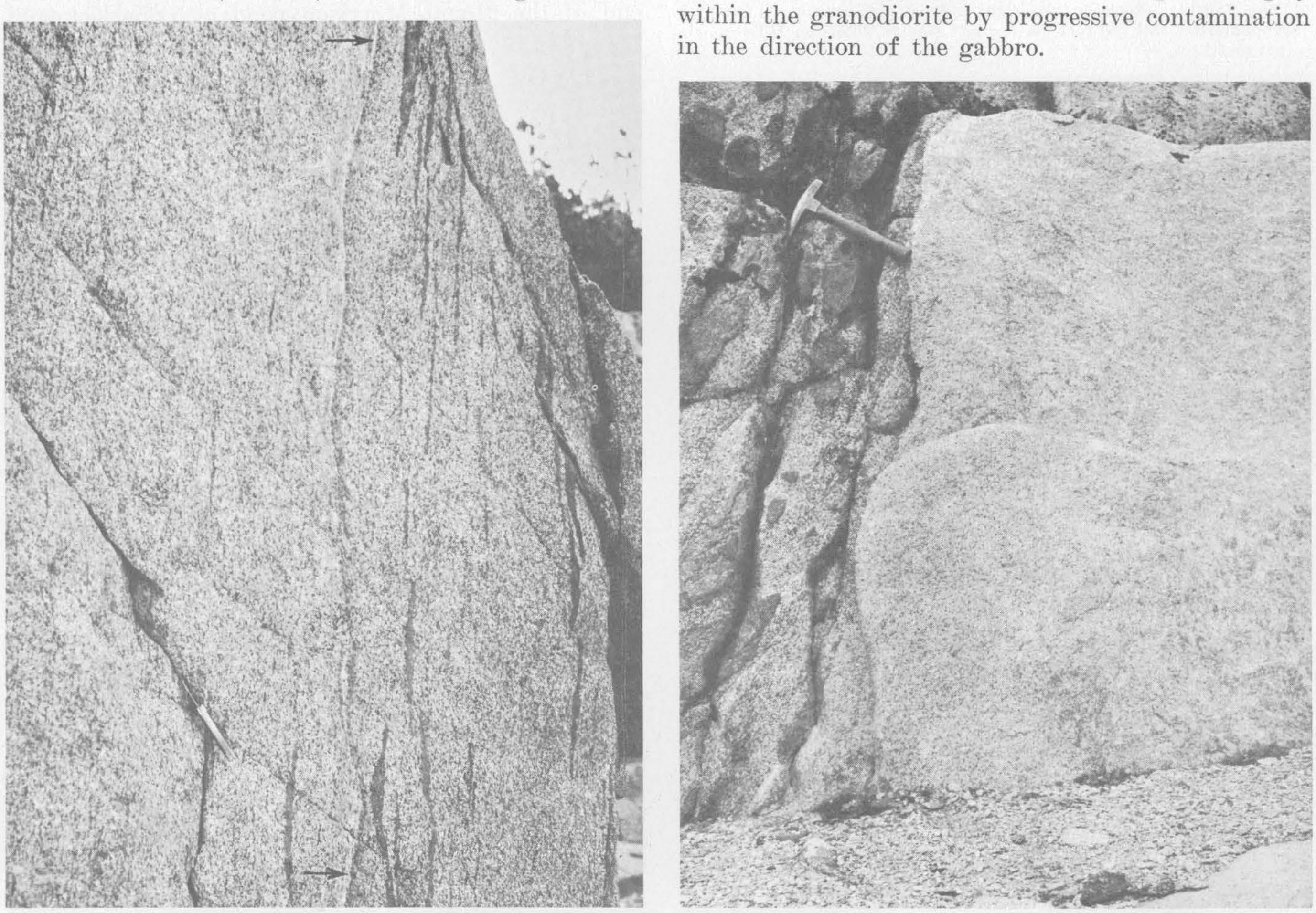
Aschistic dikes, though not abundant, generally are common enough to provide a means of determining the relative ages of the granitic rocks. In general, masses of rock similar to the Cathedral Peak granite are accompanied by more dikes than the other intrusives. The significance of swarms of marginal dikes that dip into the parent intrusive are discussed in a later section.

CONTACTS BETWEEN GRANITIC ROCKS AND METAMORPHIC ROCKS OR DIORITE

Most contacts between granitic rocks and metamorphic rocks or diorite are sharp and clean, but in places irregular contacts or broad zones of mixed rock are present. The sharpest, cleanest, and most regular contacts are between granitic rocks and metasedimentary or felsic metavolcanic rocks. Mixed zones involving these rocks are intrusive breccias that consist of angular fragments of metamorphic rocks enclosed in granitic rock. The angular outlines of the fragments and the straightness of their sides indicate that little or no chemical reaction has taken place between the fragments and the invading granitic rock.

Most contacts between granitic rocks and metamorphic rocks are steep, although gently dipping contacts occur locally. The nearly vertical sides of the Pine Creek pendant are exposed through a vertical distance of more than 5,000 feet (fig. 10) and the sides of many other pendants are exposed over vertical distances almost as great. Contacts that are concordant with the layering or bedding of the metamorphic rocks generally are more regular than contacts that are discordant. Intrusive breccias generally coincide with zones in the metamorphic rocks where the prebatholithic structures have been strongly disturbed and the rocks fragmented. In some places the fragmentation was probably caused by emplacement of the adjacent granitic rock.

Most contacts between granitic and older mafic metavolcanic rock, diorite, or hornblende gabbro are also sharp, but many are migmatitic because of chemical reaction between the granitic rocks and the invaded rocks. Sharp contacts are commonly cuspate because of reaction between granitic magma and mafic rock. The products of reaction are varied, both in composition and in texture. They are hybrid rocks of intermediate compositions, which are in part the products of contamination of the granitic rocks with more or less assimilated mafic material, and in part the products of additions of quartz and feldspar to mafic rocks. Reaction between granitic magma and hornblende gabbro is well shown on the north side of Pine Creek, northwest of Pine Lake, where hornblende gabbro grades northward over a distance of half a mile into granodiorite. The gradational zone contains a sharp though highly irregular contact between gabbro and heavily contaminated granodiorite of only slightly different appearance. The compositional gradation here appears to have taken place largely within the granodiorite by progressive contamination in the direction of the gabbro.



A *B*

FIGURE 7.—External contacts of the Inconsolable granodiorite in the drainage basin at the head of the South Fork of Big Pine Creek. *A*. Contact with Tinemaha granodiorite (right half of photograph). Mafic inclusions are present in both rocks parallel to contact and to foliation, suggesting proximity to the original intrusive contact of both masses. Inconsolable granodiorite is probably older, but relations are uncertain. *B*. Contact with quartz monzonite (right) similar to the Cathedral Peak granite. Note absence of flattened mafic inclusions in quartz monzonite, which is the younger rock.

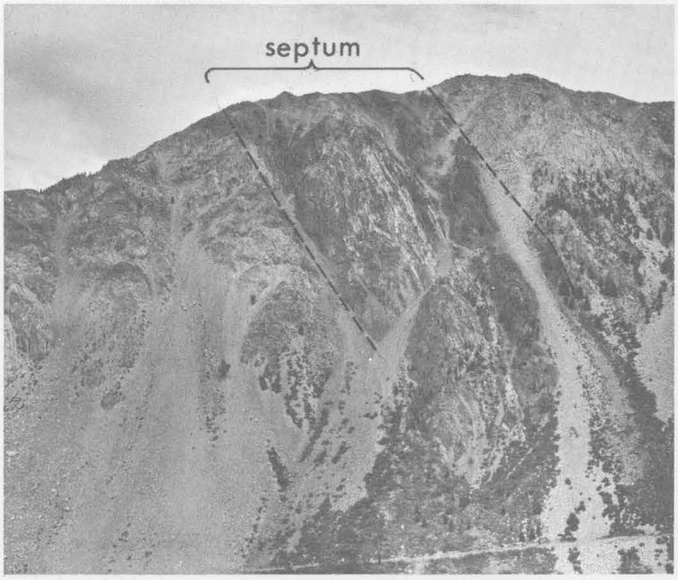
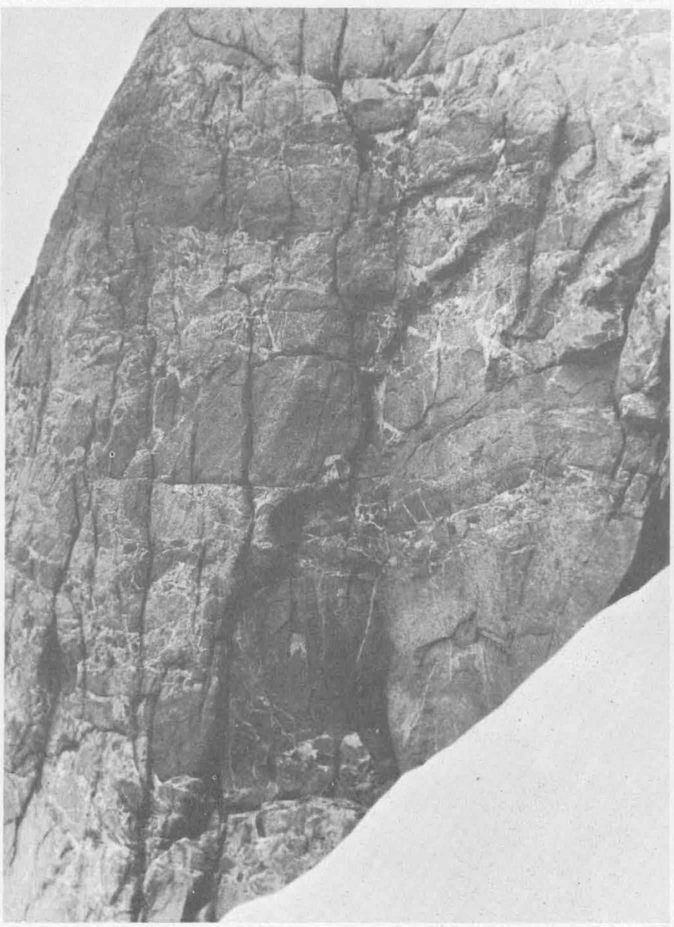


FIGURE 8.—Septum of metamorphic rock across Bishop Creek below Lake Sabrina. Rock on the right side of the septum is Lamarck granodiorite, and the rock on the left side is Tungsten Hills quartz monzonite.

FELSIC DIKE SWARMS

Swarms of dikes in two orientations have been mapped: steeply dipping dikes in conjugate sets, and gently dipping dikes marginal to certain plutons. Steeply dipping dikes have been observed in the north end of the Goddard pendant and locally in the Ritter Range pendant. The dikes in the Goddard pendant are of fine-grained, generally porphyritic rock, and are in two sets: one strikes about N. 30° W., and the other about N. 80° W. Locally, the felsic dikes are accompanied by parallel younger mafic dikes that are generally thinner but more numerous.

Gently dipping swarms of aplite, pegmatite, and alaskite dikes locally border several plutions, and are especially common along a pluton of quartz monzonite similar to the Cathedral Peak granite, which extends northwestward from the Pine Creek pendant to the Ritter Range pendant. They occur also in the south end of the Mount Morrison pendant (Sherlock and Hamilton, 1958, p. 1260–1261), and west of the Pine



A

B

FIGURE 9.—Brecciated Inconsolable granodiorite. Fragments are cemented by fine-grained felsic igneous rock and by quartz. The cementing materials almost certainly were derived from adjacent quartz monzonite similar to Cathedral Peak granite, which caused the brecciation. *A*. General view. *B*. Close view.

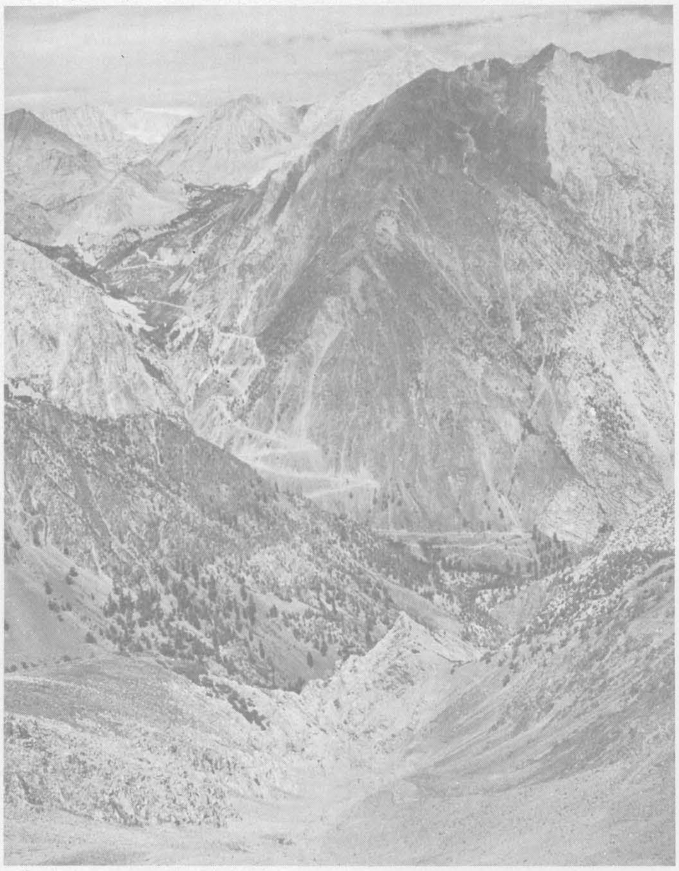


FIGURE 10.—View of Pine Creek pendant looking north across Pine Creek canyon. Contacts between the pendant rocks (dark) and the bordering Tungsten Hills quartz monzonite (light) are generally clean and sharp.

Creek pendant in upper Rock Creek and Pine Creek (fig. 11).

The dikes consist chiefly of quartz and feldspar, but locally contain garnet, sphene, biotite, magnetite, and other minerals. In texture they range from aplitic to pegmatitic, and not uncommonly both textures are present in the same dike. Dikes with dominantly pegmatitic texture almost invariably cut dikes with dominantly aplitic texture, although no great age difference seems likely. Matching irregularities in the opposing walls of many dikes indicate that they were emplaced by dilation of their walls, although the geometry of some matching walls indicates that translatory movement occurred.

Along Pine Creek, dikes extend in abundance for more than a mile from their parent mass, and a few are found at distances of more than 2 miles. Close to the parent mass, dikes comprise at least half of the rock exposed in extensive cliff faces, whereas half a mile away dikes comprise only about 10 percent of the exposed rocks. Few of these dikes continue into the Pine Creek pendant, even where it is very close to the pluton parent of the dikes; most dikes that reach the pendant finger out in the marble along the west side (fig. 11*A*).

Similar dike swarms have been described by Hans Cloos and his associates (Balk, 1948, p. 101–106). They interpret the fractures occupied by the dikes as the result of tension caused by the upward movement of magma along a steep or vertical face. According to Balk (1948, p. 104),

> The zones of marginal fissures are of the utmost importance for the evaluation of the forces acting during the consolidation of many igneous masses. Even where the rock may be structureless, marginal dikes and upthrusts must be regarded as evidence of a strong upward motion of the plutonic mass.

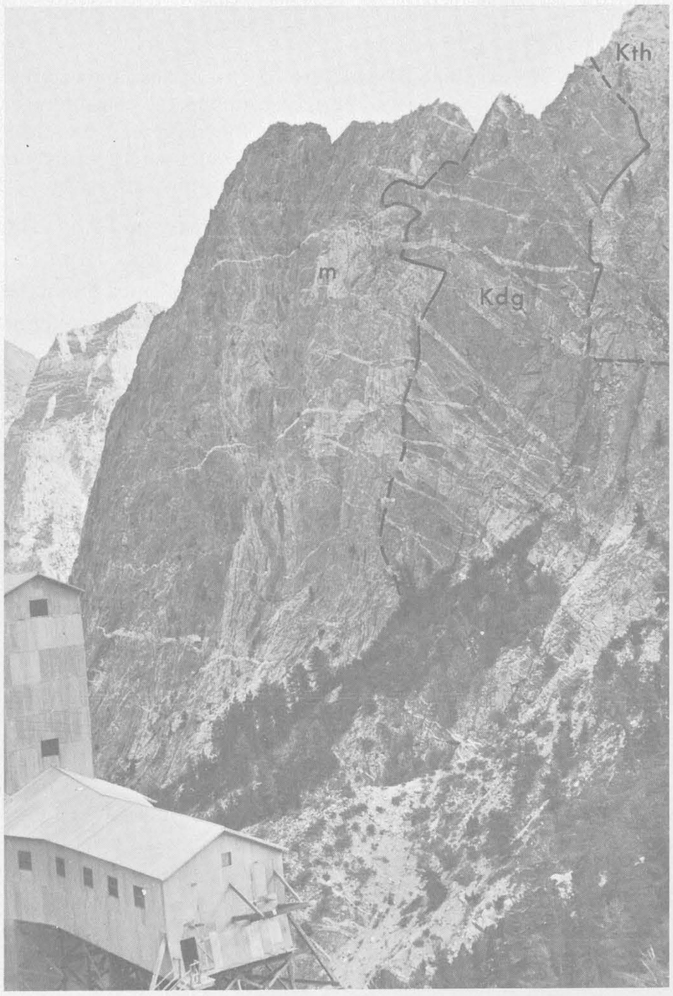
This concept of origin fits the known facts in the Sierra Nevada. Matching walls indicate dilation, locally accompanied by slight movements parallel with the walls. That the dikes die out away from the parent mass indicates progressively greater dilation toward the parent pluton, which must signify upward drag of the wall rocks adjacent to the parent pluton. That the dikes maintain a constant range of dips indicates that the fractures they occupy must have formed contemporaneously with the drag, else they would not have been formed or would have been bent upward near the parent pluton from their original configuration. These considerations indicate that the fractures are feather joints and must have been formed and the dikes emplaced in sequence from bottom to top as the parent pluton rose.

Figure 12 is a highly diagrammatic cross section through a dike swarm in which all the dikes dip 25° into the parent mass, and the external contact of the parent mass is vertical. The traces of hypothetical planes are shown which, before emplacement of the underlying dikes, were parallel to medial planes through the dikes. These traces show the amount of deformation of the wallrock that must have been caused solely by emplacement of the underlying dikes. That is, the amount of upward displacement of these traces at any place equals the aggregate thickness of dikes below the planes of reference measured perpendicular to the medial planes of the dikes. The lifting is cumulative through only limited distances, and the amount of lifting at any place equals the thickness of underlying dike rock measured perpendicular to the dip of the dikes; it does not equal the aggregate thickness of all the underlying dikes.

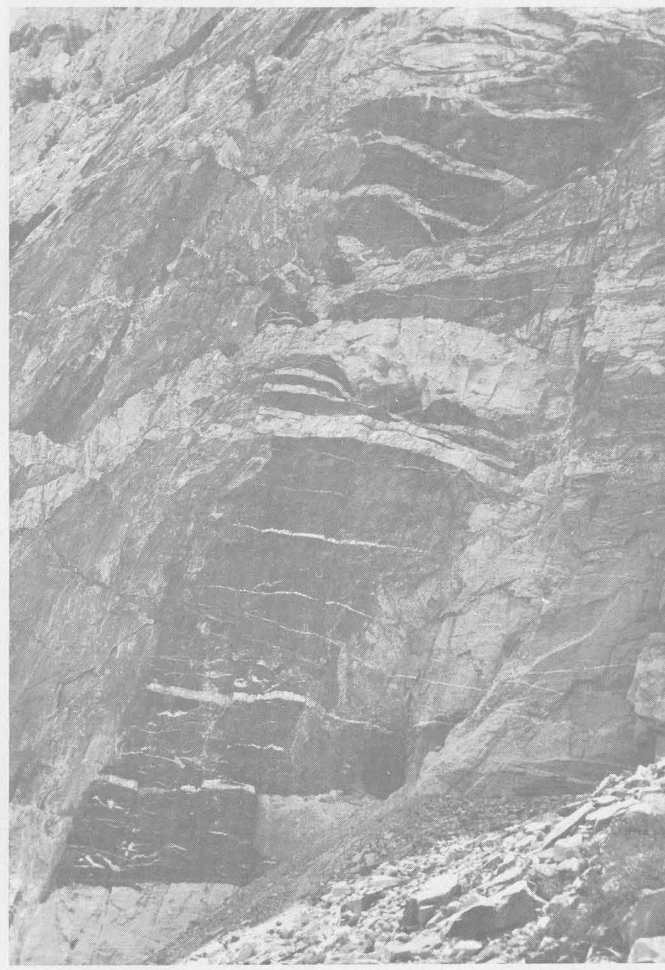
MAFIC DIKE SWARMS

Dark-colored dikes, dominantly of dioritic composition, are in many places in the central Sierra Nevada, and in the southeastern part constitutes several swarms (pl. 1). The rock in these dikes has been termed malchite by Gilbert (1941, p. 784) and diorite porphyry by Knopf (1918, p. 71). Most of the mafic dikes in swarms fall into two steeply dipping sets, one striking about N. 30° W., and the other about N. 80° W.

The most extensive swarm is in the south half of the Big Pine quadrangle and the Mount Pinchot quadrangle. Most of the dikes in this swarm strike N. 30° W., but those in the north end, in the Big Pine quadrangle, strike N. 80° W. This swarm extends discontinuously to the northwest as far as the north end of the Goddard pendant. Both sets of mafic dikes are represented there, and are parallel to older felsic dikes. Dikes striking N 80° W. also constitute a swarm that extends westward across the Tungsten Hills and along Pine Creek to the Pine Creek pendant. Mafic dikes strike N. 80° W. along Glacier Divide, and may be accompanied by dikes in other orientations. Southeast of the area shown on plate 1 the general zone of mafic dikes has been traced more than 50 miles across Owens Valley to the Inyo Mountains and the Argus Range (Moore and Hopson, 1961).



A



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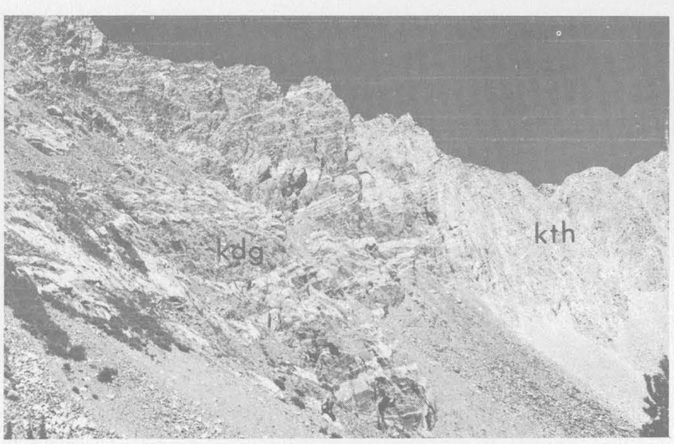


C

FIGURE 11.—Aplite, pegmatite, and alaskite dikes along Pine Creek. The dikes dip toward the parent mass of quartz monzonite similar to the Cathedral Peak granite, commonly at 20° to 30°. *A*. Along the west side of the Pine Creek pendant, dikes cut Tungsten Hills quartz monzonite (Kth) and quartz diorite (Kdg), and finger out in marble (m). *B*. At the Brownstone tungsten mine dikes cut Tungsten Hills quartz monzonite (right), tactite (center), and marble (left). *C*. West of Pine Lake dikes cut hornblende gabbro (Kdg) and Tungsten Hills quartz monzonite (Kth).

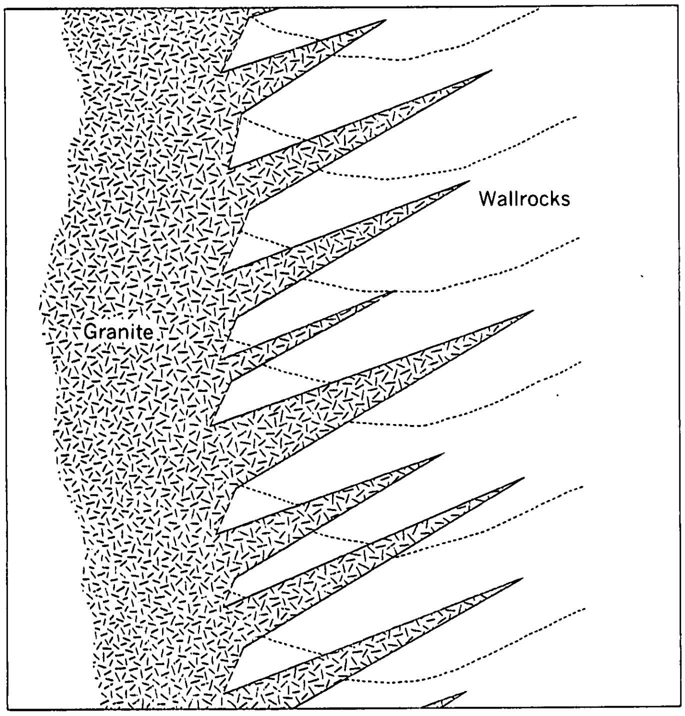


FIGURE 12.—Diagrammatic section through an intrusive and its marginal dikes to show the amount of deformation in the wall rocks caused solely by the emplacement of the dikes. Dotted lines show traces of planes of reference originally parallel to the medial planes through dikes before the underlying dikes were emplaced.

The dikes are generally a few inches to several feet thick; a few are 20 to 30 feet thick. All of the dikes are fine grained, but differences in grain size and texture are evident. Most dikes are equigranular, but some appear porphyritic. Many have been strongly sheared parallel to their walls and are schistose. Granoblastic and lepidoblastic textures prevail in sheared dikes, and the dark minerals are commonly in lenticular streaks.

Most of the dikes are green hornblende-plagioclase rocks with biotite and subordinate quartz. The plagioclase is generally oligoclase or andesine, and is zoned in undeformed dikes but unzoned in dikes with granoblastic or lepidoblastic texture. Large clots of small hornblende granules in deformed dikes probably are recrystallized relicts of original mafic phenocrysts.

The age relations of the dikes are in some doubt. In the Mount Pinchot quadrangle all the dikes appear to be about the same age, and their distribution and intrusive relationships serve to distinguish two sequences of granitic intrusions (Moore and Hopson, 1961). In this area several granitic intrusions clearly cut off mafic dikes. Northward, however, these relations do not seem to hold; both the Lamarck granodiorite, which in the Mount Pinchot quadrangle is considered to belong to the post-dike series, and the younger Tungsten Hills quartz monzonite are cut by mafic dikes.

At least two interpretations of the relations are possible: (a) mafic dike swarms were intruded at different times during the period of emplacement of the granitic rocks, but all or most of the dikes in a given swarm are of the same age; and (b) the distribution of dikes is largely controlled by the response of the different granitic rocks to stress, and though most of the mafic dikes are younger than all or most of the granitic rocks, they cut only plutons with structures favorable for the emplacement of dikes. A choice between these interpretations cannot be made with the information now in hand. Mafic dikes are most abundant in the older intrusives, but this could be a consequence of the greater number of fractures generally present in these rocks.

The source of the dike magma also is an unsolved problem. It is difficult to understand how usual processes of magmatic differentiation could produce magma of dioritic composition from a parent magma which before, and afterward, yielded magma of granodioritic or quartz monzonitic composition. In view of this difficulty two other sources for the mafic dike magma merit consideration: (a) the magma for the dikes was derived from a deep source separate from that of the granite, and (b) the dike magma was remobilized from masses of older diorite, gabbro and mafic volcanic rock. The chief argument in favor of the first suggestion is that the main dike swarm extends many miles along the strike, which suggests a single nonlocal source. The arguments for the second suggestion are (a) that on the basis of their texture and mineral content many dikes are indistinguishable from mafic inclusions and (b) that in many places, as along the north side of Glacier Divide and along the South Fork of Big Pine Creek, the dikes are closely associated with abundant inclusions of dark-colored plutonic or volcanic rocks. Further, the main dike swarm is parallel to and closely associated with the main string of Mesozoic metavolcanic roof pendants of the eastern Sierra Nevada (pl. 1). Solution to the problem of the source, as that of the age relations, of the dikes must await further study.

CHEMICAL AND MINERALOGIC TRENDS

ANALYTIC DATA

Two kinds of analytic data were obtained: limited chemical data, and extensive modal data. All of the analytic data is from the eastern part of the batholith; only chemical analyses made many years ago of rocks of uncertain affiliations are available from the western part of the batholith, and they have not been used by us. Twenty-seven samples were analyzed chemically: 10 by standard methods (including one partial analysis), and 17 by the rapid method described by Shapiro and Brannock (1956). Nine of the specimens analyzed by standard chemical methods were also analyzed by quantitative spectrographic methods for minor elements. With one exception the chemical analyses used were made in the laboratories of the Geological Survey since World War II. The exception is a partial analysis of alaskite published by Knopf (1918, p. 68). Table 3 summarizes the chemical data and gives norms calculated according to the C.I.P.W. method.

The modes of several hundred samples of granitic rock were determined during our study of the batholith. Modes of finer grained rocks were determined from thin sections using a point counter of the type described by Chayes (1949). Modes of coarser grained rocks, however, were determined using rock slabs in which K-feldspar had been stained yellow and plagioclase light red to reddish brown by the method described by Bailey and Stevens (1960). The slabs used have several times the area of an ordinary-size thin section and provide a more representative sample of rocks with an average grain size of more than 2 mm. Slabs have the disadvantage that only four constituents can be counted: quartz, K-feldspar, plagioclase, and mafic minerals. In this report individual modes are given for only a few compositionally zoned plutons. Modal averages are given, however, to show compositional variations among plutons.

BROAD CHEMICAL AND MINERALOGICAL TRENDS

A broad trend, with some local reversals, from calcic and ferromagnesian to silicic and alkalic is shown in the east-central Sierra Nevada by the sequence of intrusion. The chemical and modal analyses provide a basis for examining variations in compositions and mineral content quantitatively as well as qualitatively.

Figure 13 shows a variation diagram in which the principal oxides are plotted against SiO_2. The SiO_2 ranges from 59.0 percent in quartz diorite to 76.3 percent in Knopf's specimen of alaskite. Except for minor variations, the curves of the oxides appear regular. As SiO_2 increases, Al_2O_3, CaO, MgO, FeO and Fe_2O_3 all decrease. Na_2O is nearly constant below 70 percent of SiO_2, but increases slightly above 70 percent, and a singles analysis suggests that it rises sharply above 75 percent. K_2O increases regularly with SiO_2 to 70 percent of SiO_2, and is nearly constant above 70 percent of SiO_2.

Normative quartz, orthoclase, and plagioclase (albite plus anorthite) calculated to 100 percent are plotted on a triangular quartz-orthoclase-plagioclase diagram in figure 14. On this diagram the norms fall in a narrow band that extends from the center of the diagram toward the plagioclase corner. This pattern shows that the ratio between normative quartz and orthoclase is essentially constant and is very close to 1:1, and that quartz plus orthoclase is inversely proportional to plagioclase.

In figure 15 the modal averages of quartz, K-feldspar (and perthite) and plagioclase for different plutons and masses in the east-central Sierra Nevada have been calculated to 100 percent and plotted on a triangular diagram. In a general way the plot of modes resembles the plot of norms (fig. 14) except that the field of modes extends farther away from the plagioclase corner, past the center of the triangle, and that the long axis of the field of modes is slightly inclined to that of the norms.

The differences in the fields of norms and modes probably result chiefly from two factors: (a) crystallization of increasingly larger amounts of biotite in the rocks toward the plagioclase corner, and (b) increasingly larger amounts of albite in K-feldspar in rocks away from the plagioclase corner. Crystallization of biotite removes K_2O from the magma and results in less modal K-feldspar than normative orthoclase; because biotite increases in abundance with plagioclase toward the plagioclase corner, the modal field is shifted progressively away from orthoclase as compared with the normative field. Solid solution of albite in K-feldspar (as well as perthitic albite) increases the amount of modal K-feldspar over normative orthoclase. The amount of albite in K-feldspar is greatest in rocks with the least plagioclase because these rocks contain the most K-feldspar and because the K-feldspar is commonly perthitic. The result is that at the end of the field of modes most distant from plagioclase, the field is expanded away from plagioclase and toward K-feldspar.

Generally quartz and K-feldspar are equally abundant in all the granitic rocks, and they decrease systematically with increasing plagioclase. Biotite and hornblende increase with plagioclase, and biotite, is, on the average, about 4 percent more abundant than hornblende. A notable exception is the Tinemaha granodiorite, which contains more hornblende than biotite. Hornblende generally is present only in trace amounts in rocks that contain less than 37 percent plagioclase although a few masses of rock with less plagioclase contain several percent hornblende. Thirty-seven percent modal plagioclase falls just below the middle of the quartz monzonite field.

TABLE 3.—*Chemical analyses and norms of granitic rocks from the east-central Sierra Nevada*

	Sample numbers												
	1	2	3	4	5	6	7	8	9	10	11	12	13
Chemical analyses													
SiO_2	59.0	61.0	62.0	62.8	62.82	63.53	64.86	65.2	65.1	65.7	66.7	66.92	67.4
Al_2O_3	17.2	16.06	16.6	17.0	15.44	15.61	16.12	16.7	16.6	16.7	16.1	15.19	15.8
Fe_2O_3	2.0	1.86	2.1	2.4	2.59	2.35	1.90	1.9	2.3	1.8	2.1	1.45	1.7
FeO	4.6	4.06	3.4	2.8	3.17	3.25	2.52	2.2	2.4	2.4	2.2	2.52	2.2
MgO	2.8	3.10	2.2	1.5	2.35	2.54	1.55	1.6	1.8	1.2	1.4	1.74	1.3
CaO	6.2	5.46	4.5	3.8	5.04	4.58	3.80	4.4	4.1	4.0	3.8	3.79	3.1
Na_2O	3.3	3.45	3.7	4.8	3.15	3.31	3.44	3.7	3.4	4.0	3.6	3.16	3.5
K_2O	2.1	2.95	3.7	3.2	3.72	2.98	4.03	3.1	3.2	3.0	3.1	3.82	4.2
H_2O-	1.0	.05	.79	.62	.03	.04	.06	.63	.48	.92	.69	.06	.48
H_2O+		.58			.62	.61	.51					.48	
TiO_2	.82	.88	.77	.76	.64	.63	.57	.58	.48	.49	.48	.47	.42
CO_2	.05	.0	.19	.12	.01	.03	.0	.05	.05	.16	.08	.02	.05
P_2O_5	.34	.25	.21	.26	.30	.23	.23	.18	.19	.18	.18	.18	.14
MnO	.14	.10	.10	.15	.11	.12	.09	.09	.09	.10	.10	.08	.12
BaO	------	------	------	------	------	------	.13	------	------	------	------	------	------
Total	100.0	99.80	100.0	100.0	99.99	99.81	99.81	100.0	100.0	101.0	101.0	99.88	100.0
Quantitative spectrographic analyses for minor elements													
B	------	0.004	------	------	0.001	0.0	0.0	------	------	------	------	0.0	------
Ba	------	.1	------	------	.2	.2	.2	------	------	------	------	.1	------
Co	------	.002	------	------	.002	.001	.0009	------	------	------	------	.001	------
Cr	------	.004	------	------	.001	.002	.001	------	------	------	------	.001	------
Cu	------	.002	------	------	.001	.0009	.001	------	------	------	------	.0006	------
Ga	------	.002	------	------	.002	.002	.001	------	------	------	------	.001	------
La	------	.006	------	------	.01	.009	.009	------	------	------	------	.009	------
Nb	------	.002	------	------	.002	.002	.002	------	------	------	------	.002	------
Ni	------	.003	------	------	.0008	.0007	.0004	------	------	------	------	.0005	------
Pb	------	.002	------	------	.002	.002	.002	------	------	------	------	.002	------
Sc	------	.001	------	------	.001	.001	.0007	------	------	------	------	.001	------
Sr	------	.09	------	------	.09	.09	.06	------	------	------	------	.07	------
V	------	.01	------	------	.01	.009	.007	------	------	------	------	.006	------
Y	------	.003	------	------	.005	.003	.005	------	------	------	------	.004	------
Yb	------	.0003	------	------	.0004	.0002	.0003	------	------	------	------	.0002	------
Zn	------	.01	------	------	.0	.01	.0	------	------	------	------	.0	------
Zr	------	.01	------	------	.02	.01	.02	------	------	------	------	.01	------
Norms													
Quartz	13.30	13.32	13.1	12.6	16.69	15.84	18.90	19.9	20.8	19.9	23.3	22.62	21.8
Orthoclase	12.29	17.27	21.7	18.9	22.20	17.76	23.91	18.4	18.9	17.8	18.4	22.76	25.0
Albite	27.84	29.32	31.4	40.4	26.71	32.42	28.82	31.4	28.8	34.1	30.4	26.72	29.3
Anorthite	25.95	19.46	17.8	15.6	16.68	16.13	16.68	19.7	20.3	18.6	18.1	15.84	14.5
Diopside	2.51	4.73	2.9	1.4	5.16	5.44	.36	1.1	------	.1	------	1.83	------
Hypersthene	11.67	10.0	7.2	5.0	6.18	6.55	5.91	5.2	6.2	5.0	5.1	6.18	5.3
Magnetite	2.78	2.78	3.0	3.5	3.74	3.43	2.78	2.8	3.3	2.6	1.6	2.09	2.6
Ilmenite	1.52	1.67	1.5	1.5	1.22	1.21	1.07	1.2	.9	.9	.9	.90	.8
Apatite	.64	.64	.3	.7	.69	------	.62	.3	.9	.3	.3	.35	.3
Corundum	------	------	------	------	------	------	------	------	.1	------	.2	------	.2

	Sample numbers													
	14	15	16	17	18	19	20	21	22	23	24	25	26	27
Chemical analyses														
SiO_2	68.0	68.9	68.9	69.60	71.0	71.42	71.42	71.8	73.0	73.0	73.9	74.11	75.4	76.28
Al_2O_3	15.6	16.1	15.5	14.89	15.7	14.47	14.03	15.3	15.2	14.6	14.5	13.7	13.3	------
Fe_2O_3	1.8	1.6	1.6	1.07	.9	1.03	.89	1.0	.4	.9	.7	.60	.3	------
FeO	1.8	1.6	1.7	1.99	.81	1.38	1.63	.65	.34	1.0	1.1	.88	.74	------
MgO	1.5	.88	.93	.91	.39	.78	.70	.34	.16	.59	.26	.32	.12	------
CaO	3.6	3.2	3.2	2.70	1.6	2.86	1.91	1.8	1.1	.71	1.2	1.29	.48	.47
Na_2O	3.4	3.9	3.8	3.18	3.8	3.44	2.86	3.8	3.8	4.5	3.3	3.44	4.1	4.72
K_2O	3.4	3.4	3.4	4.45	5.0	3.69	5.35	4.1	5.0	4.9	4.9	4.92	4.5	4.73
H_2O-	.40	.50	.37	.08	.69	.06	.08	.48	.48	.27	.41	.12	.46	------
H_2O+				.31		.21	.35					.18		
TiO_2	.38	.42	.43	.42	.22	.25	.36	.22	.09	.21	.15	.18	.10	------
CO_2	.05	.03	.10	.01	.07	.03	.02	.08	.10	.08	.14	.01	.11	------
P_2O_5	.15	.14	.14	.12	.05	.10	.09	.05	.05	.02	.01	.06	.01	------
MnO	.08	.10	.09	.07	.04	.08	.05	.06	.06	.08	.06	.05	.08	------
BaO	------	------	------	------	------	------	------	------	------	------	------	------	------	------
Total	100.0	100.0	100.0	99.80	100.0	99.80	99.74	100.0	100.0	101.0	101.0	99.89	100.0	------
Quantitative spectrographic analyses for minor elements														
B	------	------	------	0.0	------	0.0	0.0	------	------	------	------	0.0	------	------
Ba	------	------	------	.2	------	.2	.1	------	------	------	------	.1	------	------
Co	------	------	------	.0005	------	.0004	.0005	------	------	------	------	.0001	------	------
Cr	------	------	------	.0005	------	.0003	.0006	------	------	------	------	.0003	------	------
Cu	------	------	------	.0008	------	.0003	.0004	------	------	------	------	.0002	------	------
Ga	------	------	------	.001	------	.001	.001	------	------	------	------	.001	------	------
La	------	------	------	.01	------	.01	.008	------	------	------	------	.005	------	------
Nb	------	------	------	.002	------	.002	.002	------	------	------	------	.001	------	------
Ni	------	------	------	.0002	------	.0001	.0003	------	------	------	------	.0001	------	------
Pb	------	------	------	.002	------	.001	.002	------	------	------	------	.002	------	------
Sc	------	------	------	.001	------	.001	.001	------	------	------	------	.0007	------	------
Sr	------	------	------	.03	------	.05	.03	------	------	------	------	.04	------	------
V	------	------	------	.005	------	.005	.004	------	------	------	------	.001	------	------
Y	------	------	------	.005	------	.003	.004	------	------	------	------	.002	------	------
Yb	------	------	------	.0004	------	.0002	.0003	------	------	------	------	.0001	------	------
Zn	------	------	------	.0	------	.0	.0	------	------	------	------	.0	------	------
Zr	------	------	------	.02	------	.01	.05	------	------	------	------	.03	------	------
Norms														
Quartz	24.8	24.5	25.6	22.52	25.68	29.94	28.38	29.64	29.22	25.2	32.6	31.92	32.58	29.76
Orthoclase	20.0	20.0	20.0	26.13	29.47	21.71	31.68	24.46	29.47	28.9	28.9	28.94	26.69	27.24
Albite	28.8	33.0	32.0	26.72	31.96	28.45	24.11	31.96	31.96	38.3	27.8	28.85	34.58	39.82
Anorthite	17.2	15.0	15.0	12.51	8.06	13.35	9.46	8.90	5.56	3.6	5.8	6.40	2.50	2.50
Diopside	.5	------	------	------	------	------	------	------	------	------	------	------	------	------
Hypersthene	4.8	3.1	3.5	4.53	1.26	3.43	3.55	.90	.66	2.2	1.8	1.72	1.36	------
Magnetite	2.6	2.3	3.2	1.57	1.39	1.46	1.39	1.39	.70	1.4	.9	.89	.46	------
Ilmenite	.8	.8	.8	.78	.46	.47	.65	.46	------	.5	.5	.32	------	------
Apatite	.8	.3	.3	.29	------	.27	------	------	------	------	------	------	------	------
Corundum	------	.5	.1	.31	1.12	------	.10	1.33	1.53	.5	1.6	.51	.71	------

1. Quartz diorite. Rapid rock analysis by H. F. Phillips, P. L. D. Elmore, and K. E. White; lat 37°20′ N.; long 118°41′ W.
2. Inconsolable granodiorite. Standard rock analysis by L. M. Kehl; lat 37°7′ N.; long 118°32′ W.
3. Coarse grained granodiorite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°54′ N.; long 118°32′ W.
4. Quartz-poor granodiorite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°46′ N.; long 118°19′ W.
5. Tinemaha granodiorite. Standard rock analysis by L. M. Kehl; lat 37°4′ N.; long 118°25′ W.
6. Round Valley Peak granodiorite. Standard rock analysis by L. M. Kehl; lat 37°27′ N.; long 118°41′ W.
7. Granodiorite of McMurry Meadows. Standard rock analysis by L. M. Kehl; lat 37°6′ N.; long 118°22′ W.
8. Coarse grained granodiorite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°54′ N.; long 118°28′ W.
9. Round Valley Peak granodiorite. Rapid rock analyses by K. E. White, P. L. D. Elmore, and P. W. Scott; lat 37°31′ N.; long 118°55′ W.
10. Quartz monzonite. Rapid rock analyses by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°54′ N.; long 118°21′ W.
11. Granodiorite. Rapid rock analyses by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°50′ N.; long 118°29′ W.
12. Lamarck granodiorite. Standard rock analysis by L. M. Kehl; lat 37°9′ N.; long 118°36′ W.
13. Tinemaha granodiorite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°54′ N.; long 118°24′ W.
14. Round Valley Peak granodiorite. Rapid rock analysis by K. E. White, P. L. D. Elmore, P. W. Scott; lat 37°31′ N.; long 118°48′ W.
15. Lamarck granodiorite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°55′ N.; long 118°23′ W.
16. Fine-grained granodiorite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°50′ N.; long 118°28′ W.
17. Tungsten Hills quartz monzonite. Standard rock analysis by L. M. Kehl; lat 37°31′ N.; long 118°34′ W.
18. Alaskite of Cathedral Peak type. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, and M. D. Mack; lat 37°14′ N.; long 118°24′ W.
19. Wheeler Crest quartz monzonite. Standard rock analysis by L. M. Kehl; lat 37°25′ N.; long 118°42′ W.
20. Tungsten Hills quartz monzonite. Standard rock analysis by L. M. Kehl; lat 37°22′ N.; long 118°43′ W.
21. Quartz monzonite of Cathedral Peak type. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, and M. D. Mack; lat 37°21′ N.; long 118°45′ W.
22. Quartz monzonite of Cathedral Peak type. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, and M. D. Mack; lat 37°16′ N., long 118°39′ W.
23. Alaskite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°49′ N.; long 118°26′ W.
24. Quartz monzonite. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, M. D. Mack, and H. H. Thomas; lat 36°57′ N.; long 118°33′ W.
25. Quartz monzonite of Cathedral Peak type. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, and M. D. Mack; lat 37°7′ N.; long 118°29′ W.
26. Alaskite of Cathedral Peak type. Rapid rock analysis by P. L. D. Elmore, S. D. Botts, and M. D. Mack; lat 37°12′ N.; long 118°25′ W.
27. Alaskite of Cathedral Peak type. Partial analysis reported by Knopf (1918, p. 68); lat 37°17′ N.; long 118°25′ W.

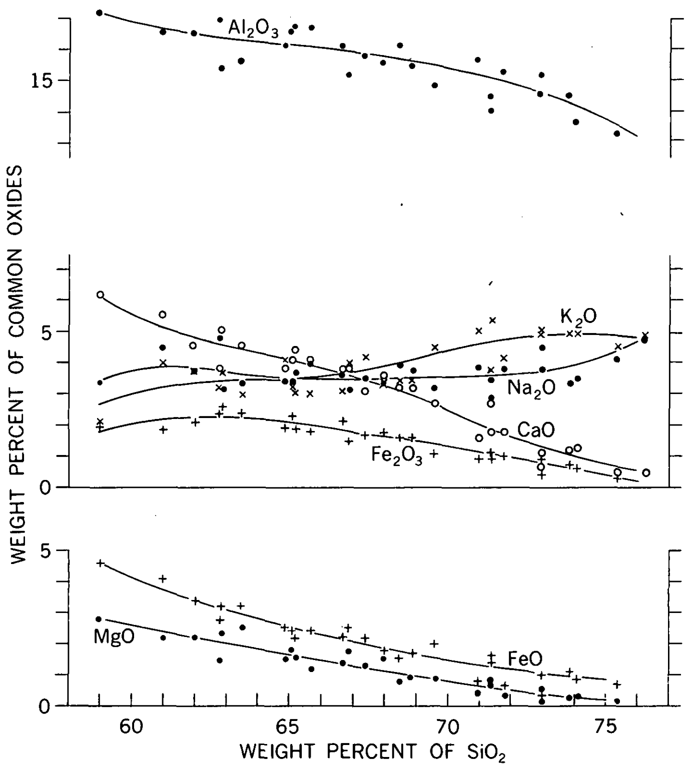


FIGURE 13.—Variation diagram of common oxides in granitic rocks of the east-central Sierra Nevada plotted against SiO_2.

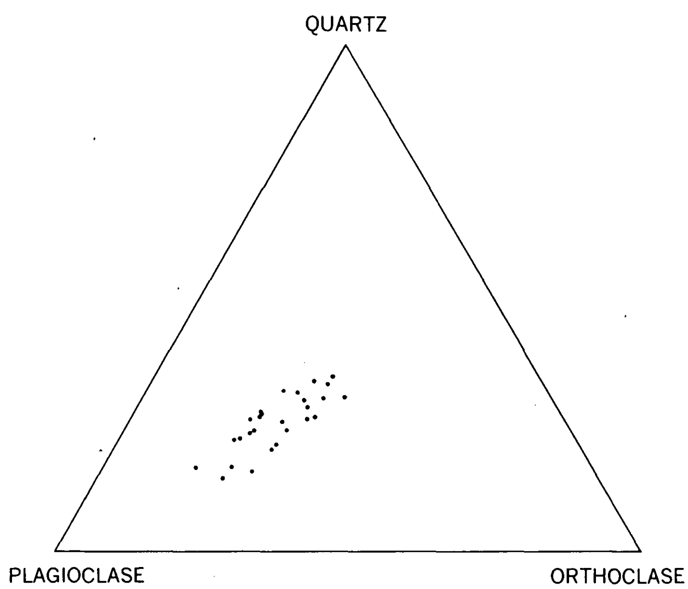


FIGURE 14.—Plot of norms (listed on table 3), of granitic rocks from the east-central Sierra Nevada.

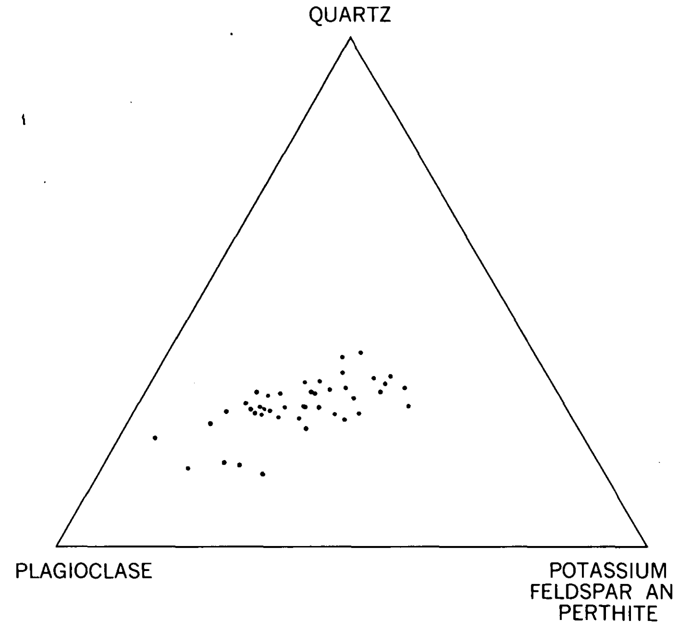


FIGURE 15.—Averages of model analyses of granitic intrusions of the east-central Sierra Nevada; the points represent a total of 597 modes.

ZONING WITHIN PLUTONS

Some masses of granitic rock show obvious compositional zoning, and many others that appear uniform are found on careful study to show systematic variation in composition from place to place. Zoning is identifiable in the field by means of variations in the content of dark minerals, generally in conjunction with changes in average grain size or in texture. The simplest and probably best way to identify zoning in the laboratory is to measure the specific gravities of systematically collected hand specimens. This method is quite sensitive and often reveals zoning where the differences in dark mineral content are too small for the zoning to be obvious in the field. Zoning is also shown by variation in the content of plagioclase, K-feldspar, or quartz, and in the anorthite content of plagioclase; it may be either lateral or concentric.

Lateral zoning is recognizable chiefly in the larger intrusives. Thus, the narrow southern part of the Lamarck granodiorite is notably lighter in color, lower in specific gravity, and less calcic than the thicker northern part. Another example of lateral zoning is the pluton of Tungsten Hills quartz monzonite in the southwestern part of the Mount Tom quadrangle, which grades toward the west from biotite quartz monzonite to hornblende-biotite granodiorites.

Concentrically zoned plutons are more felsic in their cores and more mafic in their margins. The granodiorite of Cartridge Pass in the Mount Pinchot quadrangle provides one of the best examples. The rock in the core is porphyritic quartz monzonite, and its dark mineral content is less than 6 percent; toward the margins the rock is darker and at the contact it is equigranular granodiorite and its dark mineral content is more than 12 percent. In figure 16 the zonal pattern in the Cartridge pluton is shown in terms of the content of dark minerals, specific gravity, the percent of anorthite in plagioclase, and the percentages of quartz, K-feldspar, and plagioclase. In most places the change in composition and texture from the core to margin of the pluton is gradational, but in the east side light-colored core rock is intrusive into darker marginal rock.

In most intrusive rocks zoning probably reflects differentiation during cooling, but in some masses contamination by wall rocks has been the cause of compositional differences. The interpretation as to whether compositional variation is due to differentiation or contamination usually depends upon the presence or absence of field evidence of contamination, and where there is no evidence of contamination, the variation is interpreted as an effect of differentiation. In places where contamination has been effective, the contaminating rock generally is either early mafic intrusive rock or mafic volcanic rock, but lime-rich wallrock also can affect the composition of a magma. In a granitic intrusive, early formed crystals of plagioclase, hornblende, and the accessory minerals are presumed to settle or to accrete to the walls. Convection could be effective in the marginal accretion of crystals by bringing crystals into contact with the walls, but whether convection takes place in granitic magmas is not known.

A possible contributing process may be that of thermodiffusion. At one time the Soret effect was thought to be a principal cause of differentiation in magmas, but it has been in disfavor for many years because of the presumed slow velocity of diffusion in viscous silicate melts. Wahl (1946) is one of the few advocates in recent years of diffusion in magmatic differentiation. Our concept of the kind of diffusion that might operate is that of gradual shift of calcium and the femic constituents outward from the core of a magma body in response to impoverishment of these constituents in the liquid phase in the margins as a result of crystallization. Even though the rate of diffusion may be very slow—perhaps only a few centimeters a year—over the period of crystallization of

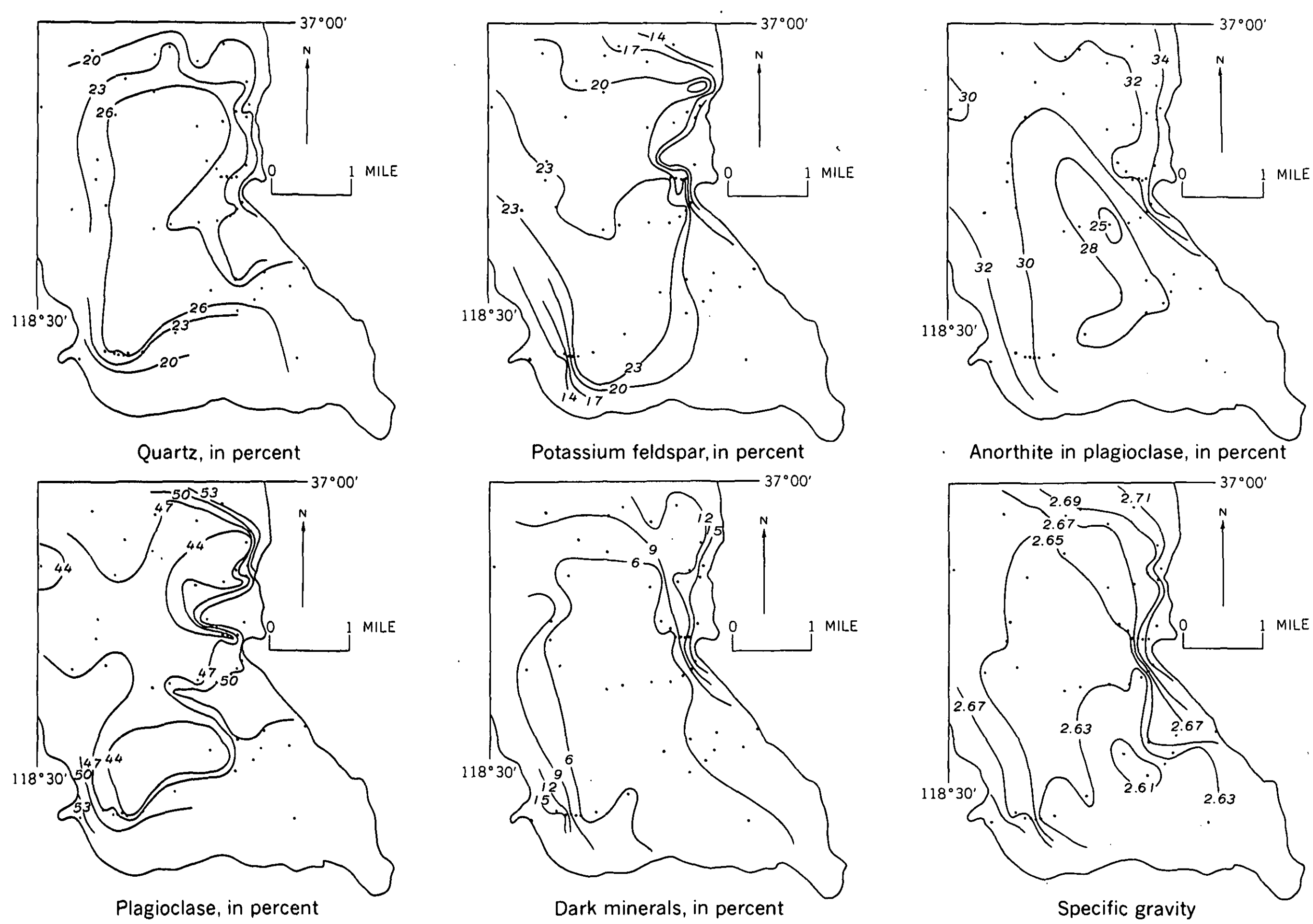


FIGURE 16.—Concentric zoning in the grandiorite of Cartridge Pass shown in different ways. Contours are based on 49 specimens whose locations are shown by dots. Compare with figure 3.

a pluton, which may extend over hundreds of thousands of years, diffusion could play a significant role in differentiation.

Both crystal fractionation and thermodiffusion permit formation of rock in the margins of an intrusion that is more mafic than the magma, and can account for a concentrically zoned pluton. The marginal part of a zoned intrusion does not necessarily represent the original composition of the magma, even if it is fine grained as compared with the interior of the intrusion (unless it is a glass). On the contrary, if it can be assumed that either or both convection and high viscosity inhibited the sinking of significant amounts of crystals, the initial composition of the magma will be represented best by the average composition of the exposed rocks.

Lateral changes in composition from one side of an intrusive to the other can be explained by assuming that after partial differentiation and solidification of the margins, renewed movements of the still liquid magma in the core took place. Abrupt contacts would result from such movements only if the moving magma came into contact with rather completely crystalized parts; otherwise, gradational contacts should result. An excellent example of a concentrically zoned intrusive in which the residual core magma moved repeatedly is to the White Creek batholith in British Columbia (Reesor, 1958). The intrusive contract in the east side of the granodiorite of Cartridge Pass doubtless reflects movement of the core magma.

CHANGES ACROSS THE BATHOLITH

In addition to chemical and mineralogic changes within plutons, broad changes take place across the batholith. In general the granitic rocks are more mafic toward the west and more silicic toward the east, but this is a gross trend and some silicic plutons are found in the western part and some mafic plutons in the eastern part.

In the discussion of zoning within plutons it was shown that changes in mineral content are accompanied by changes in specific gravity, and that specific gravity is, in a general way, an index to composition. Variations in specific gravity probably chiefly reflect variations in the content of mafic minerals and in plagioclase composition, but also may reflect, to a lesser degree, variations in the amount of K-feldspar. The specific gravities of quartz and oligoclase are close to that of average granitic rock, and variations in their amounts probably has little effect on specific gravity. The specific gravities of many hand specimens from the eastern Sierra Nevada and a belt that extends to the west as far as Shaver Lake have been determined. To show the variations in specific gravity across the batholith, the specific gravities of samples within 10 miles on either side of a line (*D–D′*, pl. 1) extending from Big Creek, south of Huntington Lake, to Owens Valley near Bishop have been projected onto a graph drawn along the line (fig. 17).

These data show a progressive increase in the average specific gravity from east to west; the average

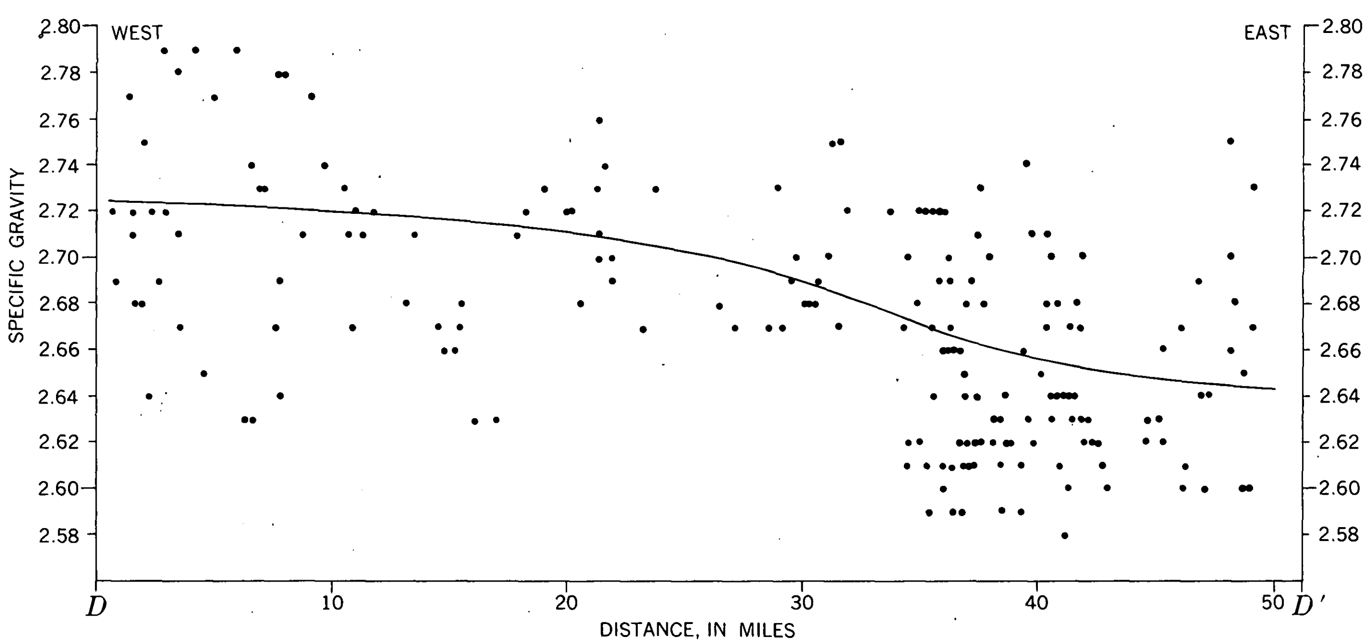


FIGURE 17.—Variation in specific gravity of granitic rocks along line *D′–D′* on plate 1 across the central Sierra Nevada from Big Creek to Owens Valley near Bishop. Each dot represents a measured sample. Points are projected up to 10 miles from either side of line.

specific gravity at the east end of the line is 2.64 and at the west end 2.72, an increase of 0.08. This increase probably is conservatixe because the west end of the line terminates in a group of small relatively silicic intrusives that lie between the Mount Givens granodiorite and the granodiorite of "Dinkey Creek" type. The average specific gravity of rocks similar to the Cathedral Peak granite, the least dense large group of granitic rocks in the eastern Sierra Nevada, is about 2.60, whereas the average specific gravity of samples of granodiorite of "Dinkey Creek" type collected between Huntington and Florence Lakes is 2.77, a difference of 0.17. The length of line D'–D' is about 50 miles; its west end falls 25 miles short of extending to the western limit of exposure of batholithic rocks. If the average specific gravity continues to increase westward, the difference in average specific gravity across the full width of the batholith may be 0.15 to 0.20.

Aside from its importance in demonstrating broad chemical changes across the batholith, the specific gravity change is important in interpreting gravity measurements. The batholith should not be treated as a mass of uniform specific gravity, but as a mosaic of different intrusives whose specific gravities are, on the average, higher toward the west.

The changes in the bulk chemistry to the granitic rocks across the Sierra Nevada appear to be part of a much broader regional pattern. All along the Pacific Coast from Alaska to Mexico the silica content and the ratio of potassium to silicon in the granitic rocks appears to increase eastward (Moore, 1959; Moore and others, 1961). The underlying cause of these changes is still little understood.

CORRELATION OF NORMATIVE COMPOSITIONS OF THE GRANITIC ROCKS WITH EXPERIMENTAL DATA

Since granitic rocks consist chiefly of quartz and feldspar, they can be approximated experimentally by mixtures of SiO_2, $KAlSi_3O_8$, $NaAlSi_3O_8$, and $CaAl_2Si_2O_8$, hereafter referred to as Qz, Or, Ab, and An. In the east-central Sierra Nevada the content of normative feldspar and quartz ranges from 79.4 in a specimen of quartz diorite to 99.3 in a specimen of alaskite. A considerable amount of experimental work has been carried on with such artificial mixtures, particularly by Bowen, Schairer, and Tuttle at the Geophysical Laboratory of the Carnegie Institution of Washington, D.C. These three workers, together with Franco, have experimentally studied the liquidus relations at atmospheric pressure in all of the binary and ternary combinations of this four-component system. The ternary system Qz-Or-Ab was studied by Bowen (1937), the system Qz-Or-An by Schairer and Bowen (1947), the system Or-Ab-An by Franco and Shairer (1951), and the system Qz-Ab-An by Schairer (written communication, 1957). Bowen and Tuttle have also determined certain liquidus relations in systems that include H_2O under pressure (Bowen, 1954; Tuttle and Bowen, 1958). Yoder, Stewart, and Smith (1957) have studied the system Or-Ab-An-H_2O at 5,000 bars, and Stewart (1958) has studied the system Qz-An-H_2O at various pressures.

The liquidus relations in the system Or-Ab-An-Qz-H_2O at 5,000 bars H_2O pressure are shown in figure 18. In figure 18-*A* projections of norms of the grantitic rocks from the east-central Sierra Nevada are shown on the faces of the tetrahedron. The positions of the norms can be misleading if allowance is not made for the fact that they are projections. Their positions were determined by calculating to 100 percent the three constituents represented in each face. The points are plotted as they would appear within the tetrahedron if they were viewed by looking at each face with the eye at the opposite corner. As a consequence of the construction, the spread of norms on the tetrahedron faces is greater than the true field within the tetrahedron. The tetrahedron is shown in three-dimensional form in figure 18-*B*. The plagioclase field occupies most of the tetrahedron, a quartz field occupies a part of the tetrahedron near the Qz corner, and a flattish Qz-Ab feldspar field extends outward from the Or corner toward Ab and Qz. Because most of the granitic rocks contain almost equal amounts of normative quartz and orthoclase, the plane within the tetrahedron that approximates most closely the true field of the norms is one which bisects the tetrahedron as shown in figure 18-*C*. The norms are projected only short distances to this plane and appear in very nearly their true positions. The plane intersects the boundary between the plagioclase and quartz fields at a small angle, and does not intersect the Or-Ab feldspar field at all. However, the boundary line between quartz, Or-Ab feldspar, and plagioclase converges on the section at a small angle and very nearly intersects it at the temperature minimum. Because this boundary line is so close, it is projected onto the section.

It is apparent from the section that the norms fall either within the plagioclase field, or very close to the field boundary between quartz and plagioclase, in a direction approaching the temperature minimum. In general, but not in detail, the oldest rocks are those whose norms plot highest in the diagram and nearest An, and the youngest plot closest to the temperature minimum. The pattern of norms is very close to that of a theoretical path for the composition of a differentiating liquid, and strongly supports the view that the granitic rocks in the east-central Sierra Nevada are fnndamentally of magmatic origin (Barth, 1952, fig. 20, p. 101). Magma of the composition of the norm with the most An (quartz diorite) would first crystallize calcic plagioclase at the H_2O saturated liquidus (slightly below 1000°C). Crystallization of the calcic plagioclase would cause the composition of the remaining liquid to be displaced away from An, and to a lesser degree, Ab. With falling temperatures the amount of Ab in the crystallizing plagioclase would increase, which would cause the composition of the remaining liquid to change along a curved path, convex toward Ab. On reaching the quartz or Or-Ab feldspar field boundary, a second mineral would begin to crystallize. If the quartz boundary surface were intersected first, as appears to have happened, quartz would begin to crystallize in addition to plagioclase, and the composiiton of the melt would move away from the SiO_2 corner of the tetrahedron as well as the An and Ab corners in a path that would be determined by the relative amounts of quartz and plagioclase crystallizing and by the composiiton of the plagioclase. This path would ultimately change the composition of the remaining liquid so that Or-Ab feldspar also would begin to crystallize. The liquid would then lie on the common boundary line of the quartz, plagioclase, and Or-Ab feldspar fields. Inasmuch as the path along this boundary curve is toward Ab, and because at the low temperature (about 700°C) the plagioclase crystallized would be Ab rich, much more quartz and Or-Ab feldspar would crystallize that plagioclase. At

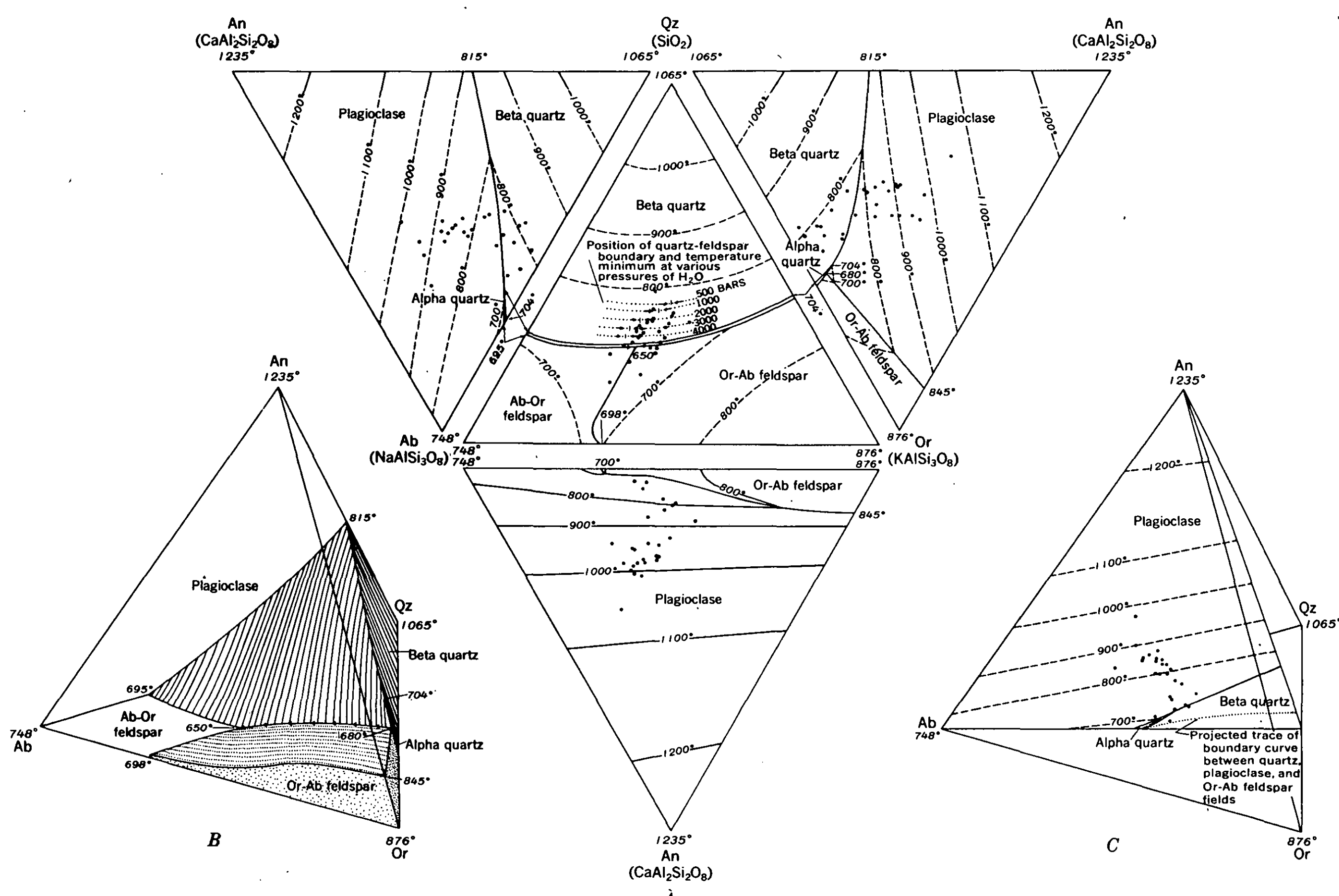


FIGURE 18.—Tetrahedron showing the liquidus relations in the system Or ($KAlSi_3O_8$)-Ab($NaAlSi_3O_8$)-An($CaAl_2Si_2O_8$)-Qz(SiO_2)-H_2O at 5,000 bars H_2O pressure. All components are in weight percent. *A*. Faces of tetrahedron showing field boundaries (heavy lines), isotherms (light lines and dashes), and projections of norms of granitic rocks from the east-central Sierra Nevada (points). Or-Ab-An face is after Yoder, Stewart, and Smith (1957). An-Qz join is after Stewart (1958). Inversion temperature of quartz is after Yoder (1950). Temperature and position of Or-Ab-Qz minimum and positions and temperautres of other field boundaries were projected by D. B. Stewart (written communication, 1958) from data of Tuttle and Bowen (1958) and H. R. Shaw (Stewart, written communication, 1960). Quartz-feldspar field boundary and temperature minimum at various H_2O pressures after Tuttle and Bowen (1958). Isotherms, except on Or-Ab-An face and along An-Qz join were constructed with reference to the above data and are approximate. *B*. Three-dimensional drawing of tetrahedron showing field boundaries. The minimum lies at 650°C. *C*. Section bisecting tetrahedron through Ab and An corners and intersecting the Or-Qz join at midpoint. On the section are plotted isotherms; field boundaries between alpha quartz, beta quartz, and plagioclase; projections of norms of rocks from the east-central Sierra Nevada; and projected trace of boundary curve between quartz, plagioclase, and Or-Ab feldspar fields.

the temperature minimum the relative amount of albitic plagioclase crystallizing probably would increase; the theoretical composition of the liquid and of the crystallizing constituents would be Ab 44.5 percent, Or 28.5 percent, and Qz 27.0 percent. However, extreme fractionation would have to occur for the minimum to be reached, and separation of a gas phase may affect the ratio of constituents in the liquid.

If no crystals were substracted from the melt during cooling and crystallization, the final rock would have the same composiion as the initial melt. However, if during crystallization some crystals were removed, the bulk composition of melt plus the remaining crystals would keep changing. If parts of this residual magma were intruded at different times, the rocks formed from them would have different compositions.

The norms of rocks with small amounts of An plot on the section (fig. 18-*C*) along the trace of the field boundary between quartz and Or-Ab feldspar at 5,000 bars of water-vapor pressure; however, it must be remembered that the norms are projected onto the section even though for only short distances, and the field boundary is cut at an unfavorable angle for accurately depicting the position of the norms with respect to this field moundary. It is therefore possible that the position of the quartz Or-Ab feldspar field boundary at some other vapor pressure would fit the distribution of norms better. The position of the boundary shifts toward the SiO_2 corner at lower pressures of H_2O, as is shown in figure 18-*A* (Bowen, 1954). However, the boundary probably does not shift much away from the Qz corner at water-vapor pressures higher than 5,000 bars because the increase in the amount of water soluble in a granitic magma at pressures above 5,000 bars is small (Tuttle and Bowen, 1958, figs. 26 and 27, p. 56 and 57).

Goranson determined the solubility of water in glass made from granite from Stone Mountain, Georgia, to be—at 800°C—about 8.9 percent at 3,000 bars, and 9.3 percent at 4,000 bars. He also concluded that at higher pressures the solubility of water in granite would rarely exceed 10 percent (Goranson, 1931). These figures are in close agreement with figures obtained by Tuttle and Bowen (1954, fig. 28, p. 58) for melts of minimum melting composition in the system Or-Ab-Qz-H_2O. In all probability an amount of H_20 equal to the amount soluble is attained only in the late stages of crystallization, after the crystallization of substantial amount of anhydrous or nearly anhydrous minerals.

The diagrams indicate that plagioclase began to crystallize in the most calcic magma represented by a point on the diagrams at about 970° and that quartz and alkali feldspars completed crystallization in the most silicic magma represented at about 650°C. However, these temperatures would be correct only if the magmas represented were saturated with H_20 throughout crystallization and if the additional components in the magma not represented on the diagrams did not affect the crystallization temperatures. Experimental evidence suggests that additional components may lower the crystallization temperatures markedly. If saturation of the melts with H_20 was achieved in the later stages of crystallization only, the first plagioclase may have crystallized at temperatures higher than 970°C.

Another possibility to be considered is that the most calcic rocks represent collections of early crystallized crystals in magma less calcic than the rock. This possibility is partly negated by the fact that the norms of these rocks plot along the theoretical path followed by a silicate melt during cooling.

If the pressure of H_2O in the closing stages of differentiation was about 5,000 bars, as the data suggest, and if the pressure of H_2O at that time was roughly equal to the rock pressure resulting from load, the depth of differentiation was at least 11 miles. The depth of final crystallization could have been less, for masses of magma from a parent body differentiating at depth could have moved higher into the crust with only local additional differentiation such as is found in compositionally zoned plutons. Nevertheless, the close agreement between the P_{H_20} inferred from the trends in the closing stages of magmatic differentiation and from the metamorphic assemblages (p. 11) suggests that crystallization and differentiation took place at roughly the same depth. We infer from these considerations that the present level of exposure probably was about 11 miles beneath the surface of exposure at the time of emplacement of the granitic rocks

COMPARISON OF COMPOSITIONAL TRENDS WITH TRENDS OF GRANITIC SUITES FROM OTHER AREAS

For comparison with suites of granitic rocks from other areas plots of norms on triangular Qz-Or-Pl (An+Ab) diagrams (fig. 19) are used because significant similarities and differences can be represented readily without resorting to oxide variation diagrams or four component Qz-Or-Ab-An diagrams. Plots of modes would be useful for making comparisons of suites of rocks, but unfortunately few suites have been analyzed with both sufficient accuracy and detail.

Figure 20 is a composite diagram of median lines through fields of norms of granitic rocks from various areas in western North America. Some median lines extend nearly to the Plagioclase corner, whereas others, including that of the east-central Sierra Nevada, terminate at considerable distances from the Pl corner. The failure to extend closer to the Pl corner results in part from a lack of chemical analyses rather than from a lack of rocks that would plot near the Pl corner. Nevertheless, the absence of analyses of plagioclase-rich rocks generally reflects a paucity of such rocks in the terrane. In the eastern Sierra Nevada, plagioclase-rich rocks are generally in small bodies of varied texture and mineral composition; consequently analyses of these rocks are not of as much significance for general purposes as those of the larger intrusive bodies, and few have been made. On the other hand, in granitic areas such as the batholith of southern California, where plagioclase-rich rocks are in large bodies and constitute a significant part of the terrane, many analyses have been made. In any event the position of the trend line is far easier to determine than the point on that line which represents the most common rock type.

Most of the median lines converge near the center of the diagram, and many also converge at the Pl corner. Between these two areas of convergence, the plots

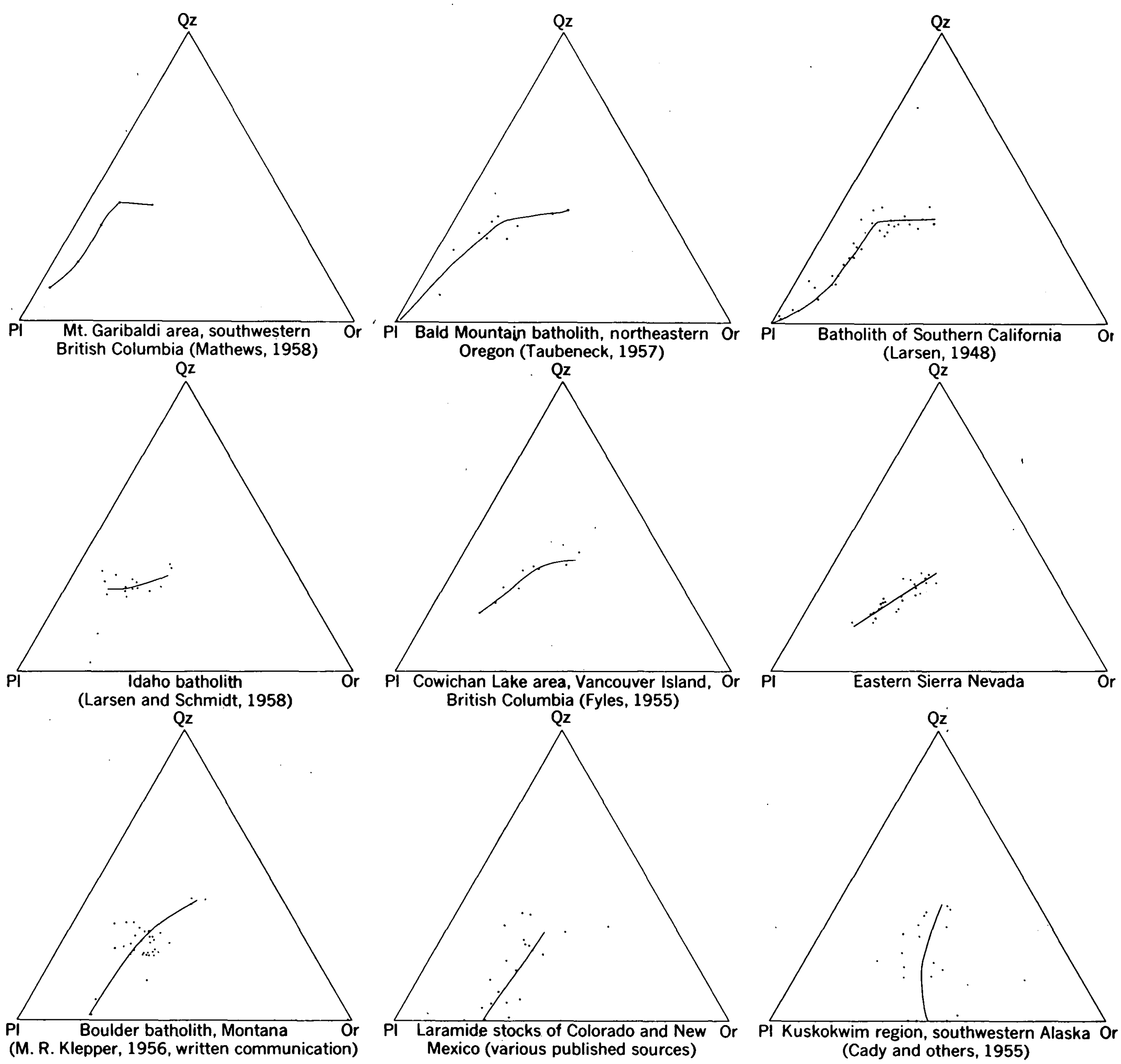


FIGURE 19.—Plots of norms for areas of granitic rocks in the western United States and Canada.

follow different trends. On the basis of these trends the plots can be categorized into three groups, although it is recognized that the groups are completely gradational. The batholith of southern California, the Bald Mountain batholith, and rocks of the Mount Garibaldi area lie along similar trends that extend away from Pl in a direction toward Qz; at about 35 to 40 percent Qz they bend toward the center of the diagram. In contrast, the Laramide stocks of Colorado and New Mexico and rocks from the Kuskokwim region, Alaska, lie along trends that extend from the Pl-Or sideline between 20 and 50 percent Or toward the center of the diagram. The third group, which includes the east-central Sierra Nevada, the Cowichan Lake area of Vancouver Island, British Columbia, and the Boulder batholith in Montana is intermediate to the other two; these plots of norms extend away from near Pl almost directly toward the center of the diagram.

The general patterns of the trends bear interesting relations to experimental data. The experimentally determined boundary curve between quartz and feldspar is shown (fig. 20) for P_{H_2O} 1,000 and 5,000 bars, since this boundary is sensitive to changes in water-vapor pressure (cf. fig. 18). The plagioclase of the rocks is, of course, not albite, but the field boundaries shown, nevertheless, are as good approximations as

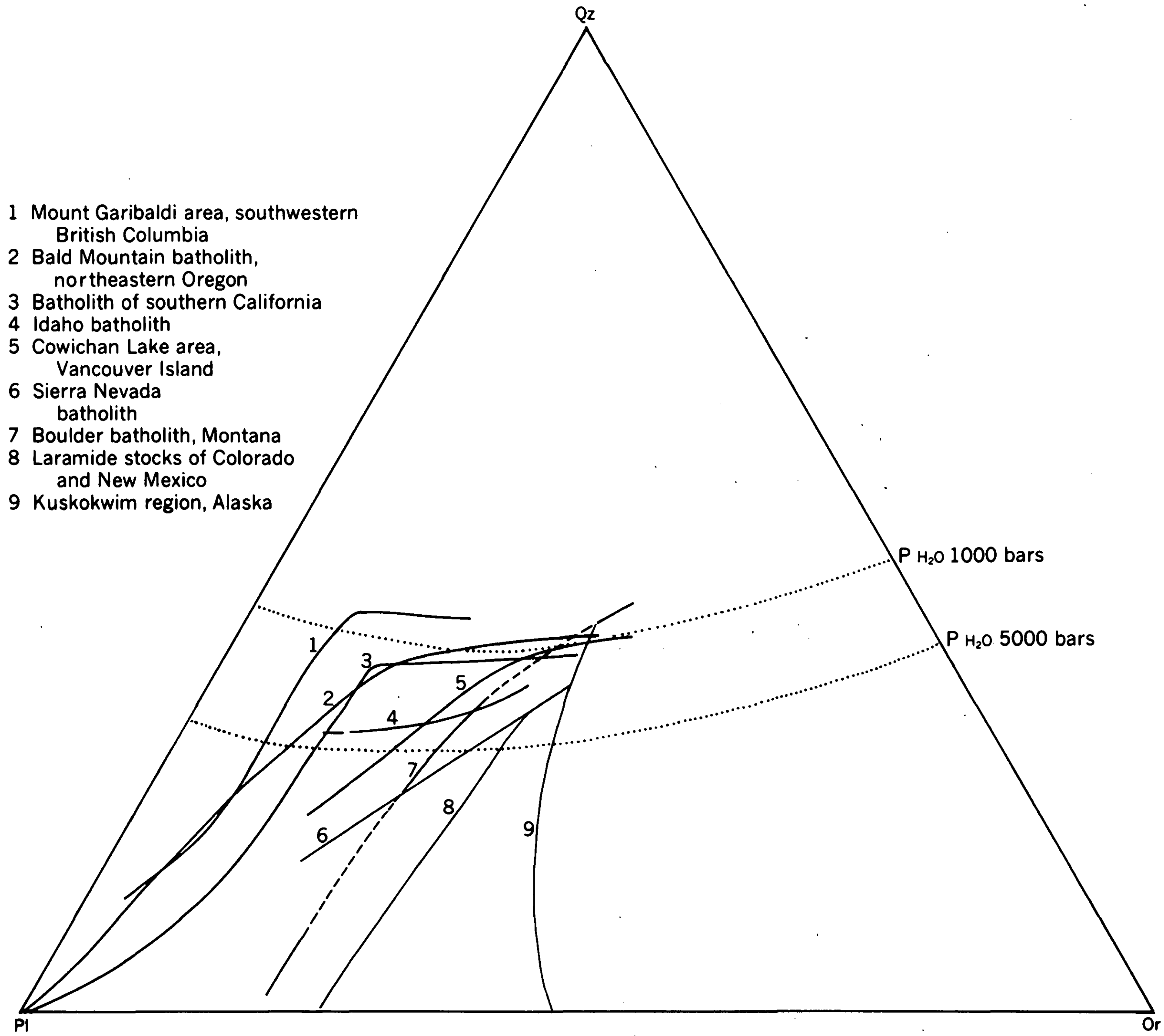


FIGURE 20.—Composite of median lines through fields of norms shown in figure 19. The experimentally determined field boundary between quartz and feldspar for the system Or-Ab-Qz-H_2O at P_{H_2O} 1,000 and 5,000 bars is shown.

can be made. By comparing figure 20 with figure 18*A* it can be seen that the norms of most suites of rocks lie either within or very close to the plagioclase (albite) field. In fact, the boundaries of the field occupied by the norm-trends in figure 20 correspond closely to the positions of the field boundaries between quartz, plagioclase, and Or-Ab feldspar in figure 18*A*.

It is known that in a general way the norms that plot closest to the Pl corner in each suite were the earliest rocks emplaced, and that those that plot near the center of the diagram were the latest, although exceptions to this generalization exist. The distributions of the norms and the sequence of emplacement of the rocks which they represent are strong arguments for considering all of the different suites to be essentially products of differentiation under conditions of general crystal-liquid equilibrium.

The different paths followed by the various plots can be explained as a result of differences in the proportion of normative quartz to orthoclase in the melt. These differences could have been original, resulting from differences in the bulk composition of the rock that was liquified, or they could have resulted from early crystallization of some such mineral as hornblende, which might have altered the proportion of normative quartz to orthoclase in the liquid. A very slight change could have a significant effect on the path of differentiation. Moore (1959) has suggested that the different trends of crystallization in the granitic rocks of different parts of the western United States are related to the composition of the initial melt, and that this initial composition is a product of the position of the magma chamber relative to the edge of the continent.

The median lines (fig. 20) of the batholiths of southern California and Bald Mountain, and rocks of the Mount Garibaldi and Cowichan Lake areas, at the ends near the center of the diagram, lie closer to the quartz-feldspar field boundary at 1,000 bars than at 5,000 bars. Smaller water-vapor pressure could reflect either smaller load pressure than that during differentiation of the eastern Sierra Nevada granitic rocks, or a deficiency of water at the same load pressure, or both.

EMPLACEMENT OF THE BATHOLITH

EVIDENCE OF MAGMA

In much of the foregoing discussions, crystallization of the granitic rocks from magma that was intruded from greater depths has been tacitly assumed. The following lines of evidence taken together indicate that the major part of the granitic rocks crystallized from a melt.

1. Contacts of individual plutons with one another and with older rocks commonly are sharp, clean, and regular, and the relative age of plutons in contact can commonly be determined by several criteria with consistent results.
2. Finer grained rock is present in the marginal parts and apophyses of some plutons.
3. In areas of diverse wallrocks most plutons are either homogeneous or are compositionally zoned in patterns that bear little or no relation to the wallrocks.
4. The geometry of some dislocations of the wallrocks suggests strongly that the dislocations were caused by the forcible emplacement of magma.
5. The internal foliation in the margins of plutons parallels external contacts and results from stretching.
6. Mafic inclusions are oriented parallel to intrusive contacts rather than to structures in the wallrocks, and in many plutons are progressively less flattened and less abundant inward. These features are compatible with a magmatic origin but not with granitization.
7. The walls of aschistic dikes marginal to some plutons are dilated.
8. Granitization and assimilation effects are confined to amphibolites and other mafic wall rocks that consist chiefly of minerals earlier in Bowen's reaction series than those that crystallized in the granitic rocks. The effects are in accord with theoretical expectations of reactions between granitic magma and wall rocks.
9. The metamorphic grade of the wallrocks and of inclusions is chiefly that of Turners's (Fyfe, and others, 1958, p. 199–239) hornblende hornfels facies which is in accord with the temperatures believed to exist in the wallrocks of crystallizing granitic magmas.
10. Variations in the compositions of the granitic rocks are in accord with variations predicted from experimental studies in melts.

EVIDENCE OF MECHANICAL PROCESSES OF EMPLACEMENT

Several lines of evidence can be cited in support of mechanical emplacement of the granitic rocks. Among these are cataclastic or protoclastic effects in the margins and contiguous wallrocks of some intrusives; lenticular mafic inclusions whose shape is believed to result from stretching of the marginal parts of a pluton during growth; and swarms of gently dipping felsic dikes marginal to some plutons, whose emplacement is inferred to have involved disclocation of the country rock. These lines of evidence have all been discussed earlier. The most convincing evidence for mechancial emplacement of the granitic rocks, however, is deformation in the metamorphic wallrocks and roofrocks that can be attributed to the intrusion of magma.

In appraising deformation in the metamorphic rocks, the deformation caused by intrusion must be distinguished from earlier regional deformation. The regional deformation resulted mostly in folding, faulting, and shearing along north- to northwest-trending lines. Although the axes of large regional folds undulate, they average about horizontal. Deformation caused by intrusion, on the other hand, only coincidentally follows north- to northwest-trending lines, and most folds caused by intrusion have steep axes. This pattern of deformation is local and directly related to intrusion.

No detailed studies have been made of the structures in the western metamorphic belt in relation to the intrusive rocks, but at two places the map pattern (pl. 1) suggests deformation by the forcible emplacement of granitic rocks: along the west side of the lobe of granitic rock just east of Sonora, in the northwestern part of plate 1, and around the stock of Jurassic granitic rock that forms the Guadalupe Mountains southwest of Mariposa near the western edge of the map. At both places the strata appear to have been bowed outward around the grantic rock.

In the eastern part of the batholith several examples can be cited of wallrock deformation that we believe was caused by forcible emplacement of granitic magma. Such deformation is shown most convincingly by structures in the Bishop Creek, Pine Creek, and Mount Morrison pendants labelled on plate 1.

The principal structures in the Bishop Creek pendant, a tight north-trending anticline and a parallel syncline, were formed as a result of regional deformation. In the southern part of the pendant, steeply dipping beds are bowed westward around a tongue of quartz monzonite similar to the Cathedral Peak granite, which appears to have penetrated the pendant from the east. Along the north edge of the pendant the strata have been spread apart by apophyses of Tungsten Hills quartz monzonite. The largest apophysis penetrates the anticline and syncline, and has spread steeply dipping beds apart by about 3 miles. A syncline present at the end of the northeast lobe may be an offset segment of the syncline in the heart of the pendant. In the northwest lobe of the pendant a salient of Tungsten Hills quartz monzonite appears to have bowed gently dipping beds upward. This is one of the few places where the separation attributed to a granitic intrusive is upward rather than lateral.

In the Pine Creek pendant, the major regional structure is a north-northwestward-trending syncline. In its south end the pendant has been bent eastward into an S-shaped structure in plan probably by intrusion of the pluton of Tungsten Hills quartz monzonite on the southwest. In this S-fold, metavolcanic rocks of Mesozoic age have been pushed and faulted across the south end of the regional syncline, which is bent to the east. The S-fold may be part of an easterly bow in strata that once were continuous with the metamorphic remnants 2 miles to the south.

The largest dislocation inferred from the arrangement of metamorphic rocks is between the south end of the Mount Morrison pendant and a septum that extends north from the Pine Creek pendant. On the basis of similar lithologic sequences the strata in the Pine Creek pendant are correlated with strata of Pennsylvanian and Permian(?) age in the Mount Morrison pendant (Rinehart and others, 1959). The strata in the Pine Creek pendant strike roughly, though not precisely, in the direction of formations in the Mount Morrison pendant with which they are correlated. The strata in the Pine Creek pendant continue to the north in a discontinuous curved septum between the Wheeler Crest quartz monzonite and Round Valley Peak granodiorite. At the north end of the septum the strata are offset 8 miles to the east of probable correlative strata in the Mount Morrison pendant. Round Valley Peak grandiorite occupies the region between the north end of the discontinuous septum and the correlated strata in the south end of the Mount Morrison pendant. The overall pattern suggests that the correlated strata in the Pine Creek and Mount Morrison pendants were once connected, and that the intervening segment has been pushed eastward to form the discontinuous septum by forcible emplacement of either the Round Valley Peak granodiorite or the younger pluton of quartz monzonite similar to the Cathedral Peak granite that is west of the Round Valley Peak granodiorite. Inasmuch as the Wheeler Crest quartz monzonite was emplaced along the east side of the septum before intrusion of the Round Valley Peak granodiorite, it must also have been involved in the eastward bulging. This may explain an abundance of cataclastic structures in the Wheeler Crest quartz monzonite in this area. Forcible emplacement of the Round Valley Peak granodiorite or the quartz monzonite similar to the Cathedral Peak granite may also have caused lateral movement on several northwesterly trending faults in the eastern half of the Mount Morrison pendant.

All but one of the examples thus far cited as evidence of deformation by forcible emplacement of magma involve lateral displacement, but it is inconceivable that the roof rocks of most plutons were not lifted by amounts of at least as much as their wallrocks were pushed aside. The wallrocks of the elongate pluton of Round Valley Peak granodiorite in the center of the Mount Morrison pendant appear to have been bowed out around the pluton, but the amount of bulging is not enough to account for all the space occupied by granitic rock; if the metamorphic wallrocks were pushed together so as to eliminate the bulging, large amounts of metamorphic rock would still be missing. There is no evidence here of stoping or of incorporation of significant amounts of metamorphic rock in the granodiorite by assimilation or granitization, and the most likely explanation is that the missing strata were pushed upward.

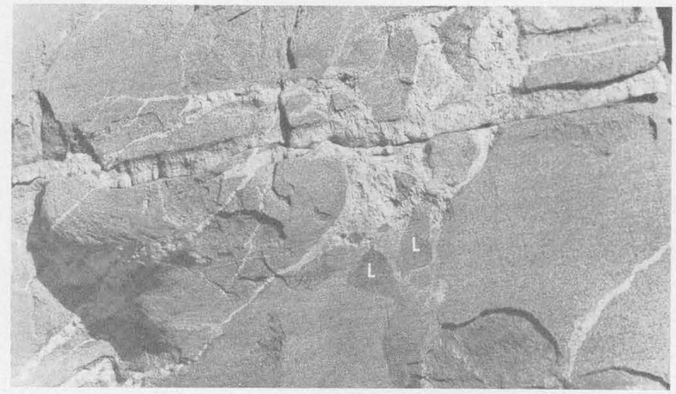
EVIDENCE FOR THERMOCHEMICAL EMPLACEMENT

Thermochemical effects include all the effects that indicate melting of wallrock and roofrock or chemical reaction between granitic rock or its magma and wallrocks or roofrocks. These effects have been variously included under granitization or assimilation. Tendencies to broaden the term "granitization" to include all reactions between magma and solid rock give the term a double or uncertain meaning, and lessen its value for precise description. In this report assimilation is used to describe the incorporation of solid rock in magma by partial reaction, solution, or melting, and granitization is reserved for the conversion essentially in the solid state of nongranitic country rock to granitic rock. Undoubtedly these processes have an area of overlap, and discrimination between them commonly is difficult or impossible. Nevertheless, theoretical distinction between the terms has some merit.

Thermochemical effects in the Sierra Nevada include two general kinds of features—mixed zones of granitic rock and wallrock in which sharp contacts predominate but in which wallrock dilation has been negligible, and hybridized granitic rocks that have been contaminated with wall or roof rock. Most of the conspicuous thermochemical effects involve mafic, generally fine grained, igneous rocks, or similar metasedimentary rocks, but in a few places pelitic metasedimentary rocks are involved. Extensive areas of sedimentary or volcanic country rock characteristized by porphyroblasts or by the local development of granitoid texture, which are features of many Precambrian terranes, are absent here.

GRANITIZATION OF MAFIC ROCK

Examples of mixed zones that result from the granitization of mafic igneous rocks are shown in figures 21, 22, and 23. The spatial relations of many irregularly shaped masses of granitic rock that penetrate mafic igneous rock indicate that the granitic rock has taken the place of mafic igneous rock and has not been emplaced by simple dilation of the walls. Small inclusions of mafic igneous rock in granitic rock could be interpreted as either residuals of chemical attack by granitic rock or its magma, and in their original positions, or as dislocated inclusions whose boundaries have been irregularly corroded chemically.



A

B

FIGURE 21.—Typical granitization effects in mafic rock. *A*. Replacement of mafic rock by aplitic material. Later aplite dike, emplaced by dilation, cuts earlier replacement aplite. Note areas of darker finer grained mafic rock (L) locally adjacent to replacement aplite. *B*. Replacement of mafic rock by medium-grained felsic granitic rock. Note irregularly embayed contact.

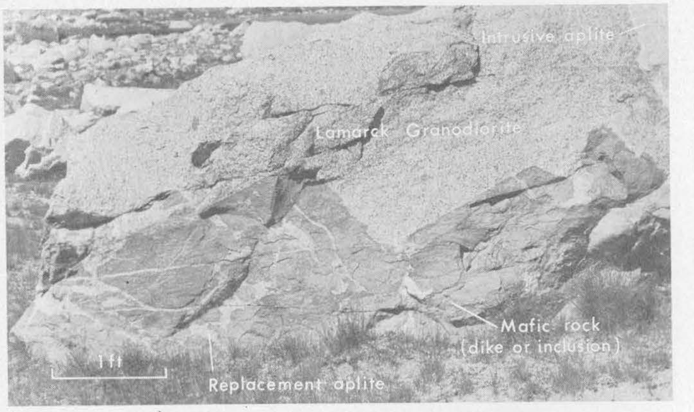


FIGURE 22.—Selective replacement of mafic rock (dike or inclusions) in preference to Lamarck granodiorite by aplitic material. Late intrusive aplite dike is in upper right-hand side of photograph.

The geometry of the walls of the early aplite shown in figure 21*A* is such that it is difficult to imagine the aplite having been emplaced forcibly by spreading of the walls, unless the mafic rock was quite soft and plastic. Darker colored, somewhat rounded areas of mafic rock adjacent to or enveloped in aplitic material are areas in which the grain size has been reduced and in which new minerals have been formed. This process is accomplished by recrystallization in conjunction with exchange of substance. Common changes are the disappearance of hornblende and plagioclase, which are the most abundant minerals in most mafic rocks, and the appearance of a fine-grained, granoblastic or schistose intergrowth of biotite, quartz, and K-feldspar. These new minerals require addition of Si and K and subtraction of Ca, Mg, and Fe. Epidote, present in places adjacent to the granitic contact, probably formed from local enrichment in CaO. Reduction of grain size is a common feature along contacts between earlier mafic rock and later granitic rock. The embayed and cuspate contacts between mafic rock and granitic rock shown in figure 21*B* are typical features of reaction contacts. The shapes of the dikelike masses of granitic rocks indicate that they replaced the mafic rocks. Careful examination of many outcrops like that shown in figure 21*B* indicated no movement in the third dimension.

Mafic dikes younger that the enclosing granitic rock and mafic inclusions in the granitic rock have been attacked and selectively replaced by aplitic material. Commonly the first signs of such replacement of a mafic body are thin stringers of aplitic material in the margins. The contact of the aplitic stringers with the enclosing granitic rock is straight and sharp, and coincides with the original contact of the mafic rock, as shown in figure 22. The contact of the aplitic material with the mafic rock, in contrast, is irregularly penetrating or conspicuously cuspate. Stringers may extend entirely across a dike or inclusion, and in places the mafic material is cut by many such stringers which may irregularly pinch and swell. Mafic dikes were observed that could be traced through segments with progressively thicker marginal stringers of aplite into a dike entirely of aplite.

Reaction between an aplite dike and an older mafic dike is illustrated in figure 23. In the lower right-hand corner of the photograph where the aplite dike cuts Tinemaha granodiorite, the positions and shapes of the dike walls indicate that they were spread apart to accomodate the dike. At the upper end where the thin dike intersects an older mafic dike, reaction has taken place within the mafic dike to produce the mixed and aplitic rock. In this reaction the granodiorite took no part, and the prominent contact across the lower part of the photograph between granodiorite and aplitic rock marks the original wall of the mafic dike.

Progressive hybridization of mafic rocks by quartz monzonite similar to Cathedral Peak granite can be seen along Big Pine Creek. The early stages of hybridization involved progressive granitization of the mafic rock through recrystallization, accompanied by interchange of substances. The later stages involved disintegration of the granitized rock at the margins and incorporation of the fragments in the granitic magma. At some distance from the mafic rock the granitic rock is even textured but has a higher content of dark minerals than uncontaminated quartz monzonite with which it is locally in sharp contact.

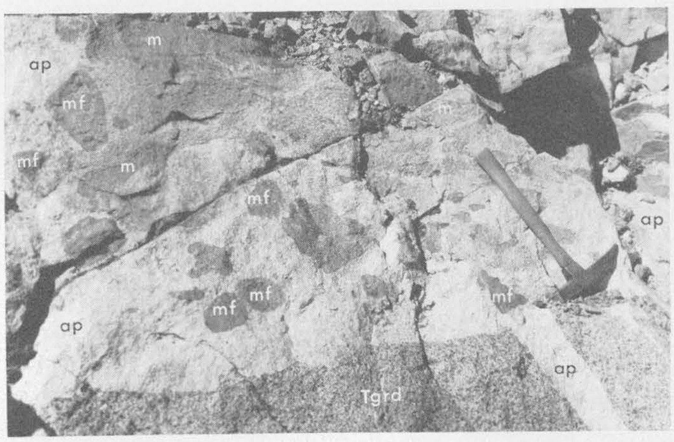


FIGURE 23.—Selective reaction of aplite dike (ap) with mafic dike (m) in Tinemaha granodiorite (Tgrd). Aplite was intruded into Tinemaha granodiorite, with which it shows no evidence of reaction. Clean straight walls and geometry of walls where aplite enters mafic dike indicate emplacement by dilation. Reaction with mafic dike is apparent. Probably both granitization and assimilation effects are represented. Note darker appearing, finer grained rounded masses of mafic rock (mf) adjacent to aplitic material.

ASSIMILATION OF MAFIC ROCK

Complementary to the process of granitization of mafic igneous and metamorphic rock is the process of assimilation of mafic rock by granitic magma. The product of assimilation is a dark-colored hybrid rock with a splotchy appearance that is caused by uneven distribution of the mafic minerals and by variations in texture. Most of the hybrid rock is granodioritic, but some is quartz dioritic. Generally the hybrid rock is darker than the rock in large intrusive masses of granodiorite. On a small scale, some dikes are progressively more contaminated with greater distance of intrusion into a mafic host rock.

Contaminated granitic rock is extensive in the south end of the Pine Creek pendant, west of the Pine Creek pendant along the north side of Pine Creek, and in Shannon Canyon. The most illustrative of these areas is a mile and a half west of the Pine Creek pendant along the north side of Pine Creek where hornblende gabbro grades laterally into granodiorite in about half a mile. Close examination, nevertheless, revealed a sharp but highly irregular intrusive contact between contaminated ganodiorite and gabbro, although at the contact the two rocks are very smiliar in appearance. The earlier hornblende gabbro does not appear to have been much affected by the granodiorite, but the gabbro obviously contaminated the later intrusive granodioritic magma.

Assimilation of the mafic material and consequent contamination of the granitic magma were probably accomplished by diffusion through volatiles or a pervasive hydrous melt of ternary minimum composition in conjunction with the physical incorporation in the magma of fragments of the wall rock. Accompanying the granitizing process of introduction of material into wallrock is the reciprocal process of removal of material (principally Ca, Mg, Fe, and probably some Al). No evidence was found to indicate that the removed material was driven ahead of the granitization into the wallrocks, and it is assumed that it moved into the magma. A difficult problem in this connection is the absence in so many places of any evidence of contamination of the granitic rock adjacent to granitized mafic rocks. Whether the material that was expelled from the wallrock traveled only far enough after reaching the magma to produce a zone of hybrid rock or whether it was dispersed so widely as to leave no recognizable trace probably was determined by the rate of dispersal after reaching the magma. Although little is known about the rate of diffusion in siliceous magma, doubtless the rate decreases as the magma cools, becomes more viscous, and crystallizes.

Marginal disintegration of the granitized or partly granitized mafic wallrock and strewing of rock or crystals through the adjacent magma may also contribute to the formation of the hybrid rock. The texture of some areas in the hybrid rock is sufficienctly similar to that of the wall rock to suggest a xenolith. Partly granitized xenoliths would, of course, be further granitized and the displaced elements diffused into the surrounding magma. The splotchy texture of the hybrid rock may be due to the partly granitized fragments of wallrock distributed through contaminated granitic rock.

ASSIMILATION OF PELITIC ROCK

Several of the granitic intrusives, notably the Wheeler Crest quartz monzonite and Tungsten Hills quartz monzonite, locally contain schistose inclusions a few inches or less in maximum dimension. These small inclusions contain abundant biotite, and probably are remnants of pelitic sedimentary rock. Wyllie and Tuttle (1960, pt. 18, p. 229–231) have determined experimentally that arkosic sediments will begin to melt at essentially the same temperature as granite, and that shale will begin to melt at only a slightly higher temperature. Further field observations are needed to establish the behavior of the pelitic rocks in granitic magma.

TENTATIVE EVALUATION OF PROCESSES INVOLVED IN EMPLACEMENT

It has been established that the Sierra Nevada batholith consists of a mosaic of separate intrusive masses and that magma was involved in the formation of these masses, although a minor amount of rock was formed by granitization. The chief problems to be considered here are the means by which space for the intrusives was provided and the relative effectiveness of forcible displacement of the country rock by magma, stoping, granitization, and assimilation. These two problems are interdependent and can be conveniently considered together.

One of the difficulties of evaluating the different processes related to the emplacement of the intrusives is that they are characteristically displayed on different scales. Neither the eye nor the camera is equally receptive to features of all sizes. Outcrops a few inches or a few feet across can be seen in considerable detail; ones a few hundred feet across, if viewed in their entirety, can be seen in much less detail, and larger features can be viewed only in a broad way and only if visibility is exceptionally good. Features associated with emplacement of the intrusive masses of about the right size for field inspection include replacement dikes and other granitization effects, zones of intrusive breccia, and dike swarms. Somewhat larger features that can be seen in the field in some places are hybridized zones in granitic rock and large foundered blocks of wallrock that are separated from an originally conjoined mass by granitic rock, but an understanding of the significance of these features commonly is best obtained from maps or diagrams. Large features that ordinarily escape the eye and which can be satisfactorily understood only after they have been represented on a map include large-scale bends or dislocations of the country rock that were caused by the forcible emplacement of intrusive magma.

Dislocations of as much as 3 miles, caused by the forcible emplacement of intrusives, have been cited for structures within individual pendants, and a dislocation of about 8 miles has been inferred between the Mount Morrison pendant and the septum that extends northeast from the Pine Creek pendant. Dislocations of such magnitudes are the most impressive effects of the emplacement of the intrusives that have been observed. On the basis of our knowledge to date, dislocation by intrusion must certainly be assigned an important role in making space for intrusive masses.

The role of stoping is difficult to assess because the boundaries of most blocks of pregranitic rock have been so modified that blocks of similar lithology cannot be related to one another by the shape of their walls. At only a few places can large blocks (ones at least a quarter of a mile long) be shown to have been split off from adjacent blocks of similar lithology. Fragmentation and piecemeal stoping is locally demonstrable in zones of intrusive breccia, but appears to be secondary to either large-scale bending or to granitization and assimilation. Nevertheless, it is probable that magma intruded with such force as to cause more than 3 miles of lateral dislocation of the wallrocks would have entered along lines of weakness and split off blocks. Once these blocks were separated they would be moved in accordance with their size and specific gravity and with the specific gravity, viscosity, and currents of the magma. Some of the exposed masses of pregranitic rock probably came from higher or lower levels than the present erosion level.

Visible traces of granitization and assimilation are quantitatively insignificant, though conspicuous because they are of optimum size for observation. Nevertheless, the end stage of granitization and assimilation is granitic rock that is virtually indistinguishable from rock that crystallized directly from magma, so that these processes are self-effacing. Granitic magma does not appear to have reacted appreciably with most metasedimentary rocks or with earlier granitic rocks. The most conspicuous reaction effects involve mafic igneous rocks and amphibolite. The possibility that large volumes of pelitic rock were incorporated in granitic magma by melting cannot be evaluated without additional fieldwork.

Time and composition are important to reaction. Dilation dikes with sharp walls generally are later features than granitization effects in the same area (figs. 21 *A* and 22), and earlier dikes generally are more strongly hybridized and have less sharply defined walls than later dikes (fig. 24). These relationships indicate that the chemical activity of granitic magma decreased with lowering temperature, and that reaction was important during the earlier stages only, when the magma was hottest.

Compton (1955) has studied an area in the western metamorphic belt where plutons intrude a terrane that is composed predominantly of mafic metavolcanic rock. For the Bald Rock batholith he estimated, from approximate projections of the country rock units through the batholith, that most of the space for the batholith was provided by forceful intrusion which forced the country rock outward, but that about a quarter of the required space was provided by stoping, assimilation, or granitization of the country rock. For the adjacent Swedes Flat pluton, which has a broad gradational contact zone with the metavolcanic country rocks, he expressed the opinion that most of the granitic rock is of replacement origin, but that the granitization was produced by fluids that emanated from a core of magma. If approximately correct, his estimates show that in terranes composed of mafic igneous rock a very significant part of the space required for

FIGURE 24.—Mafic inclusion cut by dikes of slightly different ages, all of which are offshoots from the surrounding quartz monzonite similar to Cathedral Peak granite. The earlier dikes extend from upper left to lower right. They are darker than the parent rock, apparently because of contamination. Although it seems clear that the dike walls are dilated, they are ragged and do not match perfectly across the dikes, presumably because of reaction. In contrast, later dikes, which extend from top to bottom, are uncontaminated and the walls are smooth and match perfectly.

the emplacement of an intrusive can be provided by granitization, probably in conjunction with the complementary process of assimilation.

Dislocations caused by forcible intrusion of granitic magma are most conspicuous in metasedimentary terranes, where the composition of the rocks is such that they react very slowly with the magma, or in terranes composed of igneous rocks compositionally similar to the intrusive magma. Nevertheless, there is little reason to believe that intrusions were less forcible in terranes composed of mafic igneous rocks and other rocks that react with granitic magma; however, granitization and assimilation obscure the effects of forcible intrusion in such terranes.

In summary, the most important role in the emplacement of the granitic rocks in the Sierra Nevada must be tentatively assigned to forcible intrusion, which thrust aside, and ultimately upward, the older rocks. These dislocations were accompanied locally by piecemeal stoping, and in places stoping may have accounted for substantial amounts of space now occupied by intrusive rocks. Thermochemical processes were most effective where the wallrocks were mafic igneous rocks or amphibolites, but were still probably of less importance than mechanical processes. Where volcanic rocks predominate, as in certain parts of the western Sierra Nevada, thermochemical effects may have been of greater importance in the emplacement of granitic rocks.

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U. S. GOVERNMENT PRINTING OFFICE : 1963 O - 682-196