Plutonism in the Central Part of the Sierra Nevada Batholith, California

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By PAUL C. BATEMAN

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A study of the structure, composition, and pre-Tertiary history of the Sierra Nevada batholith in the Mariposa 1° by 2° quadrangle

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By Paul C. Bateman

ABSTRACT

The Sierra Nevada batholith comprises the plutonic rocks of Mesozoic age that underlie most of the Sierra Nevada, a magnificent mountain range that originated in the Cenozoic by the westward tilting of a huge block of the Earth's crust. Scattered intrusions west of the batholith in the western metamorphic belt of the Sierra Nevada and east of the Sierra Nevada in the Benton Range and the White and Inyo Mountains are satellitic to but not strictly parts of the Sierra Nevada batholith. Nevertheless, all the plutonic rocks are related in origin.

The batholith lies along the west edge of the Paleozoic North American craton, and Paleozoic and early Mesozoic oceanic crust underlies its western margin. It was emplaced in strongly deformed but weakly metamorphosed strata ranging in age from Proterozoic to Cretaceous. Sedimentary rocks of Proterozoic and Paleozoic age crop out east of the batholith in the White and Inyo Mountains, and metamorphosed sedimentary and volcanic rocks of Paleozoic and Mesozoic age crop out west of the batholith in the western metamorphic belt. A few large and many small, generally elongate remnants of metamorphic rocks lie within the batholith. Sparse fossils from metasedimentary rocks and isotopic ages for metavolcanic rocks indicate that the metamorphic rocks in the remnants range in age from Early Cambrian to Early Cretaceous. Within the map area (the Mariposa 1° by 2° quadrangle), the bedding, cleavage, and axial surfaces of folds generally trend about N. 35° W., parallel to the long axis of the Sierra Nevada.

The country rocks comprise strongly deformed but generally coherent sequences; however, some units in the western metamorphic belt may partly consist of melanges. Most sequences are in contact with other sequences, at least for short distances, but some sequences within the batholith are bounded on one or more sides by plutonic rocks. Proterozoic and Paleozoic sedimentary strata east of the Sierra Nevada and Paleozoic strata in remnants of country rocks within the eastern part of the batholith, although strongly deformed, are autochthonous or have been displaced only short distances, whereas some Mesozoic strata in the western metamorphic belt may be allochthonous. Probably the strata in the western metamorphic belt were deposited in marginal basins and island arcs, but the possibility that they were transported from distant places has not been disproved. All the country rocks have been strongly deformed, most of them more than once. Tectonic disturbances occurred during the Devonian and Mississippian (Antler? orogeny), the Permian and (or) Early Triassic (Sonoman? orogeny), the Late Jurassic (Nevadan orogeny), and at various other times during emplacement of the batholith and uplift that accompanied and followed its emplacement. The strata in the western metamorphic belt probably were deformed in an early Mesozoic subduction complex.

The plutonic rocks range in composition from gabbro to leucogranite, but tonalite, granodiorite, and granite are the most common rock types. Most are medium to coarse grained, but some small rock masses are fine grained. Most have hypidiomorphic-granular textures and are equigranular, but some having compositions close to the boundary between granite and granodiorite contain large crystals of alkali feldspar. Serpentinized ultramafic rocks are present locally in the western metamorphic belt within and adjacent to the Melones fault zone. Except for serpentinized ultramafic rocks, trondhjemite, and most granites, all the plutonic rocks contain significant amounts of hornblende. Most of the granitoids are metaluminous or weakly peraluminous; strongly peraluminous granites are present only in the White Mountains.

Most of the granitoids are assigned to units of lithodemic rank, and most of these units are assigned to intrusive suites. Plutons assigned to the same lithodeme are composed of rock of similar composition, fabric, and age and are presumed to have been continuous at depth when they were emplaced. Intrusive suites are composed of two or more lithodemes that have similar isotopic ages and are thought to have been produced during the same magmatic episode. Lithodemes assigned to the same intrusive suite generally have common characteristics, and most show compositional and textural transitions from one to another. The lithodemes assigned to an intrusive suite commonly—but with some notable exceptions—are more felsic with decreasing age. Some intrusive suites exhibit nested patterns in which younger, more felsic rocks are enclosed in older, more mafic rocks. Except for some leucocratic granites and concentrically zoned plutons and lithodemes, the plutonic units of all ranks tend to be elongate in a northwestward direction, parallel to the dominant structures in the country rocks, and some intrusive suites only a few tens of kilometers wide extend northwestward across the map area and unknown distances beyond its north and south boundaries.

The intrusive suites show regular age patterns. The Late Triassic Scheelite Intrusive Suite crops out over a large area along and east of the eastern escarpment of the Sierra Nevada; Jurassic intrusions are present on both sides of the batholith within the map area; and voluminous Cretaceous intrusions occupy the central part of the batholith. Regionally, the axis of Cretaceous intrusion trends a little more northerly than the axis of Jurassic intrusion, and these axes cross within the map area. The locus of plutonism migrated eastward during the Cretaceous at an average rate of about 2.7 km/m.y. Nevertheless, Cretaceous plutonism was episodic, occurring within the map area chiefly about 114, 103, 98, and 90–85 Ma.

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Regular compositional changes occur across the batholith and are largely independent of the ages of the rocks. In the western foothills, tonalite is the most common rock type and is accompanied by biotite granodiorite, granite, and minor amounts of trondhjemite. Farther east, along the axis of the Sierra Nevada, hornblende-biotite granodiorite is the most common rock type and is accompanied by lesser amounts of biotite granite. In the White and Inyo Mountains, the Jurassic rock types include monzonite, quartz monzonite, granodiorite, and granite, whereas all the Cretaceous rocks are granite. The distribution of rock types is shown by average modal values. Alkali feldspar increases eastward from an average of about 4 percent in the west to about 30 percent in the eastern Sierra Nevada and the White Mountains. Plagioclase decreases irregularly from an average of a little less than 60 percent in the west to about 30 percent at the eastern base of the Sierra Nevada, then rises slightly in the White and Inyo Mountains. Total mafic minerals decrease very irregularly from an average of about 14 percent in the west to 4 percent at the eastern base of the Sierra Nevada, then rise to about 10 percent in the Jurassic rocks of the White Mountains but remain low in the Cretaceous rocks. Quartz ranges irregularly from 20 to 30 percent across most of the Sierra Nevada and the White Mountains but rises to about 35 percent in the eastern Sierra Nevada.

Chemically, K_2O increases eastward, and CaO, total Fe as FeO, and MgO decrease by smaller amounts. The alkali-lime index of the intrusive suites decreases overall eastward from about 64 in the west to about 55 in the east, reflecting the variations in K_2O, Na_2O, and CaO, but, for distinguishing intrusive suites, this index is significantly less sensitive than plots of K_2O against SiO_2 in which K_2O rises regularly eastward. Among the minor elements, U, Th, Rb, Be, Ta, Ba, and total rare-earth elements increase eastward. The oxidation ratio [mol $(2Fe_2O_3\times100)/(2Fe_2O_3+FeO)$] and initial $^{87}Sr/^{86}Sr$ increase eastward, and initial $^{143}Nd/^{144}Nd$ decreases eastward.

Most of the metavolcanic sequences in the Sierra Nevada are intruded by small hypabyssal comagmatic intrusions, but only a few metavolcanic rocks have been correlated with major intrusive suites. The most likely correlations within the map area are the mid-Cretaceous Minarets sequence with the intrusive suites of Merced Peak, Buena Vista Crest, and Washburn Lake; the Late Triassic Scheelite Intrusive Suite with the adjacent Late Triassic or Early Jurassic older succession of the Koip sequence; and the quartz monzonite of Mount Barcroft with the volcanic rocks that it intrudes, some of which have younger isotopic ages than the quartz monzonite. Easily eroded silicic tuffs could have erupted during emplacement of the voluminous late Early and early Late Cretaceous intrusions and been eroded away, but erosion does not explain the sparsity of plutonic rocks of the same age as the Early Jurassic older succession of the Koip sequence, the late Jurassic younger succession of the Koip sequence, and Early Cretaceous metavolcanic rocks.

Skarn deposits in the eastern Sierra Nevada have supplied a substantial part of the U.S. production of tungsten and significant quantities of molybdenum and copper. The Pine Creek Mine is the most productive tungsten mine in North America. Otherwise, except for by-product copper, molybdenum, silver, and gold contained in tungsten-bearing skarns and small noncommercial quartz-molybdenite veins, the batholith is almost devoid of ore deposits. The gold deposits for which the Sierra Nevada is famous are largely confined to the country rocks; the role of the batholithic intrusions was to provide the heat required to cause fluids to circulate and concentrate gold, copper, and other minerals in veins. Many K-Ar ages of vein minerals from the Mother Lode, Alleghany, and other districts of the western metamorphic belt correspond roughly with the approximate 114-Ma age of the Fine Gold Intrusive Suite, which lies adjacent to the western metamorphic belt within the map area.

The batholith was emplaced in the western margin of the North American plate, presumably as a consequence of convergence of the North American plate with plates of the Pacific Ocean. Events that occurred along an eastward-dipping subduction zone, such as shifts in its position, changes in its angle of dip, the arrival of exotic terranes, and changes in the rate of subduction, probably caused the episodic shifts in the locus of plutonism. No evidence has been recognized that indicates any intrusion was emplaced during a major compressional regional tectonic event, but some intrusions were affected by regional deformation after they were emplaced. Much of the deformation in the country rocks within and adjacent to the batholith was caused by forcible intrusion of plutonic rocks rather than by compressive regional deformation during the Nevadan orogeny. Additionally, events at the plate boundary violent enough to cause shifts in the locus of magmatism could also have deformed the country rocks. Displaced wall rocks marginal to rounded intrusions and protrusions from elongate intrusions clearly indicate ballooning during their emplacement. The presence of foliation and absence of lineation in elongate intrusions require that they also expanded as they were emplaced. Lineations are generally confined to rocks in which the magma continued to move after crystallization was far advanced or to rocks that were ductilely deformed after they had solidified.

The source magmas for the granitoids were generated in the lower crust as a result of the rise of basaltic magma and heat from the mantle. The basalt either ponded at the base of the crust and caused the generation of anatectic magmas in the overlying crustal rocks or mixed and mingled with crustal materials of variable bulk compositions to produce parent magmas having a wide range of compositions and isotopic properties. Some parent magmas were isotopically homogeneous when they rose into the upper crust, but variations in Sr and Nd isotopes within intrusions and intrusive suites indicate that either some parent magmas were isotopically inhomogeneous or that they mixed with other materials as they were emplaced. More refractory crustal materials and substantial amounts of mantle-derived magma remained in the lower crust when the magmas that crystallized to form the exposed plutonic rocks rose upward.

Possible causes of the compositional and isotopic changes that occur eastward across the batholith are (1) increasing thickness of the prebatholithic crust and, consequently, a larger crustal component in the parent magmas, (2) an increase in the sedimentary component of the crust, and (3) an increase of such constituents as potassium, uranium, and thorium in the mantle component with distance from the subduction zone. All these possibilities require further investigation for confirmation.

The parent magmas rose buoyantly, leaving refractory material in the lower crust. No inward-dipping contacts have been observed to indicate that the individual intrusions have sharply defined bottoms, and the larger masses of tonalite and granodiorite may grade downward to more mafic rocks in the lower crust. Nevertheless, intrusions with rounded outcrop patterns and tonguelike protrusions from elongate bodies, which clearly ballooned at the level now exposed, must have inward-dipping contacts at relatively shallow depths.

Mixing processes, which occurred chiefly in the lower crust, and crystal-liquid fractionation during cooling and crystallization account for the diversity of rocks that make up intrusive suites. Fractionation processes involved the partitioning of constituents such as calcium, magnesium, and iron in greater abundance in minerals crystallizing in cooler parts of the magma than in the coexisting melt and their replacement in the melt by convection of the magma, possibly aided by diffusion in the melt phase. The sidewall accretion of crystals and rise of the adjacent crystal-depleted and less dense magma to the top of the magma chamber appear to account for both horizontal and vertical compositional gradations within intrusions.

If the parent magmas had been generated entirely within the crust, vertical readjustments could have provided space for the rising magmas, and a space problem would not have existed. However, isotopic ratios, rock compositions, and uplift and erosion during the time of batholithic intrusion strongly indicate that, with the probable exception of the Cretaceous granites of the White Mountains, mantle-derived material in variable amounts was incorporated in the magmas. Space for this added material was provided by the uplift and erosion of the roof rocks and upper parts of the batholith during and following emplacement of the batholith; by the eruption of easily erodible ash-flow tuff and other pyroclastics, including airborne ash that was carried eastward by the prevailing winds; and by the spreading of the walls of the batholith. Geobarometers indicate that within the area shown on plate 1, 3.5 to 7 km of rock was eroded after the batholith was emplaced. The forceful emplacement of plutons contributed to both uplift and spreading of the batholith walls. Stoping redistributed materials within the volume now occupied by the batholith and doubtless aided in their incorporation in the magmas but provided no additional space.

INTRODUCTION

Probably no other large batholith offers conditions as favorable for geologic study as the Sierra Nevada batholith. Exposures are good almost everywhere and are superb in high glaciated areas and in the arid eastern escarpment. The western half of the Sierra Nevada is heavily forested, and the Tuolumne, Merced, and San Joaquin Rivers have cut deep, steep-walled canyons, but these are relatively minor obstacles to field study; well-maintained roads and trails make for easy access to most parts. The weather is excellent except during the winter months, when the middle and higher parts of the range are blanketed in deep snow; insects, rattlesnakes, bears, and stinging and spiny plants such as poison oak, nettles, and chinquapin are only minor nuisances.

The Sierra Nevada and the Sierra Nevada batholith are different entities. The Sierra Nevada is a magnificent mountain range that originated during the Cenozoic by westward tilting of a huge block of the Earth's crust, whereas the Sierra Nevada batholith is the part of the Sierra Nevada that is composed of almost-continuous plutonic rocks of Mesozoic age (fig. 1). Remnants of metamorphic rocks are present within the batholith, but none are large enough to interrupt its continuity. The batholith lies between older strata of the western metamorphic belt and younger sediments of the Great Valley of California on the west and the base of the eastern escarpment of the Sierra Nevada on the east. Scattered plutons west of the batholith in the western metamorphic belt and east of the Sierra Nevada in the Benton Range and the White and Inyo Mountains are satellitic to, but not strictly parts of the Sierra Nevada batholith. The extensive area composed of plutonic rocks in the White Mountains has been called the Inyo batholith (Anderson, 1937; Sylvester, Oertal, and others, 1978). Nevertheless, all the plutonic rocks are related, and some in the Benton Range and the Inyo Mountains are parts of intrusive units that have extensive outcrops in the eastern escarpment of the Sierra Nevada.

The Sierra Nevada batholith is a segment of the Mesozoic batholiths that encircle the Pacific Basin. It is continuous to the north and northeast, beneath younger sedimentary and volcanic cover, with scattered outcrops in northwestern Nevada. Southward, it terminates along the Garlock fault, but offset plutonic terranes continue southward into the Mojave Desert and the Transverse and Peninsular Ranges of southern and Baja California.

During the past 30 years, geologists of the U.S. Geological Survey (USGS) have studied all the 15-minute quadrangles within a wide area that lies across the central part of the batholith between lat 37° and 38° N. and long 118° and 120° W. (fig. 1; pl. 1), and geologic maps of most of these quadrangles have been published at a scale of 1:62,500. The principal geologic features of this area, which will be referred to henceforth as the map area, are shown in plate 1. Although only the southwest corner of the map area extends to the west edge of the Sierra Nevada, the map area includes the entire batholith where it is widest; the rocks farther west are mostly metamorphic.

The mapping of this enormous area grew out of independent studies of discrete areas or topics by individuals or small groups of individuals, most of whom spent periods ranging from a few months to many months spread over several years in the Sierra Nevada, and was not carried out as a single integrated project. Rarely were more than a few individuals actively carrying on investigations of the batholith at the same time. Geologic mapping at the 1:62,500 scale was completed in 1982, but mineralogic, chemical, isotopic, structural, and geophysical work and more detailed remapping of critical areas on larger scales can be expected to continue indefinitely as new techniques and concepts evolve. The principal purpose of this report is to provide a base for such studies.

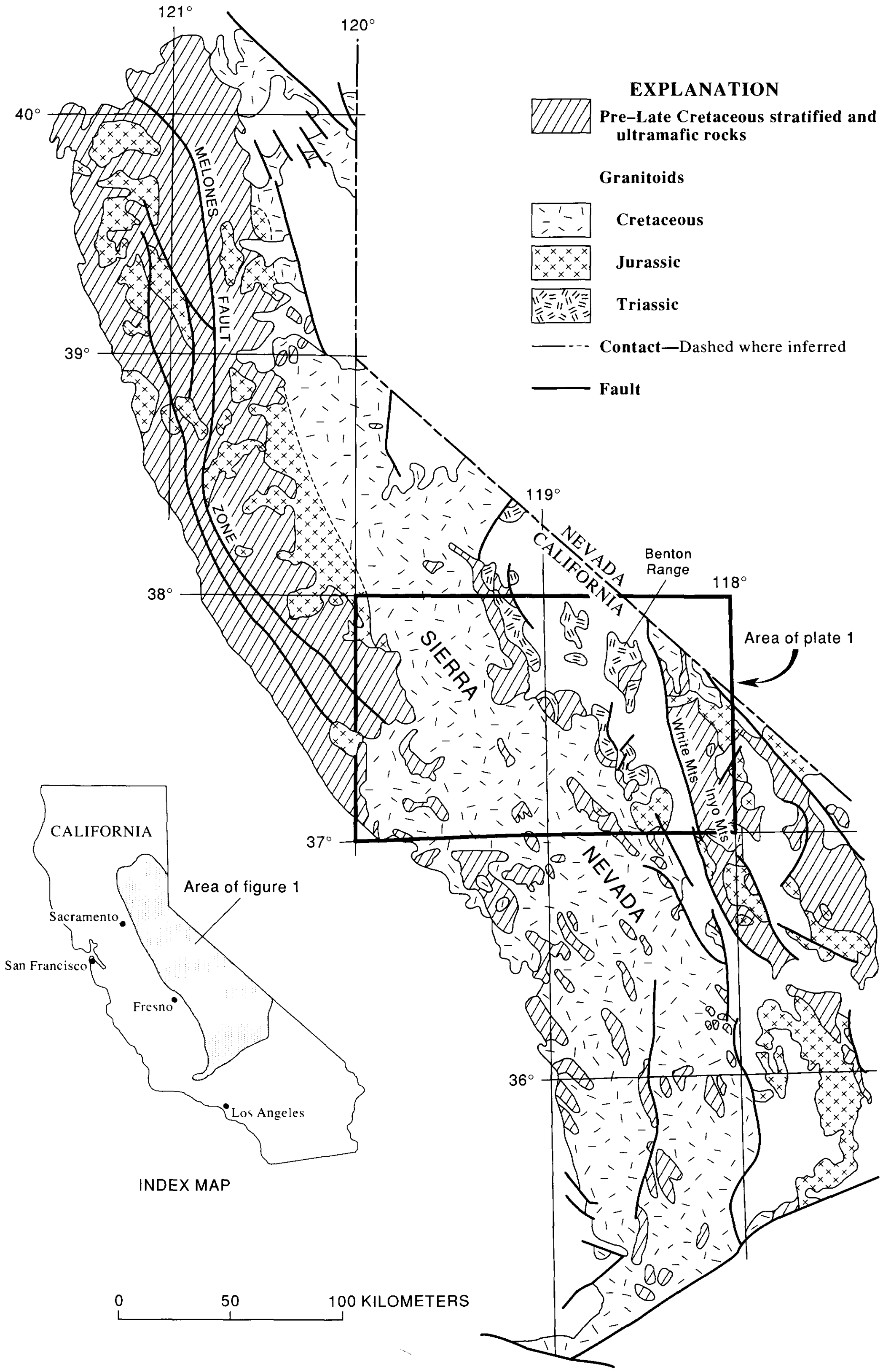


FIGURE 1.—Generalized geology of Sierra Nevada and adjacent areas in eastern California and location of Mariposa 1° by 2° quadrangle (area of pl. 1).

The mapping of the 15-minute quadrangles began in 1945, at the end of World War II, when the USGS and the California Division of Mines and Geology initiated a joint program to study selected mineralized areas of California. During World War II, USGS geologists examined most of the tungsten deposits in the Western United States, and the highly productive Bishop tungsten district, in the east-central Sierra Nevada, was a natural choice for further study under the cooperative program (Bateman, 1956, 1965). Another area that was selected for study was the western metamorphic belt (on the opposite side of the Sierra Nevada), which contains deposits of gold, copper, chrome, manganese, and limestone. Subsequently, other projects in the east-central Sierra Nevada were added to the program. When the cooperative program with the California Division of Mines and Geology terminated in the late 1950's, mapping of the central Sierra Nevada was well underway.

In 1958, I was assigned to do a comprehensive study of the belt across the batholith between lat 37° and 38° N. This assignment was designed to use the results of my own studies and those of other independent investigators. My first task was to join with others then carrying on studies in the central Sierra Nevada in preparing a summary of what then was known of the geology of the central part of the batholith; that summary was published as USGS Professional Paper 414–D (Bateman and others, 1963). In a sense, the present report is an updated and expanded version of the 1963 report. The 1963 report and a later report by Bateman and Wahrhaftig (1966) contain fairly complete summaries of geologic work that had been carried on through to the times of publication, so a summary of previous geologic work is not included in this report.

Because many geologists have contributed to the geologic mapping of the central Sierra Nevada, it is not practical to list them all. Many of them are either authors or coauthors of published maps or reports, and others are acknowledged in those reports. However, the following contributed heavily to the understanding of the batholith: D.M. Crowder, F.C.W. Dodge, J.P. Eaton, N.K. Huber, R.W. Kistler, K.B. Krauskopf, A.H. Lachenbruch, J.L. Lockwood, J.G. Moore, C.A. Nelson, H.W. Oliver, D.L. Peck, C.D. Rinehart, D.C. Ross, and D.R. Wones. Because I have been involved in this undertaking from its inception, I have had the opportunity to become familiar with the studies of all those involved. This report expresses my views, but most of them are also the consensus views of those who have participated; it also includes contrary views that nonetheless have substantial support.

Some lithodemic units and intrusive suites previously referred to only by informal names are given formal names in this report. Descriptions of these units, their type localities, and other pertinent information are given in the section "Descriptions of the Plutonic Rocks."

STRATIGRAPHIC AND STRUCTURAL SETTING

Before 1956, the Sierra Nevada batholith was generally regarded as an insuperable barrier to relating the stratified rocks in the western metamorphic belt to remnants within the batholith and to the strata east of the batholith in the Basin and Range province. However, the paradigm of plate tectonics has proved to be a great stimulus to studies of the country rocks because the Sierra Nevada is considered to lie within the zone affected by Mesozoic and Cenozoic convergence of the North American plate with plates of the Pacific Ocean. In recent years, many geologists have attempted to establish the stratigraphic and structural relations of the country rocks of the Sierra Nevada (see Schweickert and Cowan, 1975; Saleeby and Sharp, 1980; Nokleberg, 1983). In this endeavor they were severely hampered by a paucity of fossils and by the skeletal distribution and intense deformation of the country rocks. Consequently, many relations remain unresolved.

In the central Sierra Nevada, beds, cleavage, the axial surfaces of folds, and the long axes of country-rock remnants have an average strike of about N. 35° W., parallel to the long axes of the great Cretaceous intrusions. To the north, the general trend gradually becomes more northerly except at the extreme north end of the Sierra Nevada, where it bends to the northwest toward the Klamath Mountains. Although these trends have been called the Sierran trend, they are not restricted to the Sierra Nevada; major structural elements in the California Coast Ranges also parallel these trends. East of the Sierra Nevada, the Sierran trend terminates along the border between California and Nevada, in a large faulted oroflex (Albers, 1967; Stewart, 1967; Stewart and others, 1968).

The batholith lies along the edge of the Paleozoic North American craton, and its western margin crosses the foothill suture into Paleozoic and early Mesozoic oceanic crust. Proterozoic strata have been identified only in the White and northern Inyo

Mountains, but Paleozoic strata are present in roof pendants and septa in the eastern Sierra Nevada and also in the east half of the western metamorphic belt. Mesozoic strata predominate in the western metamorphic belt, compose roof pendants and septa in the High Sierra, and form the lower plate of a thrust fault in the White Mountains.

The country rocks of the Sierra Nevada are grouped into generally coherent sequences of strata, but some may be melanges. Some sequences are intruded on one or more sides by plutonic rocks, and others are bounded by faults. Precambrian and Paleozoic sedimentary rocks east of the Sierra Nevada in the White and Inyo Mountains, although complexly folded and faulted, apparently originated in the general area in which they occur, and correlative Paleozoic strata in roof remnants of the eastern Sierra Nevada can have been tectonically displaced only relatively short distances. On the west side of the batholith, in the western metamorphic belt, Mesozoic volcanic and associated sedimentary strata west of the Melones fault zone have been postulated to be allochthonous, some elements having originated in or adjacent to volcanic island arcs and some even in oceanic-rise systems (Hamilton, 1969; Dickinson, 1970; Schweickert and Cowan, 1975; Saleeby and others, 1978). Although the places of origin of these strata have not been established, it is probable that most of them originated nearby in marginal basins and island arcs.

METAMORPHISM

In the past, steeply dipping beds, folds, cleavages, greenschist- and lower amphibolite-facies metamorphism, and other characteristics of compressive deformation were commonly attributed to the Late Jurassic Nevadan orogeny, and the formation of hornfelses to thermal metamorphism caused by the later intrusion of granitoids. However, recognition that deformations have occurred repeatedly, some earlier and some later than the Nevadan orogeny (Tobisch and others, 1987), and that intrusions occurred episodically and overlapped the timespans of these deformations invalidates this simple picture. Furthermore, much of the regional metamorphism was caused by elevated thermal gradients associated with plutonism, thus blurring the distinction between thermal and regional metamorphism. For example, tonalite probably underlies greenschist-facies strata east of the Melones fault zone in the southern part of the western metamorphic belt, traditionally considered to have been regionally metamorphosed. The coloration of conodonts indicates that these strata have been heated to temperatures of 300 to 400 °C (Bateman and others, 1985).

Also, uncertainty as to the areal extent of the region affected by the Nevadan orogeny has been increasing. Both the Middle Jurassic tonalite of Granite Creek in the western Sierra Nevada and Early Cretaceous metavolcanic rocks of the Ritter Range roof pendant and Goddard septum have cleavages and lineations parallel to northwest trends considered to be Nevadan. The structures in the tonalite may have been produced during the Nevadan orogeny, but the metavolcanic rocks are much too young. Additionally, northwest-trending structures in late Paleozoic strata of the eastern Sierra Nevada were formed before these strata were intruded by Late Triassic granitoids.

Hornfelses adjacent to intrusions are clearly the products of thermal metamorphism. Thermal metamorphism was mainly in the hornblende hornfels facies, but the inner parts of some aureoles, especially adjacent to intrusions of quartz diorite and tonalite, belong to the next higher pyroxene hornfels facies, and strata distant from intrusive contacts, including the interiors of some large roof pendants, are in the lower albite-epidote hornfels facies.

In view of the sparsity of fossils to indicate the ages of many sequences, and because some adjacent, internally coherent sequences may have originated far apart, the country rocks are described from east to west—from roughly autochthonous to possibly allochthonous—rather than in the usual order of oldest to youngest.

LATE PROTEROZOIC AND PALEOZOIC STRATA OF THE WHITE AND NORTHERN INYO MOUNTAINS

The Late Proterozoic and Paleozoic strata of the White and northern Inyo Mountains are miogeoclinal or shelf deposits and are characterized by carbonate rocks (mostly dolomite in the Late Proterozoic and limestone in the Paleozoic), quartzite, and argillite (slate in places). According to Nelson (1981), these strata were deposited in tidal, subtidal, and reefal environments. The exposed Late Proterozoic and Lower Cambrian strata have an aggregate thickness of about 6,400 m and are one of the thickest accumulations of strata of this age range in North America (Nelson, 1962). These strata are overlain by another 2,500 m of Middle Cambrian to Permian strata.

Nelson (1962, 1966a) reported unconformities between the Late Proterozoic Wyman Formation and the Reed Dolomite and between the Sunday Canyon Formation, considered by Miller (1976) to be Devonian in age, and the Mississippian Perdido Formation in this thick section. The unconformity above the Wyman is shown only by lateral changes in the lithology of the upper part of the Wyman and is questionable, but the unconformity between the Sunday Canyon and Perdido Formations may reflect the widespread Devonian and Mississippian Antler(?) orogeny.

WYMAN FORMATION, REED DOLOMITE, AND DEEP SPRING FORMATION

The oldest exposed strata are assigned to the Wyman Formation. They consist of a thick succession of thin-bedded sandstone and siltstone and lenses of gray-blue, locally dolomitized oolitic limestone. The exposed thickness of this formation is about 2,700 m, but the bottom is concealed. An unconformity with the overlying Reed Dolomite is not apparent at individual exposures and is shown mainly by lateral changes in the lithologies of the upper beds of the Wyman. The Reed Dolomite consists chiefly of light-gray to buff dolomite but toward the east and southeast includes a northwest-tapering tongue of quartzite (Hines Tongue). The formation has a fairly consistent thickness of about 600 m. Conformably overlying the Reed is the Deep Spring Formation, consisting of about 500 m of limestone, dolomite, quartzite, and calcareous sandstone. The strata exposed along the range front east of Bishop are largely dolomite and sandy dolomite; quartzite is increasingly abundant eastward.

CAMPITO FORMATION

The Campito Formation comprises two members. The lower member (Andrews Mountain Member) consists of a 900-m-thick succession of dark-brown to gray quartzite, sandstone, and interbedded siltstone. The upper member (Montenegro Member) consists of about 150 m of gray-green shale in the Bishop area. The presence of olenellid fauna upward from the middle of the lower member and archaeocyathids in the thin, lenticular limestone beds of the upper member has been interpreted to indicate that the base of the Cambrian lies within the lower member (Nelson, 1962). However, correlations with Siberian platform successions suggest that the base of the Cambrian lies within the upper beds of the Reed Dolomite (Nelson, 1978).

POLETA, HARKLESS, SALINE VALLEY, AND MULE SPRING FORMATIONS

The Poleta, Harkless, Saline Valley, and Mule Spring Formations are all of Early Cambrian age. The Poleta Formation consists of a succession of archaeocyathid limestone, shale, and quartzite and varies in thickness from about 200 to 360 m, thinning toward the south. It has the most abundant trilobite fauna of any unit in the region. The formation is overlain by the Harkless Formation, which is composed of about 600 m of gray-green shale and thin-bedded siltstone and sandstone. Thin beds of pisolitic limestone and archaeocyathid limestone are present locally in the lower part of the Harkless. The overlying Saline Valley Formation is highly variable, in both lithology and thickness. It consists of medium- to coarse-grained quartzitic sandstone, blue-gray arenaceous limestone, and gray-green shale. It is highly lenticular and has a maximum thickness of 260 m. The next younger Mule Spring Formation (or Limestone) contains the stratigraphically highest olenellid fauna in the White and Inyo Mountains and therefore is the youngest Early Cambrian formation. It consists predominantly of blue-gray limestone, locally dolomitized, containing concretionary algal structures (*Girvanella*). Locally it contains a few thin interbeds of gray shale.

MONOLA FORMATION AND BONANZA KING AND TAMARACK CANYON DOLOMITES

The Monola Formation, Bonanza King Dolomite, and Tamarack Canyon Dolomite are considered to be of Middle Cambrian, Middle and Late Cambrian, and Late Cambrian age, respectively. The Monola Formation is about 365 m thick and consists of medium- to dark-gray shaly siltstone, buff limy siltstone, and fine-grained limestone. Ripple marks, crossbeds, and slump structures are locally abundant in the siltstone layers of the formation. The overlying Bonanza King Dolomite (or Formation) consists of laminated to thick-bedded, generally fine-grained, color-banded dolomite. Algal structures (*Girvanella*) and worm tracks are common. The thickness of the formation has not been established but is on the order of many hundreds of meters. The overlying Tamarack Canyon Dolomite consists of massive, light- to medium-gray, fine-grained dolomite, which locally contains black chert nodules. It is about 275 m thick.

EMIGRANT(?) FORMATION

Thrust slices of the Emigrant(?) Formation of Late Cambrian age rest locally on the Harkless Formation. They consist of thin-bedded gray limestone with beds and nodules of chert, platy gray shale, and thin-bedded silty to massive thick-bedded oolitic limestone.

CAMBRIAN STRATA, UNDIVIDED

Undivided Cambrian strata are present locally in the north end of the White Mountains and in the northern Inyo Mountains along the west side of the Cretaceous megacrystic granite of Papoose Flat. The exposures in the northern White Mountains are chiefly marble and phyllite, and their assignment to the Cambrian is uncertain. The undivided Cambrian strata peripheral to the Papoose Flat pluton include tectonically thinned Poleta, Harkless, Saline Valley, and Mule Spring Formations.

AL ROSE, BADGER FLAT, BARREL SPRING, JOHNSON SPRING, PALMETTO, AND SUNDAY CANYON FORMATIONS

The Al Rose, Badger Flat, Barrel Spring, and Johnson Spring Formations, all of Ordovician age, crop out in a small area along the south boundary of the map area (pl. 1). The Al Rose Formation consists of siltstone, shale, and mudstone; the Badger Flat of silty limestone, calcarenite, and buff limestone; the Barrel Spring of shale and siltstone underlain by gray-buff limestone and quartzite; and the Johnson Spring of vitreous quartzite interbedded with buff dolomite. The aggregate thickness of these formations does not exceed 500 m. The Palmetto Formation, also of Ordovician age, crops out in two small areas, one at the north end of the White Mountains and the other east of Dyer along the east edge of the map area. It consists of interbedded gray-to-black slate and cherty thin-bedded limestone. The Devonian Sunday Canyon Formation crops out along the west edge of the Inyo Mountains east of Tinemaha Reservoir. The lower part consists of platy gray-to-buff limy shale with a few limestone interbeds. Thin-bedded shaly limestone is progressively more abundant upward.

PERDIDO FORMATION, REST SPRING SHALE, AND KEELER CANYON FORMATION

The Perdido Formation and overlying Rest Spring Shale are Mississippian, and the Keeler Canyon Formation is Pennsylvanian and Permian. The Perdido Formation has two members. The lower member consists of dark-gray, fine-grained dolomite interbedded with layers of black chert and brown, fine- to medium-grained quartzite; conglomerate is present locally at the base. The upper member consists of thin-bedded, gray shale that contains interbeds of brown quartzite and conglomerate. The Rest Spring Shale consists of dark-gray siltstone, shale, and mudstone. The Keeler Canyon Formation is composed of thin-bedded, gray, locally cherty limestone interbedded with purple siltstone.

PALEOZOIC STRATA IN COUNTRY-ROCK REMNANTS OF THE EASTERN SIERRA NEVADA

Late Proterozoic strata have not been recognized in the roof pendants and septa of the eastern Sierra Nevada, but Paleozoic strata ranging in age from Early Cambrian to Pennsylvanian and Permian(?) have been identified. The locations of the roof pendants and septa referred to in this report are shown in figure 2. In general, the strata in these country-rock remnants contain less carbonate than approximately correlative strata in the White and Inyo Mountains. Nevertheless, Moore and Foster (1980) identified fossiliferous Lower Cambrian strata in the septum along Big Pine Creek and Ordovician strata in the Bishop Creek septum as belonging to several formations in the White and Inyo Mountains, indicating that a major structural break does not exist between the White and Inyo Mountains and the Sierra Nevada. Poorly preserved fossils indicate that the western parts of the sedimentary successions in the Ritter Range roof pendant and Saddlebag Lake septum are upper Paleozoic, possibly Mississippian(?) (Brook, 1979).

The Mount Morrison roof pendant is the largest in the eastern Sierra Nevada and is especially important because it has yielded fossils ranging in age from Early Ordovician to Pennsylvanian and Permian(?) and provides a stratigraphic succession with which the successions in other roof remnants can be compared (Rinehart and Ross, 1964). On the basis of comparison with the Mount Morrison roof pendant, the Pine Creek septum has been assigned to the Pennsylvanian and Permian(?) and the Log Cabin roof pendant west of Lee Vining to the early Paleozoic.

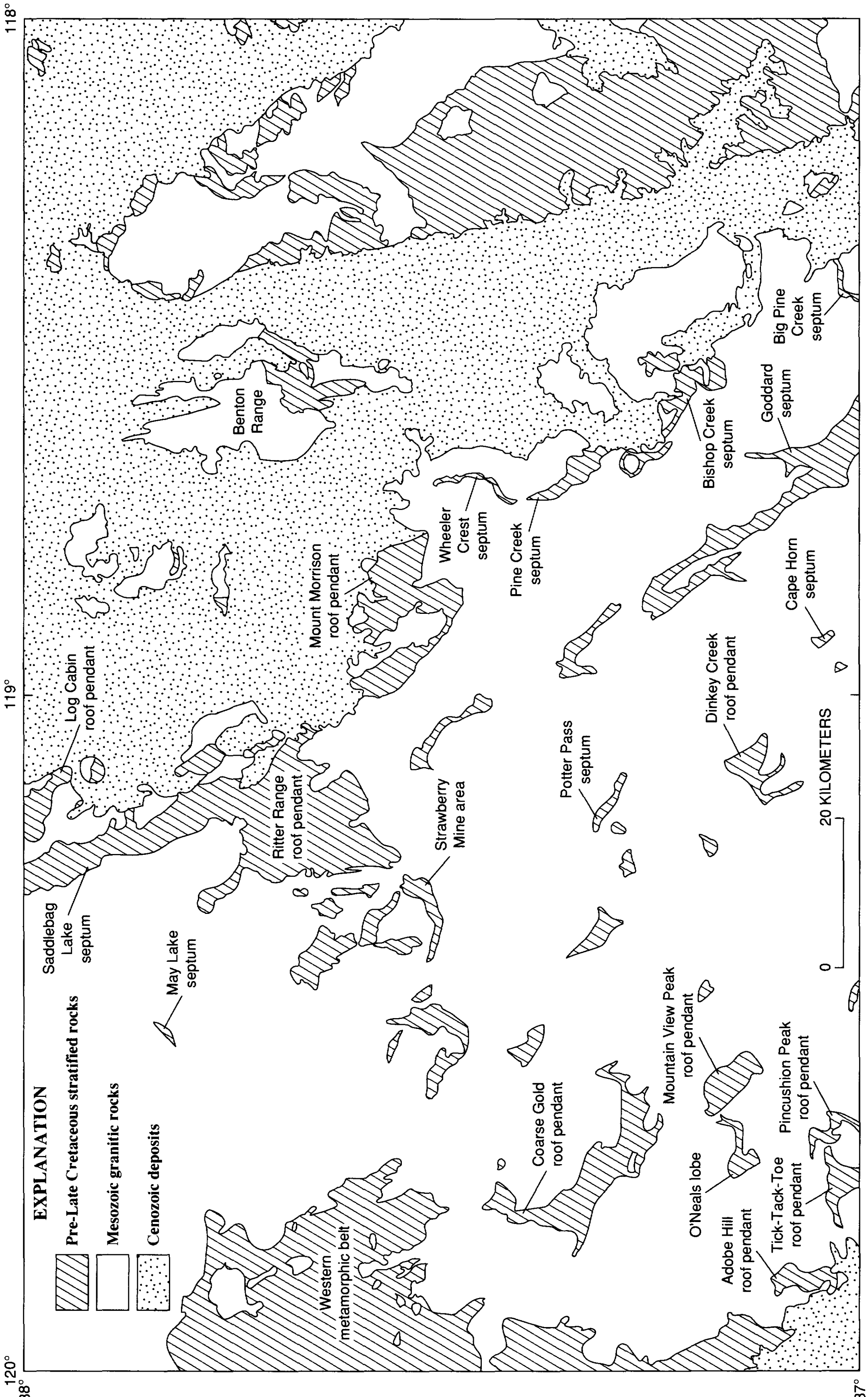


FIGURE 2.—Simplified geology and distribution of country rocks within Mariposa 1° by 2° quadrangle.

The Mount Morrison roof pendant consists of three structural blocks that are separated by two faults. Ordovician strata compose the eastern (McGee Mountain) block, Ordovician and Silurian(?) strata the middle (Convict Lake) block, and Pennsylvanian and Permian(?) strata the western (Bloody Mountain) block. The strata of the middle and western blocks dip steeply westward, in the direction of successively younger strata, and the strata of the eastern block dip steeply both east and west. Rinehart and Ross (1964) reported an aggregate stratigraphic thickness of about 10,000 m in the pendant, but the relation of this measured thickness to the original stratigraphic thickness has been questioned. Russell and Nokleberg (1977) expressed the opinion that the section originally was quite thin and has been thickened by internal folding during several episodes of deformation. On the other hand, Tobisch and others (1977), on the basis of strain studies, concluded that the contiguous, superjacent Koip sequence has been tectonically thinned about 50 percent. They suggested that other sections of stratified rocks in the Sierra Nevada have been similarly thinned. Both interpretations may be correct—the strata may have been thinned by attenuation but outcrop widths increased by internal folding.

Rinehart and Ross (1964) suggested that the Laurel-Convict fault, which separates the middle and western blocks, coincides with a significant unconformity. The presence of an unconformity is suggested by divergence of bedding and fold axes across the fault and by the fact that Ordovician and Silurian(?) strata in the western part of the middle block are juxtaposed against Pennsylvanian strata in the eastern part of the western block; Devonian and Mississippian strata appear to be missing. An unconformity in this position would probably correspond to unconformities in the Inyo Mountains between the Devonian Sunday Canyon Formation and the Mississippian Perdido Formation and, in the Candelaria Hills (about 100 km to the northeast), between Ordovician and Permian strata (Ferguson and Muller, 1949). Such a widespread unconformity is very likely an expression of the Late Devonian to Early Mississippian Antler orogeny of Nevada (Roberts, 1951).

MESOZOIC STRATA IN ROOF PENDANTS AND IN THE WHITE MOUNTAINS

The Mesozoic metavolcanic arc that extends along the crest of the Sierra Nevada is being studied by R.S. Fiske and O.T. Tobisch, and descriptions and interpretations of the metavolcanic strata given herein are mainly from their studies.

KOIP SEQUENCE AND OTHER TRIASSIC(?) AND JURASSIC METAVOLCANIC ROCKS

The Koip sequence as defined by Kistler (1966a, b) lies just west of Paleozoic strata in the Ritter Range roof pendant and its northern extension, the Saddlebag Lake septum. On-strike strata to the south in the western lobe of the Mount Morrison roof pendant and in the core of the Goddard septum have been included within this unit on plate 1 but may not be exact equivalents of the Koip sequence. The Koip sequence has been studied by Huber and Rinehart (1965), Tobisch and others (1977), Fiske and Tobisch (1978), Kistler and Swanson (1981), and Tobisch and others (1986). Huber and Rinehart (1965) designated rocks of this sequence "volcanic rocks of Shadow Creek and Mammoth Creek," and Fiske and Tobisch (1978) referred to these same rocks as "the lower section."

The sequence consists chiefly of dacitic to rhyolitic pyroclastic rocks; andesitic and basaltic flows, dikes, and sills; and more than 40 thin beds of limestone and calcareous tuff. The contact of the Koip sequence with Paleozoic strata on the east is sharp everywhere, and in most places the strata are structurally concordant. Nevertheless, folding appears to be more complex in the Paleozoic sedimentary succession than in the Koip sequence, and conglomerates that contain pebbles composed of the underlying sedimentary rocks occur in the base of the Koip sequence and within the Paleozoic sedimentary strata adjacent to the Koip sequence, especially in the Saddlebag Lake septum. These features indicate that the contact marks an erosional unconformity, except where it has been faulted as in the Pine Creek and southern part of the Mount Morrison roof pendants (Bateman, 1965b; Morgan and Rankin, 1972), and that it probably also marks a period of deformation. The overall impression is that sedimentary marine deposition during the late Paleozoic ended with uplift, erosion, and strong deformation—processes that were probably coincident with the Sonoman orogeny of central Nevada. The conglomerates within the sedimentary strata indicate uplift and erosion of an adjacent area, but their environment of deposition has not been determined. However, the basal conglomerates of the Koip sequence in the Saddlebag Lake area were deposited in stream channels (Brook, 1977) and indicate that the area was above sea level when they were deposited.

The strata of the Koip sequence form an essentially homoclinal section that is inclined 70°–80° to the southwest, except in the extreme western part, where strata west of a fault near Lake Ediza dip west much less steeply (25°–35°). The gross structure is locally complicated by soft-sediment slumps, faults, and folds, but these appear to be only minor disturbances in an orderly southwestward-dipping section (Fiske and Tobisch, 1978). Nevertheless, duplication along subtle bedding-plane faults is possible. Abundant crossbeds, small-scale channels, and less common graded beds indicate that the tops of beds face overwhelmingly to the southwest. The present thickness of the section, assuming no replication, is about 5 km, but strain studies (Tobisch and others, 1977; Fiske and Tobisch, 1978) indicate that the original thickness approached 11 km.

The presence of tabular or broadly lens-shaped depositional units having essentially planar tops and bottoms and the absence of unconformities or evidence of channeling between units indicate that the Koip sequence accumulated in an environment of low topographic relief. Thick, poorly sorted ash-flow tuffs indicate subaerial deposition, whereas soft-sediment slump structures and limestone and lime-cemented tuffs indicate subaqueous deposition. Pectinoid pelecypods at one locality indicate a shallow marine environment, but some subaqueous strata could have been deposited in fresh water. Fiske and Tobisch (1978) interpreted these relations to indicate that the rate of deposition almost equaled the rate of subsidence throughout the entire period of deposition. Shallow submarine conditions prevailed when subsidence outpaced deposition, and subaerial conditions prevailed when subsidence lagged behind deposition.

Samples collected from the eastern part of the unit, east of Shadow Lake, have yielded U-Pb ages of 214 to 187 Ma, and samples collected from the western part have yielded ages of 158 to 153 Ma, except for one sample (from the gently dipping strata in the extreme western part), which yielded an age of 177 Ma (R.S. Fiske and O.T. Tobisch, unpub. data, 1985). These ages suggest that the eastern part of the section is Late Triassic to Early Jurassic in age and that the western part is of Middle to Late Jurassic age. This simple picture is confused by the earlier discovery of a pectinoid pelecypod of the genus *Weyla*, known only from the Early Jurassic, in the eastern part of the zone that yielded Middle and Late Jurassic isotopic ages (Rinehart and others, 1959). If the U-Pb ages are correct, possibly the pectinoid was redeposited or misidentified.

Kistler and Swanson (1981) also divided the Koip sequence into younger and older successions separated by an unconformity, but their break would have occurred earlier than the hiatus defined by the two separate groups of U-Pb ages. They argued that the granite of Lee Vining Canyon, which has a Rb-Sr whole-rock age of 212±8 Ma (Kistler, 1966b) and a U-Pb zircon age of 210 Ma (Chen and Moore, 1982), intruded the older part of the section after it was deformed and before the more western metavolcanic rocks were erupted. A similar situation exists in the southern part of the Pine Creek septum, where the Tungsten Hills Granite (~200 Ma) intrudes metavolcanic rocks.

A volcanic conduit or source vent for the Koip sequence volcanic rocks has not been found within the Ritter Range roof pendant (Fiske and Tobisch, 1978), but several small bodies of andesite and monzonite intrude the Saddlebag Lake septum and adjacent Paleozoic metasedimentary strata. Thus far, the Koip sequence has not been shown to be the temporal equivalent of any of the plutonic rocks, but if the U-Pb ages for the older strata of the Koip sequence are correct, the older part of the Koip sequence may be cogenetic with the Scheelite Intrusive Suite, which has an isotopic age of about 210 Ma (Kistler, 1966b; Stern and others, 1981; Chen and Moore, 1982). Isotopic dating has not revealed any granitoids that were emplaced during the extended timespan when the younger parts of the Koip sequence apparently were erupted.

Jurassic strata in the core of the Goddard septum consist largely of medium- to thick-bedded tuff and lapilli tuff (fig. 3), thin zones of calcareous tuff, limestone, mafic lava flows, and a few felsic lava flows (Tobisch and others, 1986). Beds dip steeply west,

FIGURE 3.—Tuffaceous metavolcanic rocks in Goddard septum showing bedding, crosscutting cleavage, and downdip lineation.

and abundant sedimentary structures indicate that bedding tops face consistently to the west. The strata have yielded a U-Pb age of about 160 Ma, and a quartz syenite sill that probably is cogenetic with the volcanic rocks has an age of 157 Ma. These ages are in the same range as the younger U-Pb ages for rocks from the Koip sequence. Early Cretaceous metavolcanic strata bound these strata on both sides, probably stratigraphically on the west and in fault contact on the east (Tobisch and others, 1986).

DOMINANTLY PYROCLASTIC STRATA OF EARLY CRETACEOUS AGE

This unit comprises a succession of dominantly pyroclastic rocks that form the western part of the Ritter Range roof pendant and the margins of the Goddard septum; it probably also forms the masses between the pendant and the septum. The succession in the Ritter Range roof pendant includes pyroclastic deposits, flows, bedded tuff, and hypabyssal intrusions ranging in composition from rhyolite to basalt. U-Pb ages of 144 and 132 Ma indicate an Early Cretaceous age. These strata are in fault contact with the younger Minarets sequence on the east and are intruded by granitic rocks on the west; they are not in contact with the Koip sequence.

The Early Cretaceous strata in the Goddard septum include felsic ash-flow tuffs, interbedded lapilli tuff, medium- to thin-bedded tuff that contains lapilli and breccia fragments, strongly deformed phyllitic schist, lapilli tuff and tuff breccia, sills, dikes, and lava flows. The strata within the western part of the Goddard septum are mostly waterlain, whereas those within the eastern part are mostly subaerial deposits (Tobisch and others, 1986). All the strata dip steeply west and face west, and they are structurally conformable with the Jurassic strata that form the core of the septum. The strata within the western part of the septum may stratigraphically overly the Jurassic strata in the core of the septum, but the strata within the eastern part are in fault contact with the Jurassic strata. Three U-Pb ages for the strata of the east side of the septum are 144, 141, and 140 Ma, and two ages for strata of the west side are 137 and 136 Ma (Tobisch and others, 1986).

DANA SEQUENCE

The Dana sequence composes most of a down-dropped block that extends from Mount Dana northwest across Tioga Pass to Gaylor Peak and, according to Kistler (1966b), rests unconformably on the Koip sequence. It consists of metavolcanic tuffs, lapilli tuff, shale, calc-silicate hornfels, and marble. Dips are mostly gentle, and the strata are folded into an open northeast-trending anticline and companion syncline. Kistler and Swanson (1981) reported a Rb-Sr whole-rock age of 118±11 Ma based on an isochron drawn using only three data points representing samples collected from a unit in the upper part of the sequence. They state that although this age is not reliable, it does indicate that the succession is Cretaceous. The gentle dips and the unconformable relations with the Koip sequence suggest that this sequence is correlative with the mid-Cretaceous Minarets sequence.

MINARETS SEQUENCE

The Minarets sequence crops out continuously in the area of the Minarets, between the Koip sequence and the dominantly pyroclastic strata of Early Cretaceous age, and as isolated blocks west of the Cretaceous sequence, within the outcrop area of the granodiorite of Jackass Lakes. The strata in the eastern area rest unconformably on the Koip sequence and dip gently east, at angles that rarely exceed 35°, to a fault contact with the Early Cretaceous metavolcanic sequence. A mid-Cretaceous age is indicated by several U-Pb ages that are mostly in the range of 102 to 99 Ma (Fiske and Tobisch, 1978; Stern and others, 1981) and by a Rb-Sr whole-rock age of 99±2 Ma (Kistler and Swanson, 1981).

Fiske and Tobisch (1978) interpreted most of the rocks of the Minarets sequence to be material that filled a volcanic collapse structure which they called the Minarets caldera. The caldera fill consists of a single thick layer of ash-flow tuff, which contains a deposit of caldera-collapse breccia and is overlain by thinly laminated tuff that may have been deposited in a caldera lake. The breccia consists of material ranging from fine ash to huge blocks of rhyolite nearly 2 km across; it probably was formed by collapse of the caldera walls (Fiske and Tobisch, 1978).

The angular unconformity at the base of the Minarets sequence is a notably uneven surface that is characterized by steep-walled gullies and small canyons, which are partly filled with conglomerate that consists chiefly of volcanic clasts but that includes a few granitoid clasts. The topographic irregularities indicate that the local relief was rugged in the mid-Cretaceous, in contrast to the subdued topography inferred during deposition of the Jurassic and Early Cretaceous strata (Fiske and Tobisch, 1978). Granitoid clasts in the conglomerate at the base of the caldera fill show that some granitoids already had been unroofed in mid-Cretaceous time.

Several porphyries and other hypabyssal intrusions have the same isotopic ages as the metavolcanic rocks and appear to represent late surges of cogenetic magma. Similar isotopic ages and close spatial association with the three small intrusive suites of Merced Peak, Buena Vista Crest, and Washburn Lake and with the unassigned leucogranite of Graveyard Peak also suggest a cogenetic relation with the Minarets sequence.

METAVOLCANIC ROCKS, UNDIVIDED

The U-shaped mass of metavolcanic rock in the northern White Mountains, two small masses of metavolcanic rock in the Sierra Nevada (one west of the Minarets sequence and the other west and south of the granite porphyry of Star Lakes), and a larger mass farther southeast along the southwest side of the bulbous head of the Mount Givens Granodiorite have not been studied in sufficient detail to assign them to sequences. Probably they are correlative with the Koip sequence and (or) the dominantly pyroclastic strata of Early Cretaceous age.

In mapping the White Mountains, Krauskopf (1971) and Crowder and Sheridan (1972) distinguished upper and lower dominantly metasedimentary units from chiefly metavolcanic units. They further subdivided the metavolcanic units into felsic and mafic facies. More recently, Hanson and others (1987) restudied the area and concluded that the dominantly sedimentary strata are faulted parts of the same succession and unconformably or disconformably overlie the dominantly metavolcanic strata. All the strata dip generally westward and are younger westward.

The dominantly metavolcanic succession consists of both lava flows and tuffs. Hornblende-bearing meta-andesite is common among the mafic rocks, and metarhyolite is the most common felsic rock type. Relict pyroclastic textures and flattened pumice fragments are common among the felsic rocks. The dominantly metasedimentary succession contains rocks with lithologies as diverse as schist and phyllite, metalimestone (as disrupted lenses), metavolcanic siltstone and sandstone (in part calcareous), and coarse metaconglomerate. Metaconglomerate at the base of the metasedimentary succession contains pebbles and cobbles of metavolcanic rocks, quartzite, and dense undetermined rocks as large as 15 cm across. Locally, clasts are strongly flattened and stretched.

Zircon from a sample of metarhyolite tuff collected along the lower part of Milner Canyon in the White Mountains yielded a concordant U-Pb age of 140 Ma (table 1), and Hanson and others (1987) reported U-Pb ages of 154 Ma on zircon from an ash-flow tuff just north of White Mountain Peak and 137 Ma for a nearby hypabyssal rock. Hanson and others (1987) also reported a U-Pb age of 100 Ma for the granite of McAfee Creek, which intrudes the eastern part of the metavolcanic succession. All these ages are younger than published U-Pb ages of 165 and 161 Ma for the quartz monzonite of Mount Barcroft (Gillespie, 1979; Stern and others, 1981), which intrudes the metavolcanic succession but which, according to Hanson and others (1987), is in fault contact with the metasedimentary succession. The obvious inferences are that (1) the metavolcanic succession ranges in age from at least as old as Middle Jurassic to Early Cretaceous, (2) the metasedimentary succession is at least as young as Early Cretaceous, and (3) the quartz monzonite of Mount Barcroft intruded the dominantly volcanic succession as it was erupting and is cogenetic with the succession.

KINGS SEQUENCE

The Kings sequence (Bateman and Clark, 1974) is exposed discontinuously in a series of country-rock remnants that lie west of the Koip sequence and Early Cretaceous metavolcanic sequences and cross the west side of the Minarets sequence. Within the map area it extends from the Dinkey Creek roof pendant and small masses of quartzite farther south, close to the 37th parallel, northwest to the May Lake and other septa that lie along the west contact of the Tuolumne Intrusive Suite (pl. 1). However, the sequence has been traced both to the north and to the south beyond the limits of the map area (Saleeby and others, 1978; Clyde Wahrhaftig, oral commun., 1985), and the sequence takes its name from the fossiliferous Boyden Cave roof pendant along the Kings River, just south of the map area. Within the map area, the sequence consists of metasedimentary strata—chiefly quartzite, marble, andalusite hornfels, and schist; farther south, volcanic and volcanogenic strata are present (Knopf and Thelen, 1905; Christensen, 1963; Saleeby and others, 1978).

The Late Triassic and (or) Early Jurassic age designation for the Kings sequence is based on fossils found south of the map area. Fossils of Early Jurassic and Late Triassic ages were found in the Boyden Cave roof pendant (Moore and Dodge, 1962; Jones and Moore, 1973), the Mineral King roof pendant (Knopf and Thelen, 1905; Durrell, 1940; Christensen,

TABLE 1.—*U-Pb data for metarhyolite (unit* Mzv, *pl. 1) from the White Mountains and for the tonalite of Millerton Lake (unit Kml, pl. 1)*

[Analyses by T.W. Stern. n.d., no data]

Sample ----------	W–2	MLc–170
Unit --------------	Metarhyolite from White Mountains	Tonalite of Millerton Lake
	Ages (Ma)	
$^{206}Pb/^{238}U$ ----	147.9	133.9
$^{207}Pb/^{235}U$ ----	146.3	132.8
$^{208}Pb/^{232}U$ ----	147.6	n.d.
	Trace elements, in parts per million	
Pb -------------	14.6800	4.0653
U --------------	444.6900	189.4189
Th -------------	425.4700	.0000
	Isotopic ratios	
$^{208}Pb/^{208}Pb$ ---	0.432620	0.110470
$^{207}Pb/^{206}Pb$ ---	.102450	.060360
$^{204}Pb/^{208}Pb$ ---	.008670	.000820

1963), the Yokohl Valley roof pendant (Saleeby and others, 1978), and roof remnants near Isabella (Saleeby and others, 1978). Saleeby and others (1978) also reported Late Permian fossils from the Yokohl Valley roof pendant in a twice-reworked olistolith, but the rock types in the Yokohl Valley roof pendant, as they described them, are unlike those elsewhere in the Kings sequence and may be incorrectly assigned to the Kings sequence.

Fossils have been collected within the map area from sedimentary strata in the vicinity of the Strawberry Mine and from the Potter Pass septum (fig. 2). The first fossil found near the Strawberry Mine was a bivalve, possibly *Inoceramus pseudomytiloides* of Early Jurassic age (Nokleberg, 1981b). However, in 1982, R.S. Fiske and O.T. Tobisch collected an extensive fauna from the same general area; this fauna has been only partially studied. Preliminary determinations for about 25 percent of the collection suggest a Late Jurassic or Early Cretaceous age (R.W. Imlay, written commun., 1983). If a Late Jurassic or Early Cretaceous age for this collection is confirmed, either the sedimentary strata in the Strawberry Mine area are incorrectly assigned to the Kings sequence or the Kings sequence includes strata as young as Late Jurassic or Early Cretaceous. The fossil collected from the Potter Pass septum was identified by George Stanley (written commun., 1986) as "a scleractinian coral probably assignable to either *Koilocoenia* (Up. Tr.-Mid. Jur.) or *Actinastraea* (Tr.-Cret.)."

Within the map area, the Kings sequence has been most carefully studied in the Dinkey Creek roof pendant (Kistler and Bateman, 1966), the Strawberry Mine area (Nokleberg, 1981b), and the May Lake septum (Bateman, Kistler, and others, 1983). The strata in the Dinkey Creek roof pendant are folded and thrust faulted; nevertheless, Kistler and Bateman (1966) determined the stratigraphic succession from oldest to youngest to be (1) gray marble, (2) calc-silicate rock, (3) biotite-andalusite hornfels, (4) light-gray, conspicuously crossbedded quartzite, and (5) thinly interbedded schist and quartzite. In the Strawberry Mine area, the Kings sequence appears to form a window through the Cretaceous Minarets sequence and is intruded by hypabyssal dikes and sills. Clearly it occupied its present position when the Minarets caldera erupted. The complexly folded strata consist of calc-silicate hornfels, marble, and biotite hornfels derived from marble, limestone, and calcareous shale, respectively (Nokleberg, 1981b). The May Lake septum (fig. 2) consists of light-gray quartzite, diverse calc-silicate rocks, and marble (Bateman, Kistler, and others, 1983).

The Kings sequence is separated by granitoids from the Koip and Dana sequences and related Jurassic and Early Cretaceous metavolcanic strata to the east and from the Paleozoic(?) quartzite of Pilot Ridge to the west. The inferred ages of the Kings and Koip sequences indicate that the Kings sequence is probably older than all but the lowest beds of the Koip sequence. Although no contacts between these sequences have been observed within the map area, Saleeby and others (1978) reported that south of the map area the Kings sequence interfingers eastward with intermediate to silicic tuffaceous rocks (including dacite breccia and tuff breccia, and rhyodacite crystal tuff to lithic tuff) and westward with mafic to intermediate volcanic rocks (including basaltic andesite pillow lava, pillow breccia, tuff breccia, and crystal-lithic tuff). Thus, the lowest beds of the Koip sequence may be interlayered with the Kings sequence where both sequences are present in the same area. Saleeby and others (1978) pictured the Kings sequence as a fan complex that was shed off the continental shelf and dispersed across earlier accreted allochthonous oceanic rocks.

The west boundary of the Kings sequence within the map area is uncertain. Quartzite in the next unit to the west, the Paleozoic(?) quartzite of Pilot Ridge, resembles quartzite in the Kings sequence except that crossbedding is scarce. The Kings sequence could have been in fault contact with the quartzite of Pilot Ridge before intrusion of the granitoids, or it could have rested unconformably on the quartzite of Pilot Ridge.

STRATA EAST OF THE MELONES FAULT ZONE WITHIN AND ADJACENT TO THE WESTERN METAMORPHIC BELT

Because of poor exposures, incompetent beds, and obscure structures and because of a paucity of fossils, marker beds, and criteria for determining bedding facings, the strata east of the Melones fault zone in the western metamorphic belt, which have traditionally been assigned to the Calaveras Formation, are the most enigmatic within the map area. The diverse rock types include metavolcanic rocks of dominantly basaltic andesite composition adjacent to the Melones fault zone and chiefly metasedimentary rocks farther east. The sedimentary strata include argillite, phyllite, chert, impure quartzite, sparse limestone lenses, and abundant diamictite. In most places the strata are strongly and penetratively cleaved. The most conspicuous cleavage parallels bedding in most places, and both strike northwest and are vertical or dip steeply northeast. Nevertheless, locally abundant minor folds and divergences of cleavage and bedding suggest that major folds are present, even though none have been identified. Unrecognized faults may also be present.

The name Calaveras Formation was introduced by Turner (1893a, b) to include all the metasedimentary rocks of Paleozoic age in the western Sierra Nevada except Silurian and upper Carboniferous rocks in the Taylorsville area of the northern Sierra Nevada. The name came from fossiliferous beds in Calaveras County; these beds are part of a melange (Duffield and Sharp, 1975) and lie west of the Melones fault zone.

Schweickert and others (1977) proposed that the name Calaveras Complex be substituted for Calaveras Formation between lat 37°30' and 38°45' N. and for isolated outcrops as far south as 36°00' N. They excluded rocks west of the Melones fault zone, which had previously been included in the Calaveras Formation, and did not discuss rocks currently assigned to the Calaveras between 39°30' and 40°00' N. They identified four units from west to east—a volcanic unit, argillite, chert, and quartzite, which they assumed, following Clark (1964), ranged in age from Carboniferous or Permian in the west to Triassic and Jurassic in the east. However, the term Calaveras Complex of Schweickert and others (1977) had a short life. Soon after publication of the report by Schweickert and others (1977), Schweickert (1977) traced the Calaveras-Shoo Fly thrust fault southward into the area shown on plate 1 and identified the easternmost unit as an extension of his early Paleozoic Shoo Fly Complex (or Shoo Fly Formation). Later, Schweickert and others (1984) excluded the volcanic and argillite units, which formed the western part of their Calaveras Complex, and assigned them to the Jurassic, leaving only one of the original four units in their Calaveras Complex. Because of the uncertain status of the Calaveras Formation and because the Calaveras Complex of Schweickert and others (1977) had been reduced to a small part of its original extent, Bateman and others (1985) did not use the name Calaveras; instead they used informal names for the four units that approximately correspond to those that Schweickert and others (1977) used. These informal names are used for the units shown on plate 1. From east to west they are (1) quartzite of Pilot Ridge, of Paleozoic(?) age, (2) phyllite and chert of Hite Cove, of Triassic age, (3) phyllite of Briceburg, of Triassic age, and (4) greenstone of Bullion Mountain, of Jurassic age. Rather than being progressively younger eastward, as postulated by Clark (1964) on the basis of sparse observations of bedding facings and as assumed by Schweickert and others (1977), these units are progressively younger westward (Bateman and others, 1985). However, the westward decrease in age of units may be the result of tectonism rather than stratigraphic succession.

DePaolo (1981) interpreted strontium and neodymium isotopic data to indicate that the metavolcanic strata are continental-margin arc-type volcanic rocks and that the sedimentary rocks are largely derived from Precambrian continental crust with less than 20 percent derived from contemporaneous(?) volcanic-arc rocks. He suggested that the strata within the southern part of the western metamorphic belt east of the Melones fault zone were deposited at the distal end of a westward-flowing drainage system. A basin behind a volcanic arc seems a likely site for deposition of these strata. The Triassic and (or) Jurassic Kings sequence could also have been part of the system.

QUARTZITE OF PILOT RIDGE

The quartzite of Pilot Ridge, of Paleozoic(?) age, is characterized by massive beds of gray quartzite interstratified with argillite and phyllite. A few lenses of silicated marble are present locally. Isotopic ages of 440±15, 330±10, and 275±10 Ma for orthogneisses that intrude strata northwest of the map area, more or less on strike with the quartzite of Pilot Ridge, provide the strongest evidence that the quartzite of Pilot Ridge is Paleozoic in age (Sharp and others, 1982). The contact zone between the quartzite of Pilot Ridge and the Triassic phyllite and chert of Hite Cove is composed of thin-bedded gray quartzite, diamictite, and small masses of limestone and mafic volcanic and hypabyssal rock in a matrix of dark-gray siliceous carbonaceous metapelite. Although the contact zone is disturbed and is shown as a probable fault on plate 1, it may not mark a major fault. If the Calaveras-Shoo Fly thrust does continue southward into this area, it may lie farther east.

PHYLLITE AND CHERT OF HITE COVE

The phyllite and chert of Hite Cove unit consists of rhythmically bedded chert sequences as much as 100 m thick interbedded with argillite, phyllite, and thin beds and lenses of marble (fig. 4). The criteria that identify the chert are extremely fine grain size, granoblastic texture, restriction of impurities to iron oxides and sericite, sparse microscopic spheroids that may be relict radiolarians, and $\delta^{18}O$ values that range from 22.2 to 24.1 per mil in five samples (Bateman and others, 1985).

FIGURE 4.—Folded, rhythmically layered Triassic chert along Merced River.

Fossils were collected from two limestone beds within this unit. One extensive bed crosses the Merced River in the central part of the unit and is the site of an extensive but now inoperative limestone quarry. This bed was found to contain conodonts of Early Triassic (Griesbachian and Smithian) age (Bateman and others, 1985). The bed is interstratified with rhythmically bedded chert and phyllite and is not an olistolith. The other reported fossils, collected much earlier by Turner (1893b) along the South Fork of the Merced River near Hite Cove, are foraminifers that were identified as *Fusulina cylindrica*. According to Clark (1964), they indicate a Pennsylvanian or Permian age. This fossil collection apparently has been lost, and the exact locality where it was collected is not known; probably it was at Marble Point, where a conspicuous limestone lens crops out. Efforts to re-collect identifiable fossils from this lens and also from a lens at the crest of the ridge north of Hite Cove have been unsuccessful.

Schweickert and others (1984) showed that the phyllite of Briceburg is separated from the phyllite and chert of Hite Cove by a fault that they called the Sonora fault. However, Bateman and others (1985), unable to verify the presence of a fault within the map area in the position shown by Schweickert and others, concluded that the contact is gradational and placed it along a line that separates sparse chert to the west from abundant chert to the east.

PHYLLITE OF BRICEBURG

The phyllite of Briceburg unit includes phyllite, argillite, metagraywacke, sparse chert, and blocks of limestone that appear to be olistoliths. A sample collected from the largest of more than a dozen limestone blocks in the central part of this unit east of Briceburg yielded Early Triassic (Smithian) conodonts similar to those in the phyllite and chert of Hite Cove unit. All these blocks are probably olistoliths whose source may have been a limestone bed in the phyllite and chert of Hite Cove, probably the limestone bed that yielded the conodonts. If so, the Early Triassic age of the conodonts is merely a maximum age for the phyllite of Briceburg, and it may be younger than Early Triassic.

The phyllite of Briceburg also contains several small masses of metagabbro, which intrude and have been deformed with the phyllite. The metagabbro probably was comagmatic with the greenstone of Bullion Mountain; if so, the phyllite is older than the greenstone.

The contact between the phyllite of Briceburg and the greenstone of Bullion Mountain to the west is characterized by local interfingering of phyllite and greenstone and ductile deformation across a zone as much as a few hundred meters thick. This contact does not appear to be the locus of major faulting, but major faulting cannot be ruled out.

GREENSTONE OF BULLION MOUNTAIN

The greenstone of Bullion Mountain has the average composition of basaltic andesite (table 2) and includes metamorphosed tuff, breccia, and locally pillow lava. Small masses of medium- to fine-grained metagabbro that intrude the adjacent phyllite of Briceburg and have undergone at least one episode of regional deformation probably are hypabyssal equivalents of the greenstone. The unit is considered to be of Late Jurassic (late Oxfordian) age because an ammonite of that age, identified by Imlay as *Perisphinctes* (*Dichotomosphinctes*) cf. *P.* (*D.*) *muhlbachi* Hyatt (Clark, 1964), was collected from the western part, adjacent to the Melones fault zone. However, samples of compositionally diverse rocks (chiefly metamorphosed tuff, breccia, and possibly lava flows) collected from the northeast margin along the Merced River yielded a whole-rock Rb-Sr isochron that indicates an Early Jurassic age of 187±10 Ma (Bateman and others, 1985). Thus it is possible that the whole eastern part of the unit is Early Jurassic in age; this age is comparable to the Triassic or Early Jurassic age inferred by Morgan and Stern (1977) for the Peñon Blanco Volcanics west of the Melones fault zone.

STRATA IN ROOF PENDANTS SOUTH OF THE WESTERN METAMORPHIC BELT

Farther south, on strike with the western metamorphic belt, in the Coarsegold, Mountain View Peak, and Tick-Tack-Toe roof pendants (fig. 2), the phyllite of Briceburg and the phyllite and chert of Hite Cove are combined to form a single unit, but the greenstone of Bullion Mountain retains its identity (Bateman and Busacca, 1982; Bateman and others, 1982). Quartzite like that of Pilot Ridge is present in Goat Mountain, south of Bass Lake, and in Black Mountain along the south edge of the map area. Several remnants that lie slightly farther east and extend southeast from the remnant south of Raymond Mountain to Castle Peak are composed of thinly interbedded phyllite (or schist) and quartzite. These rocks are tentatively included in the quartzite of Pilot Ridge.

MARIPOSA FORMATION AND ADJACENT STRATA WEST OF THE MELONES FAULT ZONE

The strata west of the Melones fault zone consist of the Mariposa Formation immediately adjacent to the fault zone and a series of metasedimentary and metavolcanic units farther west and also south in the southern prong of the western metamorphic belt and in the Adobe Hill roof pendant (fig. 2). These units are not shown separately on plate 1. In all these units, steep, east-dipping cleavage is far more conspicuous than bedding. Although cleavage locally cuts across bedding, it is parallel in most places with both bedding and contacts between units. Moderately abundant graded beds in the Mariposa show that the tops of beds generally face east and suggest that the units are successively younger toward the northeast, the Mariposa being the youngest unit.

According to this interpretation, the oldest unit in this sequence is chlorite schist, which is found in the west side of the southern prong of the western metamorphic belt. This schist is stratigraphically overlain eastward by units of andalusite schist, quartz-biotite schist, metavolcanic and associated hypabyssal rocks, and biotite-quartz schist. The protoliths of these units were probably mafic tuffaceous-volcanic rocks, argillaceous sediment and silt, intermediate to silicic volcanic rocks and hypabyssal intrusions, and siltstone or fine-grained sandstone, respectively. Crystals of chiastolite, now largely altered to muscovite, characterize the andalusite schist. With the disappearance of chiastolite, the rock grades into the quartz-biotite schist unit. Northward, the Mariposa appears to overlie the biotite-quartz schist unit directly, but in the Adobe Hill roof pendant, which is separated from the southern prong of the western metamorphic belt by granitoids, an amphibolite unit lies between units considered equivalent to the biotite-quartz schist unit and the Mariposa Formation.

The Mariposa Formation consists chiefly of black carbonaceous and silty slate, but it contains graywacke, tuff, and local conglomerate. It has yielded abundant fossils of late Oxfordian and early Kimmeridgian (Late Jurassic) age (Clark, 1964). None of the thin underlying strata have yielded fossils, but I assume them to be Late Jurassic because they appear to conformably underlie the Mariposa Formation.

STRUCTURES IN THE COUNTRY-ROCK SEQUENCES

The gross structure of the country rocks into which the batholith was emplaced is still not understood. An early interpretation, that the country rocks constitute a complex faulted synclinorium (Bateman and others, 1963; Bateman and Wahrhaftig, 1966; Bateman and Eaton, 1967), has been abandoned, but a wholly satisfactory alternative interpretation of their stratigraphic and structural relations has not yet been developed.

All the rocks are strongly deformed, and many have been deformed more than once. Westward across the Sierra Nevada the structures are increasingly obscure and may be more complex. However, the strata in the western metamorphic belt are less competent than the strata in the eastern Sierra Nevada and White and Inyo Mountains, and obscurity may not equate with complexity. The strata in the White and Inyo Mountains are complexly folded and faulted, but fossils and distinctive stratigraphic units make it possible to identify and map the major structures. Generally, structures can also be determined in the roof remnants of the eastern Sierra Nevada even though fossils are scarce. However, farther west, in remnants of metamorphic rocks within the western part of the batholith and in the western metamorphic belt, structures are extremely difficult to identify because of the poor exposures, absence of distinctive stratigraphic units, paucity of fossils, and incompetence of the rocks. In the western metamorphic belt, both bedding and the most prominent cleavage characteristically are vertical or dip steeply east. Small-scale folds, many of them shear bounded, are common.

Traditionally, all the structures have been explained in terms of regional compression and crustal shortening during the Nevadan orogeny, but regional extension and batholith emplacement probably were more important. Episodic emplacement of the plutonic magmas, which expanded as they were emplaced, undoubtedly accounts for much of the deformation of the wallrocks of the batholith and of the country-rock remnants within the batholith.

STRUCTURES IN THE WHITE AND NORTHERN INYO MOUNTAINS

The strata in the White and northern Inyo Mountains are complexly folded and faulted, but competent and distinctive marker beds generally make it possible to decipher the structure. Along strike, the dips of axial surfaces commonly change from east to west and the plunges of fold axes change from horizontal to vertical. The thicknesses of units, especially in the incompetent strata that stratigraphically overlie the Late Proterozoic and Cambrian Campito Formation, increase and decrease irregularly. Thicknesses in the cores of folds may be several times those in limbs.

The largest fold structures are two complexly faulted anticlinoria and an intervening synclinorium. The anticlinoria are shown by outcrops of the Late Proterozoic Wyman Formation, Reed Dolomite, and Deep Spring Formation, and the synclinorium is shown by outcrops of the Early Cambrian Poleta and Harkless Formations (pl. 1). The northern anticlinorium lies north of Deep Spring Valley in the southern White Mountains. It is intruded by two small plutons, the granites of Sage Hen Flat and of Birch Creek, and is truncated on the east and northeast by the Cottonwood Granite (see section "Cottonwood Granite (Jc)") and the granodiorite of Beer Creek. The southern anticlinorium is in the northern part of the Inyo Mountains just south of Deep Spring Valley and is intruded by the megacrystic granite of Papoose Flat. The axis of the northern anticlinorium bears north, whereas the axis of the southern anticlinorium bears northwest (pl. 1). At its north end, the synclinorium lies west of and parallels the northern anticlinorium, but southward it bends toward the east between the two anticlinoria, crosses Deep Spring Valley, then bends southeast along the northeast side of the southern anticlinorium.

Extending west from the northern anticlinorium and the northern part of the synclinorium to the range front are a series of tight folds in which the thick and competent Andrews Mountain member of the Campito Formation has squeezed the incompetent Montenegro shale member in the top of the Campito and the equally incompetent overlying Poleta and Harkless Formations into a series of discontinuous masses in which the strata have been markedly thickened or thinned. Exposures of the Wyman, Reed, and Deep Spring along the base of the range south of Poleta Canyon suggest that the southern anticlinorium extends northward just west of the range front. Westward from the core of the southern anticlinorium, the strata dip generally toward the southwest at a slightly greater angle than the slope of the topography. Thus, progressively younger strata are exposed toward the range front.

The irregular trace of the axis of the synclinorium and the divergent trends of the axes of the anticlinoria suggest either that these structures developed in an inhomogeneous stress and (or) strain field or, more likely, that they were disturbed during a second episode of deformation. The structural attitudes of these major folds suggest that in this region the northwesterly Nevadan trend is superimposed on an older northerly trend (probably dating from the mid-Paleozoic Antler orogeny) and that the southern anticlinorium has been rotated counterclockwise and displaced toward the east. The northern anticlinorium more or less parallels structures farther east in early Paleozoic rocks and may have been less disturbed by a second deformation than the southern synclinorium because of buttressing by the Middle Jurassic Soldier Pass Intrusive Suite (see section "White and Inyo Mountains" under "Descriptions of the Plutonic Rocks").

The complexity of the folding and faulting in the White and northern Inyo Mountains suggests that most of the strata exposed within the map area overlie a major thrust fault—most likely the Last Chance thrust, which is exposed in a window a few kilometers east of the map area (Stewart and others, 1966). Stevens and Olson (1972) reported that along the eastern base of the Inyo Range near Tinemaha Reservoir, Cambrian and Ordovician strata overlie Mississippian to Permian strata along a folded thrust fault, but the presence of this fault is uncertain. Nelson (1966b), who first mapped the area, recognized only normal faults and subsidiary landslides, and Dunne and Gulliver (1978) restudied the area and concluded that Nelson's interpretations are essentially correct.

Another area where the distribution of rocks and structures indicates a major thrust fault is in the northern White Mountains, where, according to Crowder and Ross (1973), Proterozoic and early Paleozoic strata have been thrust westward over Mesozoic metavolcanic strata along the "Barcroft structural break." The metavolcanic rocks form a roughly U-shaped mass, open at the north end, which encloses the Cretaceous(?) granite of Pellisier Flats. On the south, east, and north sides, Cretaceous and Jurassic granitoids separate the metavolcanic rocks from Late Proterozoic and Cambrian strata. Alluvial deposits of Hamill and Chalfant Valleys border the metavolcanic rocks on the west except at one place where the metavolcanic rocks are in fault contact with Paleozoic(?) marble. Thus, the metavolcanic rocks are almost entirely surrounded by Late Proterozoic and Paleozoic rocks.

The western prong of metavolcanic rocks, discontinuous masses of Paleozoic strata, and adjacent peripheral parts of the granite of Pellisier Flats are strongly and pervasively sheared (pl. 1). Some shearing occurred during Cenozoic faulting along the range front, but some occurred before intrusion of the Cretaceous granite of Boundary Peak. Hanson and others (1987) interpreted the Jurassic quartz monzonite of Mount Barcroft, which intrudes the metavolcanic strata, and the granodiorite of Cabin Creek to have been thrust westward over the metavolcanic succession in the Early Cretaceous, after the youngest volcanogenic strata were deposited (that is, no earlier than 137 Ma) and before the granite of McAfee Creek intruded the strata (about 100 Ma). Although contemporaneity cannot be demonstrated, thrusting of Late Cambrian strata over Ordovician strata at the north end of the White Mountains and of the Early Cambrian Harkless Formation over the Proterozoic Reed Dolomite along the east side, eliminating about 1,500 m of strata, probably occurred during the same interval. The map pattern suggests that the thrusting that carried the Paleozoic and Proterozoic strata and the quartz monzonite of Mount Barcroft and the granodiorite of Cabin Creek westward over the metavolcanic strata also accounts for some of the shearing in the western lobe of metavolcanic strata and the granite of Pellisier Flats. However, Hanson and others (1987) state that D_1 foliation and lineation, which they associate with the shearing within the western lobe of metavolcanic rocks, were formed before the metavolcanic rocks were intruded by the quartz monzonite of Mount Barcroft and are therefore older. If these D_1 foliations and lineations also are present in the granite of Pellisier Flats, the granite of Pellisier Flats is older than the Jurassic quartz monzonite of Mount Barcroft and not of middle Cretaceous age as shown on plate 1. Isotopic ages for the granite of Pellisier Flats are conflicting.

STRUCTURES IN REMNANTS OF COUNTRY ROCKS OF THE EASTERN SIERRA NEVADA

Although the structures in roof pendants and septa of the eastern Sierra are better understood than those in the western metamorphic belt and adjacent remnants of country rocks, many problems of interpretation remain. Inferred unconformities between early and late Paleozoic strata, between the late Paleozoic metasedimentary strata and the Koip sequence, possibly within the Koip sequence, between the Koip and Dana sequences, and between the Koip and Minarets sequences indicate repeated disturbances.

Major and first-formed folds in the early Paleozoic strata of the Log Cabin and Mount Morrison roof pendants and the Bishop Creek septum trend north or a little east of north. These folds, which follow a distinctly different trend from folds in late Paleozoic and Mesozoic rocks, are interpreted as having formed during the mid-Paleozoic, probably coincident with the Antler orogeny of Nevada. Although the early Paleozoic strata have been affected by later deformations, these first folds have not been obliterated. Rinehart and Ross (1964) have shown that the steeply plunging folds in the early Paleozoic strata of the Mount Morrison roof pendant would parallel other early Paleozoic folds if they were rotated the amount needed to return the late Paleozoic strata to their original nearly horizontal position.

This rotation probably occurred chiefly during the Mesozoic but could have begun in the late Paleozoic. Northwest-striking folds along Nevadan trends in the late Paleozoic strata of the Mount Morrison roof pendant and the Pine Creek septum are truncated by granitoids of the Late Triassic Scheelite Intrusive Suite. The dominant structures in the Koip, Dana, and Minarets sequences are fault-bounded, westward-facing homoclines. Bateman and others (1963) interpreted the homoclinal structures to reflect the east limb of the discredited major synclinorium whose west limb was in the western metamorphic belt. The westward tilt of the western block of the Minarets sequence and angular unconformities between the Koip and Dana sequences and between the Koip and Minarets sequences indicate that the steep dips resulted from repeated tilting, the last of which occurred after deposition of the Minarets sequence (~100-Ma) and well after the Nevadan orogeny.

Tobisch and others (1986) suggested that rather than being folded into their steep west-facing positions, these strata were tilted westward on listric faults that developed in response to regional extension which accompanied emplacement of the batholith. Most of the tilting apparently occurred after deposition of the Early Cretaceous metavolcanic rocks and before deposition of the ~100-Ma Minarets sequence, but renewed tilting must have occurred after emplacement of the Minarets sequence and before intrusion of the mid-Cretaceous (~98-Ma) granodiorite of Jackass Lakes. Minor faulting postdated intrusion of the granodiorite of Jackass Lakes.

Tobisch and others (1986) also postulated that the later rise and expansion of intrusions caused the cleavages and lineations in the country rocks (fig. 3). However, faint cleavage and low strain in the Minarets sequence requires that the high strain and conspicuous cleavages and lineations in the Early Cretaceous and Jurassic metavolcanic rocks were imposed before the mid-Cretaceous Minarets sequence was erupted. The plutonic rocks adjacent to the Jurassic and Early Cretaceous metavolcanic rocks are of Late Cretaceous age and so are too young to have caused the cleavages and lineations, but the Shaver Intrusive Suite, the intrusive suite of Yosemite Valley, and the Fine Gold Intrusive Suite farther west do have appropriate ages. Further study is required to establish the cause and time of origin of the cleavages and lineations in the country rocks.

Tobisch and Fiske (1982) concluded that structures formed in the post-Paleozoic strata of the Ritter Range roof pendant during different episodes of Mesozoic deformation are essentially parallel, and that only in the Minarets sequence, which has undergone just one deformation, is it possible to designate a relative time sequence in the formation of slaty cleavage and crenulations. These observations together with the parallel outcrop patterns of the intrusive suites indicate that the Nevadan trend was imposed over a long timespan in the Mesozoic and is not solely the product of the Late Jurassic Nevadan orogeny.

STRUCTURES IN THE KINGS SEQUENCE

Within the map area, most of the remnants of metamorphic rocks that have been assigned to the Kings sequence are small, and structural patterns have been determined only within the Dinkey Creek roof pendant and in the Strawberry Mine area (Kistler and Bateman, 1966; Nokleberg, 1981b). Kistler and Bateman (1966) recognized three tectonic episodes in the Dinkey Creek roof pendant, the first identified with strongly overturned folds with axial surfaces striking about N. 5° E. and dipping 45° W., the second with open folds that have near-vertical axial surfaces striking N. 20° W., and the third with sporadically distributed open folds of small amplitude with near-vertical axial surfaces striking N. 60° W. The first and second sets are coaxial, and their axes plunge 10° to 20° W. The strata in the Dinkey Creek roof pendant closely resemble strata in the Boyden Cave roof pendant just south of the map area along the Kings River, which have yielded Early Jurassic fossils (Moore and Dodge, 1962; Jones and Moore, 1973). Nevertheless, Kistler and Bateman (1966) left undecided the question of the age of the strata in the Dinkey Creek roof pendant because the north-to-northeast strike of the axial surfaces of the first folds parallels the strike of axial surfaces of folds that were formed during the mid-Paleozoic orogeny. It is quite possible that the first and second folds were formed during the same deformation and that the strike of the axial surfaces of the first folds do not reflect the fold pattern of the early Paleozoic strata. Rotation around their common axis would bring second-fold axial surfaces parallel to first-fold axial surfaces. The first folds are spatially associated with and parallel to a zone of thrust faulting, whereas the second folds are largely wrinkles in the lower plate in front of the thrust fault. In this interpretation, the first folds are merely overturned second folds in the region of greatest strain. Intrusion of the Dinkey Creek Granodiorite along the northwest side of the pendant could have caused southeastward thrusting, which overturned earlier open folds.

The strata in the Strawberry Mine area are chiefly calc-silicate hornfels and marble presumed to be of Early Jurassic age (Nokleberg, 1981b). They are surrounded by metavolcanic rocks of the mid-Cretaceous Minarets sequence and have been cut by metaigneous intrusions. Nokleberg (1981b) recognized four generations of folds, but the latest one is poorly represented. The first three fold systems are similar in their orientations to those in the Dinkey Creek roof pendant. The first fold system affects only the Jurassic metasedimentary strata, whereas the two younger generations of folds affect both the metasedimentary strata and the mid-Cretaceous metaigneous rocks. Although the first folds have been strongly disturbed by the succeeding deformations, Nokleberg (1981b) interpreted them as moderately appressed and as having had axial surfaces originally striking east or northeast. On the basis of their attitude, he assigned these folds a Middle Jurassic age and related them to folds of similar orientation in the Calaveras Formation of the western metamorphic belt and in the Boyden Cave roof pendant just south of the map area (Nokleberg, 1981b). Nokleberg assigned the next younger generation of folds, which strike N. 25° W., to the middle Cretaceous because they are cut by the early Late Cretaceous (mid-Cretaceous) granodiorite of Jackass Lakes. He assigned a middle or Late Cretaceous age to the youngest folds, which strike N. 65°–90° W., because the granodiorite of Jackass Lakes was affected by the deformation that caused the folding.

STRUCTURES WITHIN THE WESTERN METAMORPHIC BELT

Although the strata east of the Melones fault zone in the western metamorphic belt are younger toward the west, the succession of units may be tectonic rather than stratigraphic. Three possible interpretations of the structural relations are shown in figure 5. The strata forming the overturned limb of a west-vergent shear fold that is truncated on the east by the Calaveras-Shoo Fly thrust of Schweickert (1977) and on the west by the Melones fault zone, both of which are represented as east-dipping reverse faults, are shown in figure 5*A*. Successively younger strata thrust eastward over older strata and then modified

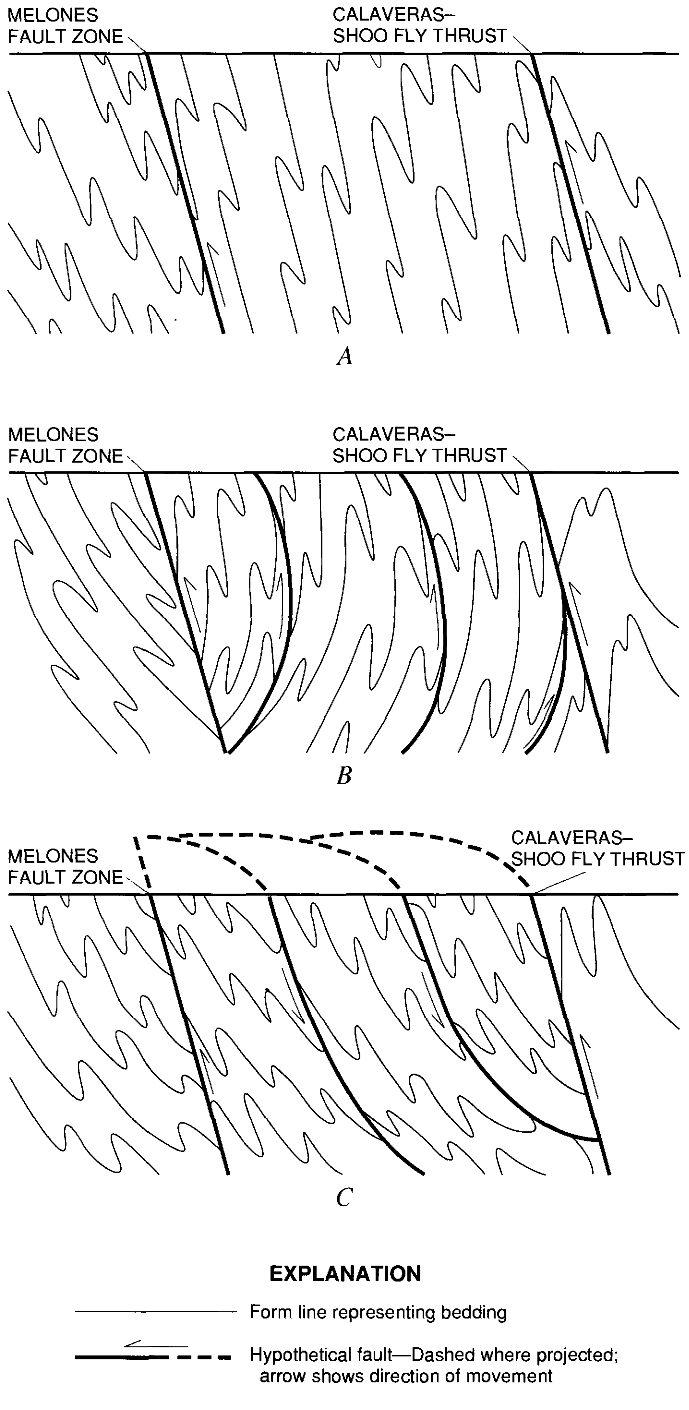


FIGURE 5.—Structure in western metamorphic belt between Melones and Calaveras-Shoo Fly faults. Three possible interpretations (*A–C*) are shown; for details, see text.

by backfolding and westward-directed folds and faults are shown in figure 5*B*. This agrees with Moores and Day's (1984) interpretation of relations in the northern Sierra Nevada; however, early eastward overthrusting is not required to account for the known relations within the map area. Younger units thrust eastward beneath older units, producing a pattern similar to that in accretionary prisms, are shown in figure 5*C*. In the first two interpretations the strata are younger toward the west, whereas in the third interpretation they face predominantly east, contrary to the age relations indicated by the phyllite of Briceburg and the phyllite and chert of Hite Cove, which appear to constitute a relatively unbroken sequence.

The Melones fault zone extends the full length of the western metamorphic belt and bisects it longitudinally (Clark, 1960). It has been considered by some to mark the boundary between accreted slabs of oceanic, largely island-arc-related strata on the west and autochthonous or earlier accreted strata on the east. The principal evidence within the map area that indicates that the Melones fault zone represents a major dislocation is the presence of serpentinite lenses and the slightly lower metamorphic grade of the Mariposa Slate than that of the phyllite of Briceburg. Schweickert and others (1982) state that in the northern Sierra Nevada, south of Placerville, the Melones fault zone forms the boundary between two related belts of Jurassic rocks, a slate belt on the west and a phyllite-greenschist belt on the east, and that it is not a major suture. The Late Jurassic age of the strata on both sides of the fault zone in the map area is in agreement with this interpretation. On the other hand, studies by Scott Paterson (written commun., 1985) indicate changes in strain, structure, lithology, and metamorphic grade across the fault, which support major ductile movement.

Only the south end of the Melones fault zone lies within the map area, and it is truncated near Mariposa by the Bass Lake Tonalite. There, the principal line of movement presumably was along the contact between the Mariposa Formation and the greenstone of Bullion Mountain. Discontinuous lenticular masses of serpentinite along this contact and for distances of as much as 2 km on either side indicate that the zone affected by faulting is at least several kilometers wide. Forcible intrusion of the Bass Lake Tonalite bent the south end of the serpentinite lens that lies along the contact between the Mariposa Formation and the greenstone of Bullion Mountain sharply west, in which direction it continues for about 3 km before being cut off by the tonalite (pl. 1).

South of Mariposa, the former trace of the Melones fault zone must pass between sedimentary strata along the west edge of the map area, which are continuous northward with strata west of the Melones fault zone, and the Coarsegold roof pendant, the O'Neals lobe, and the Tick-Tack-Toe roof pendant (fig. 2), which are composed of strata correlative with strata east of the Melones fault zone. This distribution of strata requires that the ancestral trace of the fault zone bends toward the south and passes into the Central Valley between the Adobe Hill and Tick-Tack-Toe roof pendants. Bateman, Busacca, and Sawka (1983) suggested that the leucotonalite of Ward Mountain (Ward Mountain Trondhjemite, of this report) occupies the ancient trace of the fault zone and that after emplacement of the Bass Lake Tonalite renewed activity at depth in some way may have guided emplacement of the Ward Mountain magma. The tongue of trondhjemite that extends northwest from the northern pluton of Ward Mountain and the thin crescent-shaped metamorphic inclusion within the Bass Lake Tonalite near the north end of the tongue are especially suggestive of this location for the trace of the fault zone.

Schweickert (1977) traced the Calaveras-Shoo Fly thrust southward from the northern Sierra Nevada, but its position within the map area is uncertain. If present, the thrust may lie between the quartzite of Pilot Ridge and the phyllite and chert of Hite Cove. The principal evidence that the thrust coincides with this contact and is not farther east is the Paleozoic isotopic ages for orthogneisses that reportedly intrude strata west and northwest of the map area and are believed to be correlative with the quartzite of Pilot Ridge (Sharp and others, 1982).

In addition to the presence of the Melones fault zone and the Calaveras-Shoo Fly thrust fault in the western metamorphic belt, unpublished studies by S.R. Paterson (oral commun., 1985) suggest the possibility of faulting between the greenstone of Bullion Mountain and the phyllite of Briceburg. Fault movement could also have occurred within the phyllite of Briceburg north of the Merced River along a northwest-trending line defined by divergent cleavages. West of this line, cleavage strikes about N. 85° W., whereas east of the line it strikes N. 30° W. on the average. The significance of this line has not been determined. A fault may separate the cleavage domains (pl. 1), but it is also possible that the divergent cleavages are of different ages or that the line separating them marks a sharp flexure.

TECTONIC EPISODES

Two and possibly three major episodes of regional deformation occurred during the Paleozoic and early Mesozoic; deformation accompanied emplacement of the batholith during the middle and late Mesozoic; and tectonic movements of local extent or of lesser intensity occurred during the Late Cretaceous and Cenozoic. Deformation that occurred during emplacement of the batholith is discussed in the sections "Deformation Associated With Emplacement of Intrusions" and "Shifts in the Locus of Magmatism."

The most reliable means of determining the time when a tectonic disturbance occurred is by bracketing it between the ages of strata above and below an angular unconformity, between the ages of a deformed stratigraphic succession and a body of intrusive rock that was emplaced after the succession was deformed, and between isotopically dated intrusive rocks, the older of which was involved in an episode of regional deformation. These are the criteria used here. Some use has been made of the attitudes of planar and linear structures in correlating and distinguishing different fold systems (Kistler, 1966b; Kistler and Bateman, 1966; Brook, 1977; Russell and Nokleberg, 1977; Nokleberg and Kistler, 1980). However, these criteria must be used with extreme caution because the attitudes of fold axes and axial surfaces of the same age can vary from place to place and because in some places folds of different ages are very nearly parallel (Tobisch and Fiske, 1982).

The following episodes of deformation have been identified with moderate, but not absolute, certainty.

MISSISSIPPIAN (ANTLER?) OROGENY

Evidence of the Antler orogeny within the map area includes (1) the unconformity in the Inyo Mountains between the Devonian Sunday Canyon Formation and the overlying Mississippian Perdido Formation, (2) a probable unconformity coincident with the Laurel-Convict fault in the Mount Morrison roof pendant, which separates Ordovician and Silurian strata on the east from Pennsylvanian and Permian(?) strata on the west, and (3) regular north- to northeast-trending folds in the early Paleozoic strata of the Bishop Creek, Mount Morrison, and Log Cabin roof pendants and of the White Mountains, which contrast with the northwesterly trends in late Paleozoic and Mesozoic strata.

EARLY TRIASSIC (SONOMAN?) OROGENY

An orogeny in the Early Triassic, which may be the Early Triassic and (or) Late Permian Sonoman orogeny, is indicated by the unconformity in the Ritter Range roof pendant and in the Mount Morrison roof pendant and its northern extension (the Saddlebag Lake septum) between Paleozoic metasedimentary strata and the Mesozoic Koip sequence. The northwest-trending folds in the Pine Creek septum and in the late Paleozoic rocks of the Mount Morrison roof pendant presumably formed during this deformation but could have formed during the Late Triassic deformation.

LATE TRIASSIC DEFORMATION

Kistler and Swanson (1981) proposed an unconformity between what they called the younger and older successions of the Koip sequence. According to them, the older part of the Koip sequence was already folded when it was intruded by the granite of Lee Vining Canyon, which has a U-Pb age of 210 Ma and a K-Ar age of 212±8 Ma and is older than isotopic ages on the younger part of the Koip sequence. Similar relations are present in the Pine Creek septum where metavolcanic rocks faulted against isoclinally folded Paleozoic strata are intruded by the ~200-Ma Tungsten Hills Granite. The precise age of this deformation is suspect because of disagreement between the U-Pb ages and the Rb-Sr ages for the Koip sequence. Nokleberg and Kistler (1980) attributed folding in so-called Calaveras strata and in the strata of the Boyden Cave roof pendant (just south of the map area) and of the Strawberry Mine area to this deformation. Sharp and others (1982) reported a possibly coeval episode of deformation about 215±15 Ma that affected the isotopic characteristics of several bodies of orthogneiss which previously had intruded the early Paleozoic Shoo Fly Complex of Schweickert (1977).

LATE JURASSIC NEVADAN OROGENY

The term "Nevadan orogeny" has been used loosely to designate deformations that occurred during the Late Jurassic. In the northern Sierra Nevada, it has been identified with an unconformity between Late Jurassic strata and the Knoxville Formation of latest Jurassic (Tithonian) age, which is considered to be the basal unit of the Great Valley sequence. Within the map area, the youngest stratigraphic unit involved in the Nevadan orogeny is the Late Jurassic Mariposa Formation.

The Mariposa Formation contains late Oxfordian and early Kimmeridgian fossils and therefore was deformed after the beginning of the Kimmeridgian Stage and probably before the Tithonian. According to the Decade of North American Geology 1983 time scale (Palmer, 1983), the Kimmeridgian lasted from about 156 to 152 Ma. Schweickert and others (1984) summarized data that bear on the age of the Nevadan orogeny, including the isotopic ages of granitic rocks that were emplaced before and were involved in the Nevadan orogeny and of some that were emplaced shortly thereafter, and concluded that the orogeny occurred 155±3 Ma. Within the map area, in addition to the age of the Mariposa Formation, the most confining ages for the Nevadan orogeny are U-Pb ages of 166 and 163 Ma for the tonalite of Granite Creek, which was deformed and lineated, presumably during the Nevadan orogeny, and 151 Ma for the contiguous undeformed granite of Woods Ridge.

LATE EARLY CRETACEOUS DEFORMATION

An angular unconformity at the base of the 100-Ma Minarets sequence requires that the underlying Koip sequence was tilted westward before the Minarets sequence was erupted and after the Early Cretaceous metavolcanic strata conformable with the Koip sequence were deposited. An accurate isotopic age for the Dana sequence, which also rests unconformably on the Koip sequence, would narrow the timespan during which the westward tilting could have occurred. Tobisch and others (1986) suggested that the tilting, rather than being a compressional event, occurred by rotation on listric faults and reflects regional extension rather than shortening.

MID-CRETACEOUS DEFORMATION

Most of the westward tilting and minor folding of the large coherent block of the ~100-Ma Minarets sequence centered on the Minarets must have taken place before intrusion of the 98-Ma granodiorite of Red Devil Lake because displacement on the fault that bounds this block on the west is much greater in the metavolcanic rocks than in the granodiorite of Red Devil Lake. Movement after the granodiorite of Red Devil Lake was emplaced probably was minor. The smaller blocks farther west may have been parts of a second westward-rotated block, which was jostled and fragmented when it was intruded by the ~98-Ma granodiorite of Jackass Lakes.

Faulting that produced approximately parallel shear zones within the granodiorite of Jackass Lakes, farther west, within the granite of Shuteye Peak and the granodiorite of Illilouette Creek, and farther south, at Courtright Reservoir within the Dinkey Creek Granodiorite, could also have occurred during this general time.

LATE CRETACEOUS (AND TERTIARY?) DEFORMATION

Shear zones similar to those within the granodiorite of Jackass Lakes, the granite of Shuteye Peak, the granodiorite of Red Devil Lake, and the Dinkey Creek Granodiorite also occur within the younger Lake Edison Granodiorite and the Mono Creek Granite. These rocks are too young for the shearing to have been contemporaneous with shearing that accompanied tilting of the Minarets sequence but could be the result of similar movements at a later time. A mineral foliation that strikes about N. 80° W. across the ~90-Ma Tuolumne Intrusive Suite and into older rocks must also be younger than the deformation that affected the Minarets sequence.

Still younger deformation is indicated by strike-slip movements on many regional joints (Lockwood and Moore, 1979). The joints are independent of individual intrusions and probably were formed after the last intrusions were emplaced, in the Late Cretaceous when the Sierra Nevada batholith was being uplifted and eroded. The strike-slip movements occurred still later, probably during the Tertiary when extensional strain occurred in the Basin and Range province to the east. The most recent movements have been uplift and westward tilting of the Sierra Nevada, which continue to the present.

HIERARCHICAL ORGANIZATION OF GRANITIC UNITS

The Sierra Nevada batholith is composed of hundreds of separate granitic plutons that must be assigned positions within a hierarchical system in order to deal effectively with the larger problems of the batholith. To arrange the granitic rocks in a hierarchy requires a vast amount of data concerning the ages and affiliations of the individual plutons.

Because the data now available are limited, the organization of the granitic rocks given herein must be considered provisional and subject to modification. Most of the plutons within the map area are assigned to lithodemes, and lithodemes are assigned to intrusive suites in accordance with the guidelines of the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983). Plutons that were emplaced between the major intrusive events or whose affiliations are uncertain are left unassigned.

Most workers have referred to the dominantly granitic terrane of the Sierra Nevada as a batholith, but Cloos (1936), following German usage, referred to it as "der Sierra-Nevada-pluton" and to the discrete units of which it is composed as "partiplutons." The generally accepted definition of a batholith is a plutonic terrane encompassing at least 100 km^2 composed predominantly of medium- to coarse-grained granitic rocks, whereas a pluton is defined simply as an igneous intrusion. To reconcile these terms and also to fulfill a need for clear terminology, in recent reports the predominantly granitic terrane of the Sierra Nevada is called the Sierra Nevada batholith, and the discrete units of which it is composed are called plutons (Bateman, 1965b).

PLUTONS

As used in this report, a pluton is a body of intrusive rock that is expressed at the surface by a single exposure that is continuous except for a veneer of younger cover. It is bounded at the surface by sharp contacts or by gradational zones caused by the mixing of magma and older wallrocks or of younger magma with the rocks of the pluton, and it is unbroken by continuous internal contacts. However, discontinuous contacts between facies that differ slightly in composition or texture may be traceable for short distances. The probability that the rock exposed in two adjacent but disconnected areas was once continuous or is now continuous at depth does not allow the rock in the two outcrop areas to be designated as two parts of one pluton. Only a few plutons within the map area have been (informally) named, using prominent features within them. If a pluton assigned to a lithodeme (see section "Lithodemes") contains the type locality of the lithodeme, it is given the same name as the lithodeme. For example, the part of the Ward Mountain Trondhjemite (lithodeme) that contains Ward Mountain (type locality) is called the Ward Mountain pluton, whereas the second pluton of this lithodeme is called the Experimental Range pluton (Bateman and Busacca, 1982).

The plutons of the central Sierra Nevada range in outcrop area from less than 1 km^2 to more than 1,000 km^2. Most of the large plutons are elongate in a northwesterly direction, but many smaller ones, especially those with leucocratic compositions, are subcircular. The rock in most small plutons is compositionally and texturally almost homogeneous, whereas the rock in most larger plutons varies from place to place. It is common for melanocratic rock near the margin of one of the larger plutons to grade inward to more leucocratic rock. However, some of the larger plutons grade compositionally from one side or one end to the other, and some are more leucocratic upward. Plutons are either in sharp contact with one another or are separated by septa (screens) of metamorphic or older igneous rocks or by younger dikes. Where plutons meet, the contact between them usually is obvious because of differences in composition, grain size, and (or) texture. The relative ages of plutons in contact can generally be determined by means of the crosscutting relations of dikes, by the presence of inclusions of older rock in younger, or by truncated fabric, structures, dikes, and compositional or textural patterns in the older pluton. During geologic mapping, considerable effort was expended to determine the relative ages of plutons in contact, but some age relations were not determined because the plutons meet along smooth, featureless, or poorly exposed contacts. Relative ages are especially difficult to determine in the northwestern part of the map area where deep alluvial cover and slope wash conceal contacts.

LITHODEMES

In the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983), the term "lithodeme" is defined as a mappable unit of plutonic and highly metamorphosed or pervasively deformed rock and is a term equivalent in rank to "formation" among stratified rocks. The formal name of a lithodeme consists of a geographic name followed by a descriptive term that denotes the average modal composition of the rock—for example, Cathedral Peak Granodiorite. The informal name of a lithodeme consists of a lithologic term followed by a geographic name—for example, the granodiorite of Kuna Crest. Two or more plutons composed of rock of similar composition, texture, and age relative to other rocks in the same general area are assumed to be outcrops of the same lithodeme—

that is, to have solidified from a common magma and, at the time of their intrusion, to have been connected at depth. Older plutons that are composed of rock of similar composition, texture, and age relations and that are now separated only by younger intrusions are assumed originally to have been parts of a single body; nevertheless, they are now considered to be separate plutons.

In this report, lithologic designations for the intrusive-rock units (fig. 6) generally follow the classification recommended by the International Union of Geological Sciences (Streckeisen, 1973), which required changing some lithologic designations used in earlier reports. To conform to this classification, most rocks previously called quartz monzonite are now called granite and most rocks previously called alaskite are now called leucogranite. However, in plots of modes in this report (mostly in the section "Descriptions of the Plutonic Rocks"), the upper limit for quartz is placed at 50 percent rather than 60 percent because none of the modes for Sierran granitoids plot between 50 and 60 percent. Moreover, in this report, rocks whose compositions plot in the monzogranite subfield of granite are referred to simply as granite, whereas rocks whose modes plot in the syenogranite subfield are referred to as syenogranite.

Because some units are small and (or) unimportant or because only part of the unit is exposed within the area being mapped (and hence its full dimension and average composition are unknown), many lithodemes have been given informal names. For example, Bateman and Wones (1972a, b) called a rock unit in the southwestern part of the Huntington Lake quadrangle the granodiorite of Blue Canyon. Later mapping farther west showed that this rock is part of one of the largest lithodemes in the central Sierra Nevada and that its average composition is tonalite rather than granodiorite; therefore in subsequent reports it was called the tonalite of Blue Canyon. Because the name Blue Canyon has been preempted in formal nomenclature, this lithodeme is formally named the Bass Lake Tonalite in this report.

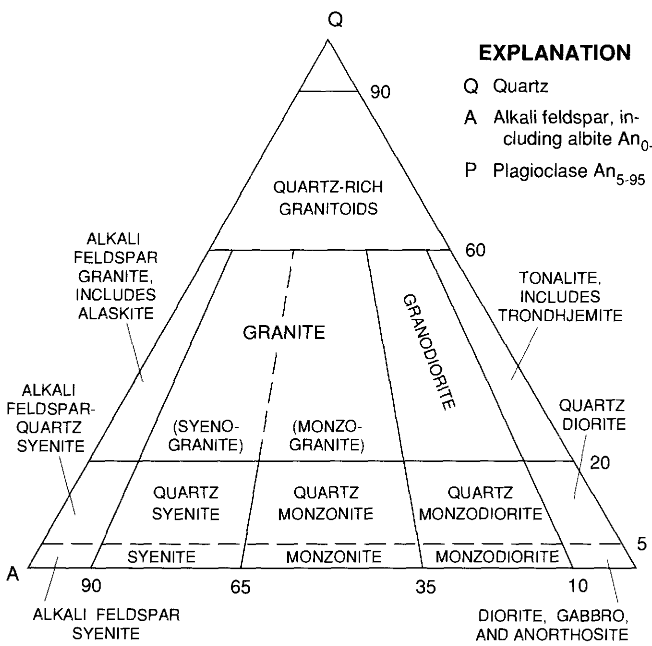


FIGURE 6.—Q-A-P diagram showing IUGS classification of plutonic rocks (Streckeisen, 1973).

INTRUSIVE SUITES

Lithodemes and plutons that crop out in the same general area and that have similar ages and similar or related compositions and fabrics may be combined into units of higher rank called intrusive suites. This term was adopted by the North American Commission on Stratigraphic Nomenclature (1983) for plutonic and high-grade metamorphic rocks as the term equivalent to "group" for stratified rocks. The formal name of an intrusive suite consists of a geographic name followed by "Intrusive Suite"—for example, the Tuolumne Intrusive Suite. However, many intrusive suites in the central Sierra Nevada have not been given formal names because the assignment of lithodemes to these intrusive suites involves considerable uncertainty; future reassignments can be expected. The phrase "intrusive suite of" precedes the geographic name of an informally named suite—for example, intrusive suite of Washburn Lake.

The underlying concept of an intrusive suite is that all included units are in some manner cogenetic, though not necessarily comagmatic, and that they are the products of a single fusion episode (Presnall and Bateman, 1973). The most unequivocal suites are those that have distinctive modal, chemical, isotopic, and textural characteristics and in which the units that make up the suite are transitional to one another. Such suites are zoned compositionally and texturally and may exhibit partial or complete nested patterns in which more melanocratic rock in the margins gives way inward to younger, more leucocratic rocks. The Tuolumne Intrusive Suite, originally called the Tuolumne Intrusive Series by Calkins (1930), was the first intrusive suite to be identified in the Sierra Nevada and is the most firmly established. It is a splendid example of the ideal kind of intrusive suite (Bateman and Chappell, 1979).

Most of the intrusive suites shown on plate 1 are less than ideal. They are composed of rocks that crop out in the same general area and have similar or identical isotopic ages; in most of them, the younger units are progressively more leucocratic. However, the units may not have enough similar characteristics and transitions one to another to unambiguously demonstrate their consanguinity. Some of the units that compose these less-than-ideal suites may have solidified from magmas that were derived from different parts of an inhomogeneous parent magma or from different coexisting parent magmas rather than from a common homogeneous parent magma.

In this report, lithodemes and some plutons are assigned to intrusive suites on the basis of the following empirical criteria: (1) The lithodemes and plutons are in the same general area, and some or all may be contiguous; (2) they have similar textural, mineralogical, chemical, or isotopic characteristics; (3) successively younger units are generally, but not invariably, more felsic; (4) units may be arranged concentrically with the more felsic units in the interior; (5) if deformed as a result of external forces, all the units have undergone the same number and kinds of deformation; (6) textural and compositional changes among units are generally in the same order as they are in concentrically zoned plutons of the same compositional range; (7) screens (septa) of metamorphic or older igneous rocks are more likely to separate intrusive suites than lithodemes of the same intrusive suite; (8) the units of a younger suite may truncate dike swarms or cataclastic or mylonitic zones in units belonging to an older suite; and (9) isotopic ages of the members of an intrusive suite indicate true age differences of no more than a few million years.

ROOF PENDANTS, SEPTA, AND INCLUSIONS

Because the terms "roof pendant," "septum," and "inclusion" are confusing, their usage in this report is defined as follows:

Roof pendant: A mass of metamorphic rock that is entirely surrounded by plutonic rock and presumed to be a downward projection of rocks that overlaid the plutonic rocks before erosion. Roof pendants may be enclosed within a single intrusion or be bounded by two or more intrusions. Usually they are large—several square kilometers—but size is not an essential criterion.

Septum: A mass of rock (usually metamorphic) that lies between two intrusions. It may be of any size and shape but usually is elongate in outcrop. It may be discontinuous horizontally and (or) vertically. "Screen" is an equivalent term. A septum may also be a roof pendant, but it is not necessarily a roof pendant.

Inclusion: A mass of rock (usually metamorphic) that is surrounded by rock of the same intrusion and is assumed to have been covered by the same intrusive rock before erosion.

SAMPLING, ANALYTICAL PROCEDURES, AND REPORTING OF DATA

During the mapping of most of the 15-minute quadrangles that lie within the map area (pl. 1), an effort was made to collect samples of the plutonic rocks about 1.6 km apart, but owing to the rough terrane no attempt was made to follow a rigid pattern. Care was taken to collect fresh and representative rock samples at each locality.

CHEMICAL ANALYSES

A summary of major-oxide analyses, CIPW norms, modes, and bulk specific gravities of the samples of plutonic rocks collected within the map area from 1953 to 1983 has been published (Bateman, Dodge, and Bruggman, 1984), and only chemical analyses needed to support interpretations are given in this report. Only carefully selected samples were analyzed chemically, especially during the early stages of mapping. However, the number of chemical analyses increased as mapping progressed westward across the batholith because new methods and improved techniques reduced costs and increased the analytical capacity of the chemical laboratories at the U.S. Geological Survey. Some chemical analyses published in early reports were made using classical wet-chemical methods (Peck, 1964), which are both time consuming and costly, but most analyses for major elements were made by the "rapid method" of Shapiro and Brannock (1962). Recently, increasing numbers of analyses have been made by X-ray fluorescence. Because only in recent years have quantitative analyses of minor elements become readily available, few are given in this report.

MODAL ANALYSES

In this report modal analyses are included where appropriate. The volume percentages of quartz, alkali feldspar, plagioclase, and total mafic minerals were determined for almost all the samples of granitic rocks by counting regularly spaced points on sawed and polished slabs on which plagioclase was stained red and alkali feldspar yellow (Norman, 1974). Some modes determined in the early 1950's for rocks from the eastern Sierra Nevada were counted on small slabs with areas of less than 50 cm^2, and a few modes of fine-grained rocks were counted on thin sections. However, modes published after 1965 were made by counting a minimum of 1,000 points on slabs of at least 70 cm^2. Where the percentages of biotite, hornblende, and accessory minerals were determined, point counts of these minerals were made on thin sections and apportioned over the total mafic content as determined on stained slabs.

The number of point counts made on the stained slabs would have been sufficient to permit assigning limits of error of less than ±3 percent at the 95-percent-confidence level, the limits increasing with decreasing abundance of a mineral, except that the spacing of points was generally less than half the distance across individual grains. According to Van der Plas and Tobi (1965), points must be spaced at least as far apart as the distance across the largest grains for accurate statistical treatment. Nevertheless, modes of rocks collected within a few meters of one another generally are in good agreement, and gradual changes over large areas indicate systematic distribution patterns.

The relative modal abundances of quartz, alkali feldspar, and plagioclase are plotted on quartz-alkali feldspar-plagioclase (Q-A-P) diagrams (fig. 6) in which the fields of the different granitic rocks are shown, according to the International Union of Geological Sciences (IUGS) classification (Streckeisen, 1973). For most rock units, the color index (sum of volume percentages of mafic and accessory minerals) is shown by histograms, but if the volume percentages of hornblende and biotite were determined separately, such as for the Bass Lake Tonalite (fig. 64) and the Tuolumne Intrusive Suite (fig. 74), their abundances are shown on auxiliary diagrams.

ISOTOPIC AGES

Isotopic ages of most of the more important lithodemes and plutons have been determined by the U-Pb, K-Ar, and (or) Rb-Sr methods, but many more isotopic ages are needed to establish the ages of all the rocks. The U-Pb ages are generally better indicators of times of crystallization than K-Ar ages because the temperature of closure is significantly higher than for K-Ar ages. The presence of inherited zircon has not been a serious problem in dating the plutonic rocks but may account for some anomalous ages. Unless otherwise indicated, the U-Pb ages cited are ^{206}Pb-^{238}U ages on zircon. Concordant U-Pb ages are ones in which $^{206}Pb/^{238}U$ values agree with $^{207}Pb/^{235}U$ values within the limits of analytical error, which is estimated to be ±2 percent (Stern and others, 1981). R.W. Kistler (written communs., 1984–87) determined a few Rb-Sr isochrons for both plutonic and metavolcanic rocks. The ages determined for plutonic rocks by this method appear to be generally reliable, but the ages determined for volcanic rocks have been variable. Redistribution of Rb and Sr by volatiles during or after consolidation may explain some anomalous Rb-Sr ages for volcanic rocks.

K-Ar ages were determined on either biotite or hornblende. The blocking temperature for hornblende is higher than for biotite, and hornblende ages generally are more reliable indicators of the magmatic age of the rock. However, both types of ages may be younger than the age of crystallization because of slow cooling to the blocking temperature or because of reheating by younger intrusions. Some K-Ar hornblende ages are inexplicably older than U-Pb ages for the same rocks.

I-TYPE AND S-TYPE AND MAGNETITE-SERIES AND ILMENITE-SERIES GRANITOIDS

Most Sierran granitoids are I types in the classification of Chappell and White (1974) and belong to the magnetite series of Ishihara (1977). However, a large part of the Fine Gold Intrusive Suite belongs to the ilmenite series, and further study may show that the Cretaceous granites of the White Mountains are S types in the sense in which they were originally defined (Chappell and White, 1974).

"S type" refers to a sedimentary source and "I type" to an igneous source, but the criteria that distinguish these two types of granitoids are complex and have been modified from time to time. In the original report on I-type and S-type granitoids, Chappell and White (1974) distinguished I- from S-type granitoids by the following criteria: In I-type granitoids Na_2O is greater than 2.2 percent and mol $[Al_2O_3/(Na_2O+K_2O+CaO)]$ is less than 1.1; they contain CIPW-normative diopside or less than 1.1 percent CIPW-normative corundum; they cover a broad spectrum of compositions; and they show regular interelement variations (on linear-variation diagrams) within plutons. They envisaged both I- and S-type granitoids as having their source regions within the crust. Recently White and others (1986) narrowly redefined S types to agree with the properties of cordierite-bearing S-type granites in the Lachlan Fold Belt of southeastern Australia, where S-type granites were first identified. According to these more restrictive criteria, S-type granitoids are uncommon rocks, and none are present within the Mariposa 1° by 2° quadrangle (pl. 1).

Ishihara (1977) defined magnetite-series granitoids as containing greater than 0.1 volume percent magnetite and ilmenite-series granitoids as containing less than 0.1 percent, but in recent studies the two series have been distinguished by their magnetic susceptibility, the boundary being placed at 50×10^{-6} emu/g (Ishihara, 1979). Ishihara (1977) interpreted ilmenite-series granitoids as having originated in the middle to lower continental crust, where they incorporated carbonaceous material that reduced Fe^{+3} to Fe^{+2}, and magnetite-series granitoids as having been generated in a deep level of the lower crust or upper mantle. However, preliminary studies by P.C. Bateman and F.C.W. Dodge suggest that regional variations of magnetic susceptibility in the granitoids of the central Sierra Nevada result chiefly from intrinsic differences in the oxidation ratios of the source components for the magmas.

PHYSICAL FEATURES OF THE GRANITIC ROCKS

MINERALS

The minerals of Sierran granitic rocks are typical of granitoids in the compositional range of tonalite or quartz diorite to granite. The essential minerals are quartz, alkali feldspar, and plagioclase; the varietal minerals are clinopyroxene, hornblende, and biotite; and the common accessories are opaque minerals, titanite, apatite, and zircon. Orthopyroxene, monazite, thorite, and allanite are present in some rocks. Epidote, chlorite, and sericite are common alteration products. In the western foothills, muscovite is present in the Ward Mountain Trondhjemite but may be secondary. Cordierite, andalusite, sillimanite, and garnet have not been reported except locally in the granite of Dinkey Dome, an epizonal pluton that was water saturated before it was completely crystallized and that also may have assimilated aluminous metasedimentary rocks.

QUARTZ

Quartz occurs in dark-gray to white anhedral grains that have a wide range of size and shape. Tiny fluid inclusions are common, whereas mineral inclusions are uncommon. When thin sections are viewed by polarized-light microscopy, the larger grains in many rocks show irregular or undulatory extinction or a mosaic extinction revealing sharply defined and diversely oriented polygonal components. This extinction behavior reflects late- or post-crystallization strain; quartz responds to strain more readily than other minerals and in many rocks is the only indicator of strain.

ALKALI FELDSPAR

Alkali feldspar is white, light gray, or pinkish in hand specimen and forms subhedral grains in most undeformed granitoids, but it is interstitial in the low-potassium rocks of the Fine Gold Intrusive Suite and forms euhedral to subhedral megacrysts in some rocks that contain relatively large amounts of potassium. In thin section, alkali feldspar generally has quadrille structure (grid twinning, gridiron structure, or grating structure), but in some rocks it is inconspicuous or absent. Most alkali feldspar is also perthitic, and plagioclase grains bordering the alkali feldspar are rimmed with exsolved albite.

Microprobe studies of alkali feldspar in the Mount Givens Granodiorite show that it contains only 4 to 12 percent albite and less than 1 percent anorthite—far less than the amounts required for the alkali feldspar to coexist in equilibrium with the plagioclase in the magma (Seck, 1971; Stormer, 1975; Bateman and Nokleberg, 1978). Similar results have been obtained (Noyes, Frey, and Wones, 1983; Noyes, Wones, and Frey, 1983) for the granodiorites of Red Lake and Eagle Peak. Clearly, the alkali feldspar in these rocks, and by inference in other granitic rocks as well, has at least partially reequilibrated at subsolidus temperatures. Albite has been expelled to form perthitic intergrowths and albitic rims on plagioclase, as similarly observed by Tuttle and Bowen (1958).

ALKALI FELDSPAR MEGACRYSTS

Megacrysts are abundant in rocks whose compositions are transitional between granite and granodiorite. Their size and shape varies widely from well-formed crystals several centimeters on a side to small, poorly formed grains 1 cm across. Some of the largest and best formed megacrysts are in the Wheeler Crest and Cathedral Peak Granodiorites. The megacrysts in the Wheeler Crest Granodiorite are generally tabular parallel to (010), and common dimensions in outcrop are 1 to 2 cm wide and as long as 10 cm. Megacrysts in the Cathedral Peak Granodiorite are more nearly equant and appear blocky; 3 by 5 cm are common dimensions in the outer part of the intrusion, and 1 by 1.5 cm is usual in the inner part where megacrysts are sparse. Megacrysts of these sizes and dimensions are clearly distinguishable from groundmass feldspar (fig. 7). In other intrusions, such as the Half Dome, Mount Givens, and Tinemaha Granodiorites, megacrysts are smaller, less well formed, and grade to groundmass feldspar in a seriate texture. Most megacrysts are twinned, generally according to the Carlsbad law, and selected growth zones that parallel the crystal faces commonly contain abundant inclusions of all the other minerals in the rock. Tiny tabular or elongate plagioclase crystals oriented with their longer dimensions parallel to crystal faces are the most common mineral inclusions. The megacrysts commonly are accentuated by peripheral concentrations of mafic minerals, and the faces of the megacrysts are uneven because of interference with bordering crystals during late stages of growth. In some rocks, alkali feldspar anastomoses from the margins of megacrysts into the surrounding equigranular rock.

The rocks in which megacrysts occur plot on or near the boundary between the granodiorite and granite fields on a Q-A-P diagram. They are rarely the last or most leucocratic rocks in a suite to solidify; commonly they are succeeded by equigranular granite. Thus, inward from the margins of the concentrically zoned Tuolumne Intrusive Suite, small megacrysts appear about midway in the Half Dome Granodiorite, increase in size and abundance to the contact between the Half Dome and Cathedral Peak Granodiorites, then decrease; the inner part of the Cathedral Peak Granodiorite and the Johnson Granite Porphyry contain only a few small scattered megacrysts. This pattern is duplicated entirely or partly in other zoned plutons or intrusive suites where the succession of units is well established (Reesor, 1958; Wagener, 1965; Bateman and Wones, 1972a, b; Chappell and White, 1976; Bateman and Chappell, 1979). Lockwood (1975) determined that the abundance of megacrysts decreases inward in the southern part of the Mono Creek Granite and is not related to elevation (fig. 8). In the Tuolumne Intrusive Suite, these changes take place without significant changes in the total amount of alkali feldspar in the rocks.

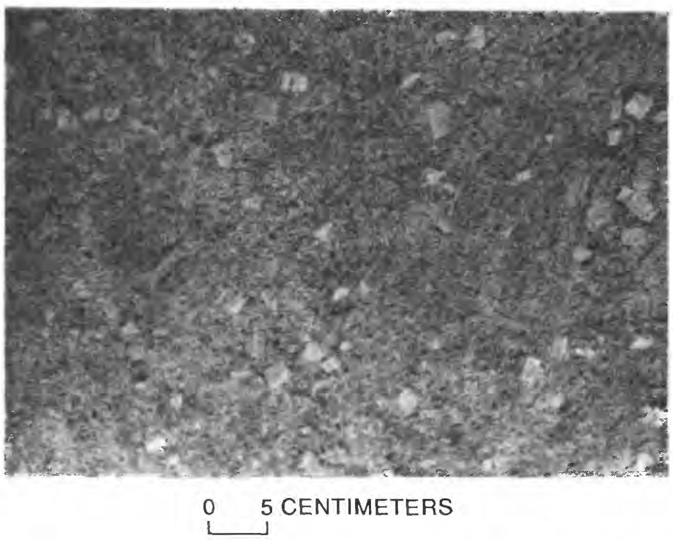


FIGURE 7.—Typical distribution of megacrysts in outer part of the Cathedral Peak Granodiorite.

A plot of all the modes of megacryst-bearing granites and granodiorites within the map area (pl. 1) is shown in figure 9*A*. Much of the scatter probably reflects analytical inaccuracies caused by the coarse and uneven grain size of the rocks. The most accurate modal analyses of megacrystic rocks within the map area, those made in the field by Bateman and Chappell (1979) on smooth glaciated outcrops of the Cathedral Peak Granodiorite, show much less scatter and plot mostly within the 10-percent contour for all the modal determinations (fig. 9*B*).

Compared to groundmass alkali feldspar, megacrysts typically contain more barium, and the cores of megacrysts are more barium enriched than the rims (Mehnart and Büsch, 1981). Kerrick (1969) interpreted the high barium content of the megacrysts of the Cathedral Peak Granodiorite to indicate that they formed in a higher temperature range and earlier than the groundmass alkali feldspar.

Both magmatic and metasomatic origins have been proposed for alkali feldspar megacrysts, but most, and presumably all, of those in the Sierra Nevada batholith are magmatic. The evidence that they are magmatic is that they have been carried in dikes, are oriented parallel to the magmatic foliations, have been flow sorted by size in schlieren, vary systematically in size and abundance relative to their position in plutons or intrusive suites, occur preferentially in rocks whose compositions plot close to the boundary between the granite and granodiorite fields on a Q-A-P diagram, and were formed at a higher temperature than the groundmass alkali feldspar. Several temperature-dependent processes may be involved—the rate of nucleation of alkali feldspar relative to other minerals, the rate of crystal growth of alkali feldspar, the length of time the alkali feldspar crystals had to grow in the magma before they were sealed in solid rock, and the abundance of crystals already present in the magma.

Swanson (1977) showed experimentally that in magma of granodiorite composition, the growth rate of alkali feldspar reaches a maximum at a temperature only slightly below the temperature at the beginning of nucleation, whereas the nucleation rate reaches a maximum at a lower temperature. Bateman and Chappell (1979) observed that in the

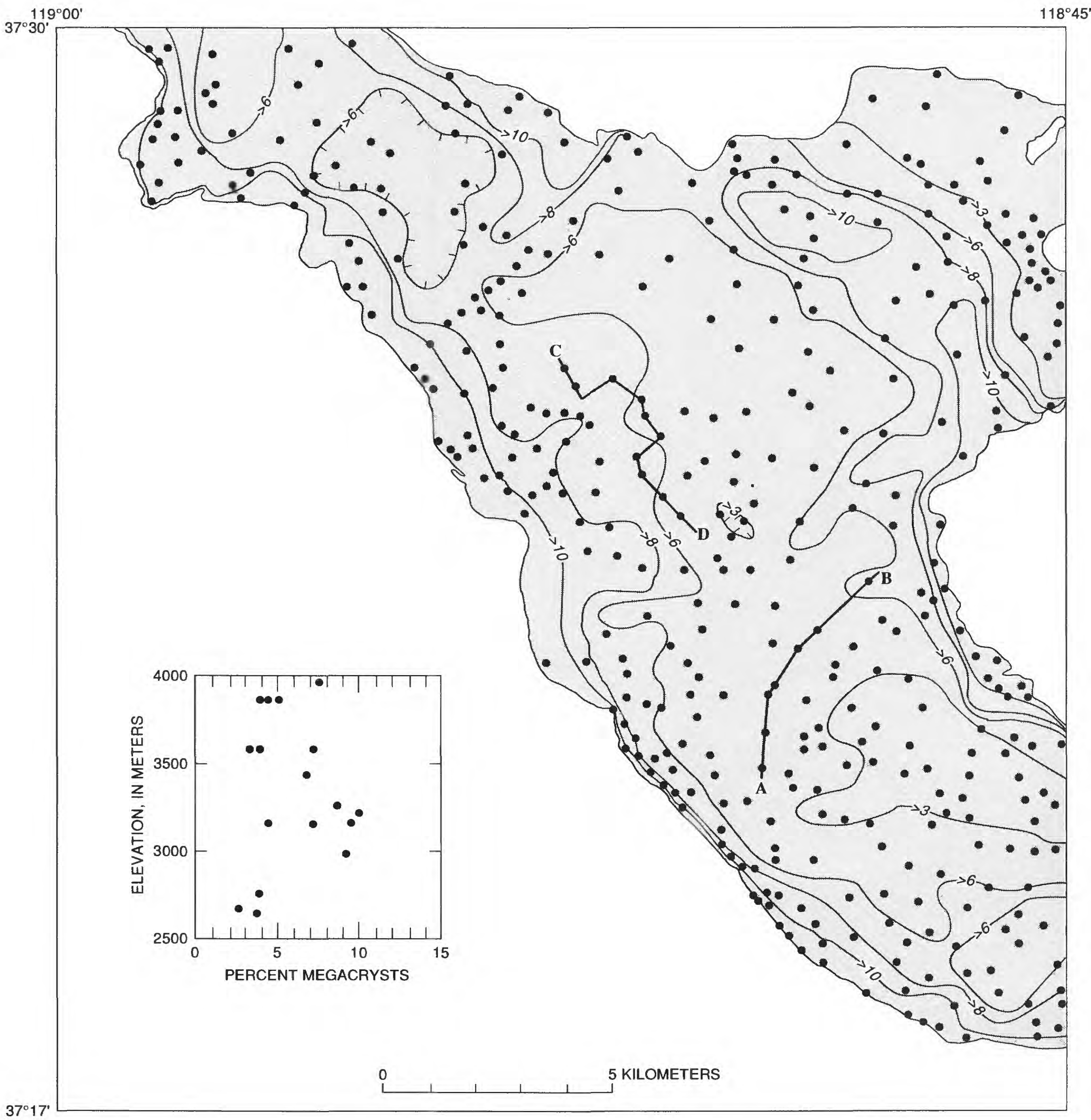


FIGURE 8.—Abundance of alkali feldspar megacrysts in central and southern parts of the Mono Creek Granite (shaded). Dots show sample localities. Isopleths show percent megacrysts in rock; hachures point toward lower values. Insert shows abundance of alkali feldspar megacrysts relative to altitude along traverses A–B and C–D. Modified from Lockwood (1975).

Tuolumne Intrusive Suite equant grains of alkali feldspar first appear in the interval between precipitation of plagioclase of An_{32} and An_{29} compositions and that compositions of plagioclase in the most calcic megacrystic rocks range from An_{39} to An_{23} and in the least calcic megacrystic rocks from An_{23} to An_9. Thus, An_{23} is the only plagioclase composition common to all the rocks that contain megacrysts and may be the composition of the plagioclase that was crystallizing when the growth rate for alkali feldspar was at the maximum. Thus, the growth rate of alkali feldspar in the Tuolumne magma may have increased as the An content of coexisting plagioclase decreased from An_{30} to An_{23}, then decreased as the composition of the plagioclase fell below An_{23}. Microprobe determination of the compositions of plagioclase grains enclosed in megacrysts would provide data needed to evaluate this speculation.

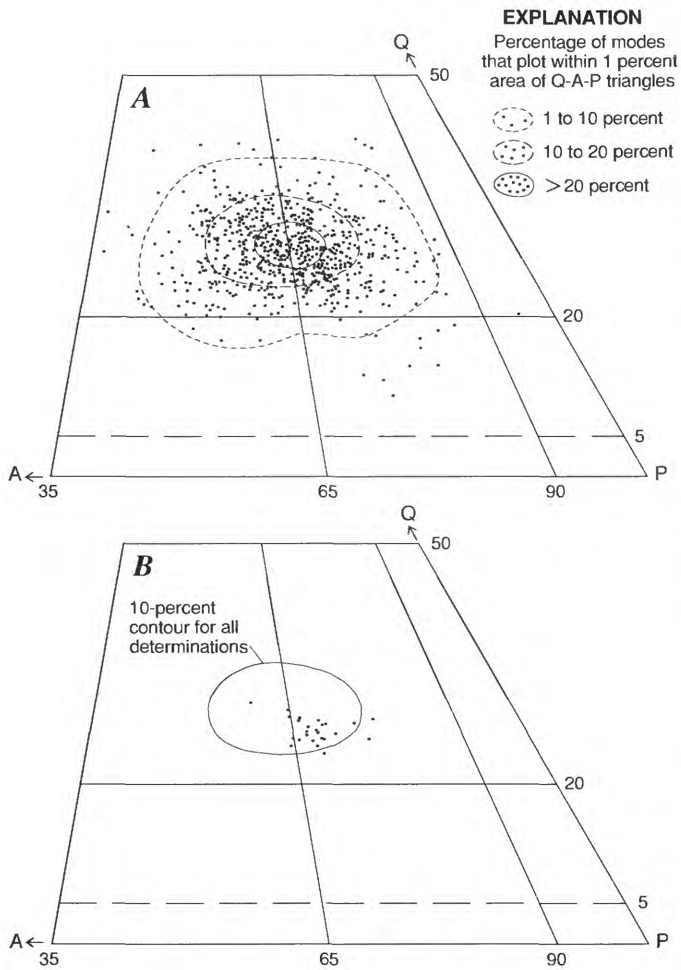


FIGURE 9.—Modes of plutonic rocks (dots) containing alkali feldspar megacrysts. *A*, All rocks within map area (pl. 1), including the Mono Creek and Cottonwood Granites, the Cathedral Peak and Wheeler Crest Granodiorites, granodiorite of McKinley Grove, megacrystic facies of the Mount Givens, Half Dome, and Dinkey Creek Granodiorites, and megacrystic facies of granodiorite of Eagle Peak. *B*, Most-accurate modes determined in field by Bateman and Chappell (1979) on glacially polished outcrops of the Cathedral Peak Granodiorite. See figure 6 for Q-A-P classification diagram.

The abundance of crystals present in the magma probably also is critical to the size and abundance of megacrysts because abundant crystals would interfere with the growth of megacrysts. Mafic minerals and plagioclase probably were abundant in the outer part of the Tuolumne Intrusive Suite magma when alkali feldspar first began to crystallize, but if the margin rock was formed by sidewall accretion of crystals, the adjacent magma may have been relatively free of crystals. During the later stages of solidification, the magma in the inner parts of the intrusion must have been increasingly crowded with crystalline material. An increase in the abundance of crystalline material rather than a decrease in the rate of crystal growth may be the principal cause of the decreasing size and abundance of megacrysts in the more felsic parts of intrusions.

PLAGIOCLASE

Plagioclase occurs as white to light-gray grains that are generally smaller than those of either quartz or alkali feldspar. Most grains are twinned on the albite law, and some are also twinned on the pericline or Carlsbad laws. The main part of almost all plagioclase grains is compositionally zoned, with the anorthite content being highest in the interior and decreasing outward. Progressive zoning from core to margin is commonly interrupted by discontinuities and (or) reversals. Thin compositional oscillations are common. Zoning reflects changes in the temperature, pressure, composition, or dissolved-volatile content of the melt phase during crystallization. The anorthite content of plagioclase commonly varies by 10 to 20 percent from the inner to the outer zone, but the range generally is smaller in granite than in granodiorite or tonalite. In granodiorite, plagioclase commonly ranges in composition from An_{45} to An_{30} in the cores of grains to An_{25} to An_{22} in their margins, whereas in granite plagioclase compositions range from An_{20} in the cores of grains to about An_{10} in their margins. Plagioclase that contains less than 10 percent anorthite may, at least in part, represent material that was exsolved from alkali feldspar.

Many plagioclase grains in granodiorite, tonalite, and quartz diorite contain mottled and altered cores that are discontinuous with the surrounding zoned plagioclase. The cores are generally distinctly more calcic than the most calcic zones in the surrounding zoned plagioclase, and a few exhibit reverse zoning. Although these cores have been interpreted to be restite, remnants of refractory material that survived melting during generation of the magma (Presnall and Bateman, 1973; Bateman and Chappell, 1979), other origins are more likely. The cores could have crystallized within the granitoid magma at depths where pressures were high, and very calcic cores could represent relic crystals that were present in mantle-derived basalt at the time the parent magmas were formed.

BIOTITE AND HORNBLENDE

Biotite and hornblende occur both as discrete, well-formed crystals and as anhedral to subhedral grains in clusters with opaque minerals, titanite, apatite, zircon, epidote, and chlorite. Some rocks, such as the Half Dome, Lamarck, and Round Valley Peak Granodiorites, are characterized by discrete hornblende prisms and biotite books, whereas others, most notably the Dinkey Creek Granodiorite, are characterized by clusters of anhedral to subhedral minerals and only scattered discrete crystals. The clusters range in size from tiny, vaguely defined clots of a few minerals to mafic inclusions as much as several meters in outcrop length. The well-formed discrete crystals appear to have precipitated from the melt phase of the magma, whereas the inclusions and clots probably were present in the magma since it was generated. The presence of euhedral hornblende crystals bordering some mafic inclusions and in trains terminating at mafic inclusions suggests that the mafic inclusions supplied material for the formation of the euhedral crystals.

The ratio of hornblende to biotite changes as their combined amount increases. Rocks whose color index is low generally contain little or no hornblende, but as the color index increases to values greater than some minimum range (5–20 for the Bass Lake Tonalite and 1–10 for the Tuolumne Intrusive Suite), hornblende appears and increases, generally at a faster rate than biotite (figs. 64, 65, 74, and 75).

Dodge and others (1968, 1969) examined biotite and hornblende from samples that were collected within and just south of the map area (pl. 1). They concluded that the composition of these minerals is similar in all the granitoids.

In studying the variations within the compositionally zoned bulge at the north end of the Mount Givens Granodiorite, Nokleberg did microprobe studies of both biotite and hornblende (fig. 10; Bateman and Nokleberg, 1978). The biotite shows little variation in composition except for an increase in the amount of MnO from the more mafic margin to the more felsic core of the bulge. However, the hornblende shows compositional zoning both within individual grains and from the margin to the core of the bulge. MgO decreases from the core to the rim of individual hornblende crystals as total Fe (as FeO) increases, and both the maximum and minimum amounts of MgO in hornblende decrease inward from the margins of the intrusion with an increase of Fe (as FeO). The TiO_2 and MnO concentrations in hornblende also increase slightly inward from the margins of the pluton. The behavior of MgO in hornblende is similar to that of CaO in plagioclase and doubtless reflects falling temperatures during solidification of the pluton from its margins inward. It also shows that hornblende continued to precipitate from the melt after the earliest exposed rocks began to solidify and that it is not entirely residual solid material from an earlier cycle.

In many granodiorites and tonalites, such as the Dinkey Creek and Tinemaha Granodiorites, hornblende crystals are partly or entirely replaced by tiny grains of biotite and plagioclase. This alteration requires the addition of potassium, which may have been concentrated in late interstitial melt or in volatiles. The irregular distribution of alkali feldspar as streaks and patches in some rocks clearly shows the mobility of potassium in the late stages of crystallization.

According to Burnham (1979), hornblende forms in andesitic magma from earlier crystallized anhydrous phases (such as orthopyroxene, clinopyroxene, plagioclase, magnetite, and ilmenite) when the temperature of the magma falls below approximately 920 °C and the water content is at least 3 percent. This seems to be a viable explanation for the aggregates of minerals in mafic inclusions and the small clots of anhedral minerals, but discrete euhedral hornblende crystals probably precipitated directly from the melt phase. Hornblende grains that have pyroxene cores show no evidence of alteration of the pyroxene to form hornblende. Burnham (1979) stated that although a 3-percent water content would allow hornblende to coexist stably with the melt, 4 percent is required for the formation of euhedral, or even subhedral, crystals. These water concentrations generally agree with the results of experiments conducted by Naney

(1977), who also determined that at 2-kbar water pressure hornblende forms at temperatures less than about 850 °C. Most granitic magmas probably began to crystallize hornblende only after precipitation of anhydrous minerals had increased the volatile content of the melt phase to some critical amount and the temperature had fallen to about 850 °C. Small amounts of fluorine would signifi-

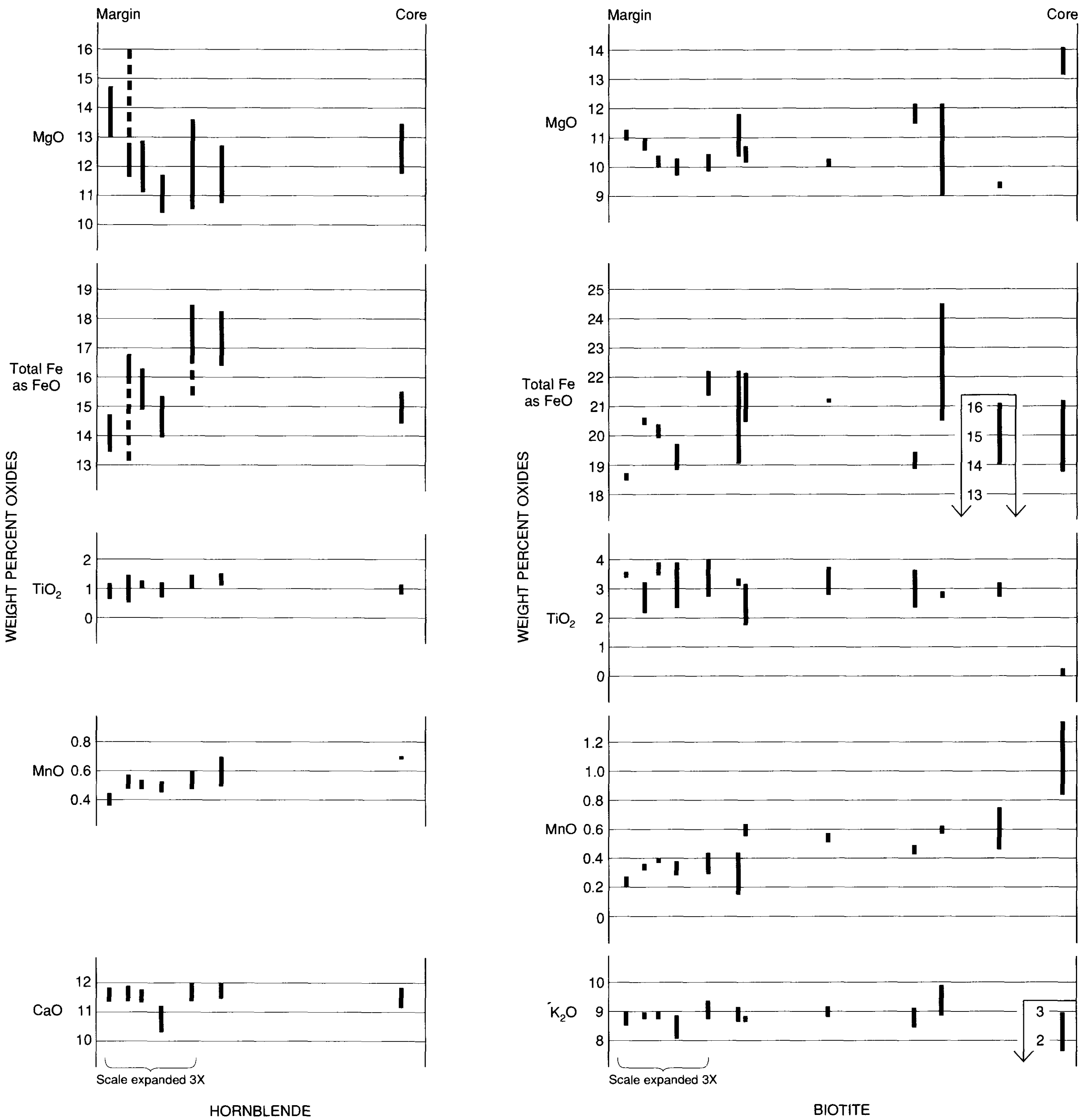


FIGURE 10.—Composition of hornblende and biotite in samples collected from margin to core of compositionally zoned bulbous head at north end of the Mount Givens Granodiorite. Dashes indicate range in one or two determinations for several samples. From Bateman and Nokleberg (1978).

cantly lower the amount of water required and raise the temperature at which hornblende first appears.

ACCESSORY MINERALS

The common accessory minerals are magnetite, ilmenite, titanite, apatite, and zircon. Thorite, monazite, and allanite have also been identified but are less common. Magnetite is present in most rocks but is scarce in the tonalites of the western foothills, probably chiefly because of low f_{O_2} in the magma (Dodge, 1972). Magnetite and ilmenite occur as small equant grains. These two minerals have not been separately identified in most rocks, but microprobe studies show that ilmenite is present in the Mount Givens Granodiorite and absent in the Tuolumne Intrusive Suite, which contains abundant titanite. Czamanske and Mahalik (1972) and Czamanske and Wones (1973) suggested several reactions in which an increase in f_{O_2} can cause an increase in the ratio of magnetite to ilmenite. Titanite that occurs as discrete crystals appears to be primary, but that occurring as irregular grains associated with an opaque mineral, presumably magnetite, probably exsolved from preexisting ilmenite. Apatite forms euhedral prisms that are stubby in most rocks but acicular in others. Many zircon crystals in biotite have pleochroic halos. Generally allanite is uncommon but is conspicuous in some rocks of the eastern Sierra Nevada. Where present, it occurs as cigar-shaped grains that are variable in color and pleochroism; colors range from reddish brown through grayish orange to yellowish orange.

TEXTURES

TEXTURES IN UNDEFORMED GRANITOIDS

Most undeformed granitoids are medium grained and have hypidiomorphic granular textures, but a few small intrusions of felsic rock (for example, the Johnson Granite Porphyry, the granite of Hogan Mountain, the quartz syenite of the Goddard septum, the alaskite of Evolution Basin, the leucogranite of Taboose Creek, and parts of the granite of Pellisier Flats and Ward Mountain Trondhjemite) are fine grained or composed of anhedral grains of variable size. The magmas for the predominant hypidiomorphic granular rocks were undersaturated with water throughout their periods of crystallization, whereas those composed of fine grains or of anhedral grains of variable size crystallized from magmas that suffered loss of pressure and consequent quenching, volatile fluxing, and alkali metasomatism.

In hypidiomorphic-granular textured rocks the felsic minerals form the largest crystals, hornblende and biotite, the next largest, and the accessory minerals the smallest. Hornblende and biotite are euhedral in some rocks, such as the Half Dome Granodiorite, but form clots of anhedral to subhedral grains in other rocks, such as the Dinkey Creek Granodiorite. Euhedral hornblende prisms generally are 5 to 10 mm long, but some are as long as 2 cm; euhedral biotite books are generally 2 to 5 cm across, but in some rocks they are as much as 1 cm across. Miarolitic cavities are uncommon but are present in a few fine-grained felsic rocks.

TEXTURES IN DEFORMED GRANITOIDS

Gneissic foliation, commonly accompanied by lineation, occurs in granitoids that were strained when they were largely or completely crystallized. In some rocks the strain is the result of continued intrusion during late stages of crystallization, and in others it is the result of regional stress. The minerals respond in a variety of ways during ductile deformation. Quartz is the mineral most sensitive to strain and readily breaks down to form a microcrystalline mortar between the other minerals in the rock, especially the feldspars. Biotite and hornblende are almost as responsive as quartz to strain and are reduced to elongate strands, shreds, and tiny isolated fragments. Plagioclase is relatively resistant, and in moderately deformed rocks porphyroclasts commonly survive, although fracturing and rotation may have produced optical discontinuities visible in thin section. Under moderate strain, interstitial alkali feldspar appears to break down with plagioclase, forming granoblastic mortar with quartz and plagioclase. However, equant grains and megacrysts are resistant to deformation and can appear relatively undeformed in a granoblastic groundmass, often giving the erroneous impression that they grew as porphyroblasts. In moderately and strongly strained rocks, the long diagonal dimension of megacrysts lies in the plane of foliation. Some severely strained rocks are composed of thin, needlelike aggregates of quartz and feldspar, which are commonly sheathed in biotite. In these rocks, much of the hornblende is converted to biotite. Such a reaction would necessarily involve the feldspars and other minerals. Mehnert (1968) suggested the following reaction:

$$\text{Biotite+plagioclase}(An_{45})\text{+quartz=hornblende}$$
$$\text{+titanite+plagioclase}(An_{35})\text{+orthoclase.}$$

FOLIATION AND LINEATION

PRIMARY MAGMATIC FOLIATION AND LINEATION

Most bodies of quartz diorite, tonalite, granodiorite, quartz monzodiorite, and quartz monzonite and some bodies of granite are foliated. Most of the foliation that has been observed in the Sierra Nevada is primary and formed as a result of the differential flow velocities in the magma as it solidified. In some places, primary foliation has been deflected by forcefully emplaced younger intrusions (fig. 11).

Primary foliation is shown both by the preferred orientation of tabular and prismatic minerals and by flattened mafic inclusions shaped by magmatic flow. Mineral foliation in the host generally parallels foliation shown by mafic inclusions but in places is less regular and varies nonsystematically a few degrees from the foliation shown by mafic inclusions. Lineation generally is difficult to identify in rocks that have not undergone ductile deformation, but oriented hornblende prisms define a weak lineation in the plane of mineral foliation in some granitoids. Markedly elongate mafic inclusions have not been observed in undeformed granitoids. Inclusions that appear elongate in outcrop generally are lens-shaped or faintly triaxial. However, Kistler and Swanson (1981) described spindle-shaped mafic clots in a hypabyssal intrusion. Where identifiable, primary lineation generally plunges down the dip of the foliation.

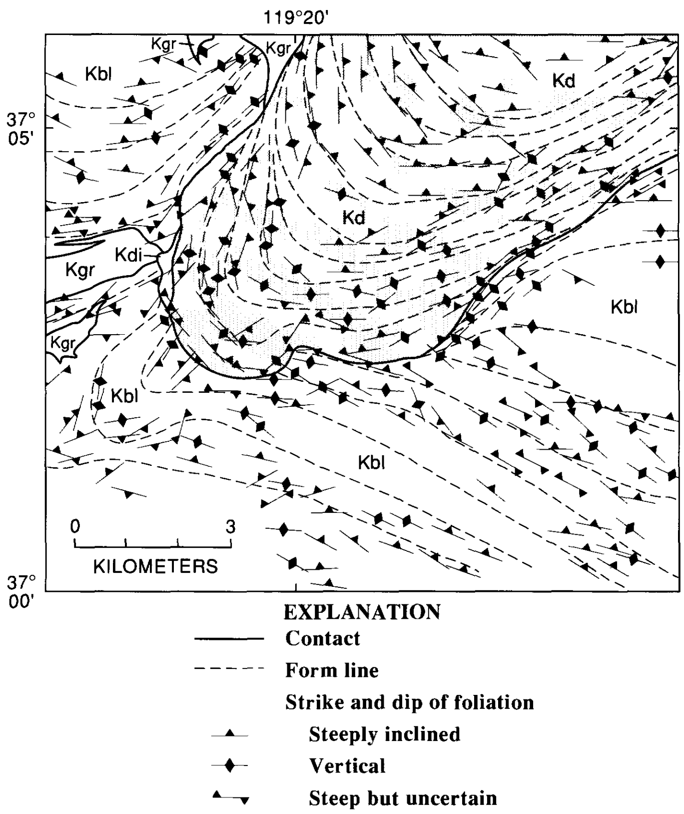


FIGURE 11.—Original and deformed foliation patterns (shown by dashed form lines) south of Shaver Lake. Arcuate pattern of foliation in the Dinkey Creek Granodiorite (Kd) is original, whereas foliation pattern in the older Bass Lake Tonalite (Kbl) was disturbed during emplacement of the Dinkey Creek Granodiorite. Kgr, younger granite; Kdi, older diorite.

In general, foliation approximately parallels and is strongest close to external intrusive contacts with older rocks. With distance from the contact, the foliation is progressively weaker because the preferred orientation of minerals is less perfect and because mafic inclusions are less flattened. In compositionally zoned intrusions, the abundance of mafic inclusions decreases as the rock becomes more leucocratic. Inward, the foliation may diverge from the contact, but generally it does so in broad, sweeping curves that are subparallel to the margins of the intrusion (fig. 11).

Balk (1937) summarized and extended the pioneering work of Cloos (1923) and his school of geologists in Europe who explained the foliation in igneous rocks in terms of flow movements in the magma during solidification. Later Mackin (1947) identified three types of nonuniform flow that can produce foliation (fig. 12); he termed these three types of flow "deceleration flow," "acceleration flow," and "velocity gradient flow." Velocity changes occur in the direction of flow in deceleration flow, across the direction of flow in velocity gradient flow, and both across and in the direction of flow in acceleration flow.

Deceleration flow occurs where a body of magma balloons outward from a central feeder. In this type of flow, an influx of magma pushes the chamber walls outward, the velocity of flow diminishing with distance from the feeder. Because trailing edges and ends of crystals and mafic inclusions move faster than leading ends or edges, inclusions are flattened and crystals are rotated toward parallelism with the chamber walls. This type of flow produces a strong foliation but generally only a weak lineation or no lineation. It explains almost all the observed magmatic foliation in the granitoids and provides strong evidence that the magmas ballooned as they were emplaced.

Acceleration flow occurs when magma moves through an orifice of diminished size, a process requiring increased velocity of flow. The increased rate of flow in the direction of flow stretches mafic inclusions and orients the long axes of prismatic crystals in the direction of flow because the leading ends move faster than the trailing ends. This type of flow might take place in the narrowing ends of elongate or dikelike intrusions. However, it should produce a magmatic lineation, which has not been observed in the Sierra Nevada.

In velocity gradient flow, drag along an interface causes the rate of flow parallel to the interface to decrease toward the interface. Initially, the interface is a sharp contact of crystal-laden magma with adjacent rock, but with inward crystallization and solidification a transitional zone between more fluid and near-solid magma moves away from and effectively replaces the original solid-liquid interface. The increased rate of flow with distance from the contact rotates tabular and prismatic crystals and stretches soft mafic inclusions parallel to the contact (fig. 12). This type of flow, like acceleration flow, should produce a lineation, but lineation has not been observed in undeformed granitoids.

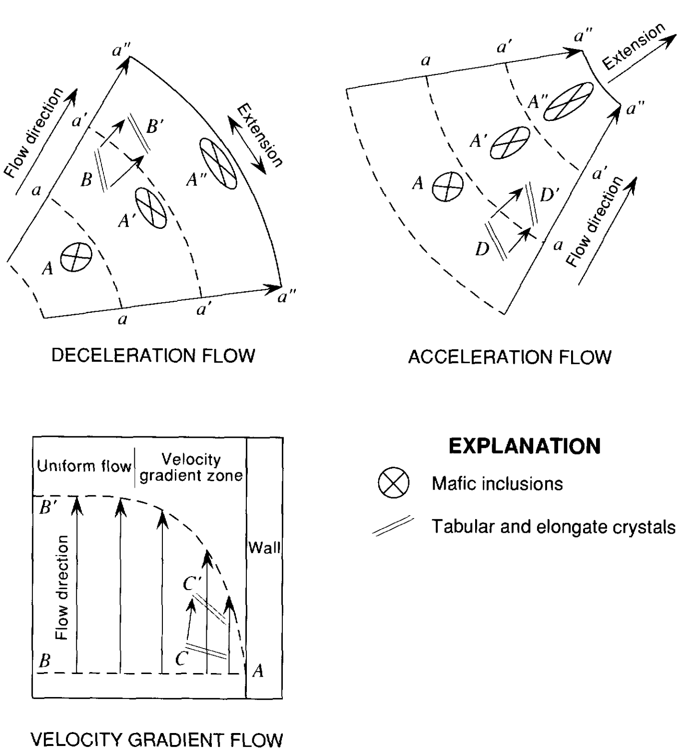


FIGURE 12.—Three types of nonuniform flow that can produce foliation and lineation in plutonic rocks. Diagrams represent either horizontal or vertical sections. Modified from Mackin (1947). See text for discussion.

DISTURBED PRIMARY FOLIATION

In many places, primary foliation bends around younger intrusions. Such patterns are predictable where the intrusions belong to the same intrusive suite and the older intrusion may not have been completely crystallized when the younger intrusion was emplaced. However, where the foliation in an older intrusion is deflected around a much younger intrusion, the foliation in the older intrusion must have been disturbed after the rock was completely crystallized. Original northwest-trending foliation in the Bass Lake Tonalite (fig. 11) is bent toward the southwest around a tongue of Dinkey Creek Granodiorite that intruded it from the northeast. The isotopic ages of these two intrusions differ by more than 10 m.y., so it is unlikely that the Bass Lake Tonalite was still fluid when the Dinkey Creek Granodiorite was emplaced. Nevertheless, thin sections of samples collected from the Bass Lake Tonalite within the area of deflected foliation show little evidence of post-consolidation strain. Presumably ambient temperatures in the Bass Lake Tonalite were still high enough when the Dinkey Creek Granodiorite intruded it for the granodiorite magma to raise temperatures in the tonalite to above its solidus.

DUCTILE FOLIATION AND LINEATION

Ductile foliation and lineation in the granitic rocks have two causes: (1) regional stress and (2) stresses caused by the emplacement of intrusions.

DEFORMATION ATTRIBUTED TO REGIONAL STRESS

Locally, the granitic rocks are cut by zones of ductile shearing that occurred after they solidified, but evidence of deformation caused by regional stress as intrusions cooled and crystallized is sparse, and only a few examples can be cited. Sylvester, Oertal, and others (1978) interpret a conjugate system of short strike-slip faults that cross the Papoose Flat pluton in the northern Inyo Mountains to be the result of regional stresses that were overshadowed during forcible emplacement of the pluton but renewed their dominance soon thereafter. The faults displace the margins of the pluton but can be traced only short distances into its interior. These relations indicate that the faulting took place after the border facies of the pluton was cool enough to fracture but while the interior was still hot enough to seal and obscure the fractures. When the offsets along the faults are removed, the original shape of the pluton in plan view can be seen to have been elliptical and elongate to the northwest, parallel with other plutons in the region.

The Tuolumne Intrusive Suite shows two foliation patterns: a weak and discontinuous primary magmatic foliation that more or less parallels the external and internal contacts of the suite and a steeply dipping foliation that strikes about N. 70° W. across the suite. The second foliation is shown chiefly by the preferred orientation of small biotite flakes. The yellow-brown color of the biotite in thin section and the absence of evidence of ductile deformation suggest that the second foliation was produced at relatively high—magmatic or near-magmatic—temperatures, probably during the waning stages of crystallization, and reflects regional stress. Cray (1981) reported a somewhat-similar second west-northwest foliation in the Morgan Creek pluton of the Lake Edison Granodiorite and in adjacent older rocks but not in younger rocks. However, the foliation in the Morgan Creek pluton is also shown by closely spaced microfractures (Cray, 1981).

The tonalite of Granite Creek in the western foothills was pervasively deformed during regional deformation, apparently long after it was emplaced. Northwest-trending ductile foliation and local lineation in the tonalite is continuous with foliation and lineation in the adjacent metamorphic rocks and obviously is the result of regional deformation, presumably during the Nevadan orogeny. A concordant U-Pb age of 163 Ma and a slightly discordant age of 166 Ma show that the intrusion was emplaced about 10 m.y. before the Nevadan orogeny.

Shear Zones in the Eastern Sierra Nevada

Several northwest-trending shear zones, 500 to 1,000 m wide and as much as 15 km long, occur in the higher parts of the Sierra Nevada within rocks peripheral to the Mount Givens Granodiorite, and other related shear zones could have been present within the area now occupied by the Mount Givens Granodiorite. One zone lies along the southwest side of the Mount Givens Granodiorite, three lie north of the bulbous head at the north end of the Mount Givens, and four are east of the Mount Givens. Most of the shear zones originated during the early Late Cretaceous but at somewhat-different times. They cut rocks as young as about 97 Ma and are truncated by rocks as old as 90 Ma.

An extensive zone of shearing within the Goddard septum is present in both metamorphic and plutonic rocks. Mafic dikes thought to be part of the ~148-Ma Independence dike swarm (Chen and Moore, 1979) intrude the granitic rocks and are themselves sheared. However, more study is needed to determine whether the mafic dikes were emplaced before any shearing occurred or were emplaced after an earlier period of shearing and were themselves sheared during a later period. Another less extensive zone of gneissic granite that may have been deformed during the Jurassic lies along the South Fork of Bishop Creek. This granite is intruded by the Inconsolable Quartz Monzodiorite, which has been dated by the Rb-Sr method at 105±11 Ma, and by the Late Cretaceous Lamarck Granodiorite.

The best-exposed and most easily accessible shear zone extends southeast from Courtright Reservoir to the Cape Horn septum (Bateman, Kistler, and DeGraff, 1984). There, the Dinkey Creek Granodiorite on the west is in contact with the Mount Givens Granodiorite on the east except where a thin discontinuous septum of metamorphic rock separates them. Shearing in the margin of the Dinkey Creek Granodiorite continues into the granites of Short Hair Creek and Lost Peak, small plutons that intrude the Dinkey Creek Granodiorite, but not into the Mount Givens Granodiorite.

As the contact between the Dinkey Creek and Mount Givens Granodiorites is approached from the west just south of Courtright Reservoir (fig. 13), evidence of deformation in the Dinkey Creek Granodiorite increases. The grain size is gradually reduced, and conspicuous steeply dipping foliation and downdip lineation become increasingly apparent. The foliation and lineation are shown by slivers and needles of mineral aggregates sheathed in biotite and by abundant and evenly distributed flattened and elongate mafic inclusions. Extremely fine grained mylonitic zones are present close to the contact. Measurements made close to the contact with the Mount Givens Granodiorite on joint surfaces both parallel and perpendicular to the foliation show that the mafic inclusions have been drawn into blades with approximate axial ratios of 1:7:30 (fig. 14).

Various felsic dikes intrude the deformed Dinkey Creek Granodiorite where a septum separates it from the Mount Givens Granodiorite. Most of the fine-grained dikes are deformed and lineated, whereas the coarser grained dikes, including pegmatites, are not. The lineated fine-grained dikes are offshoots of the granite of Lost Peak, and the

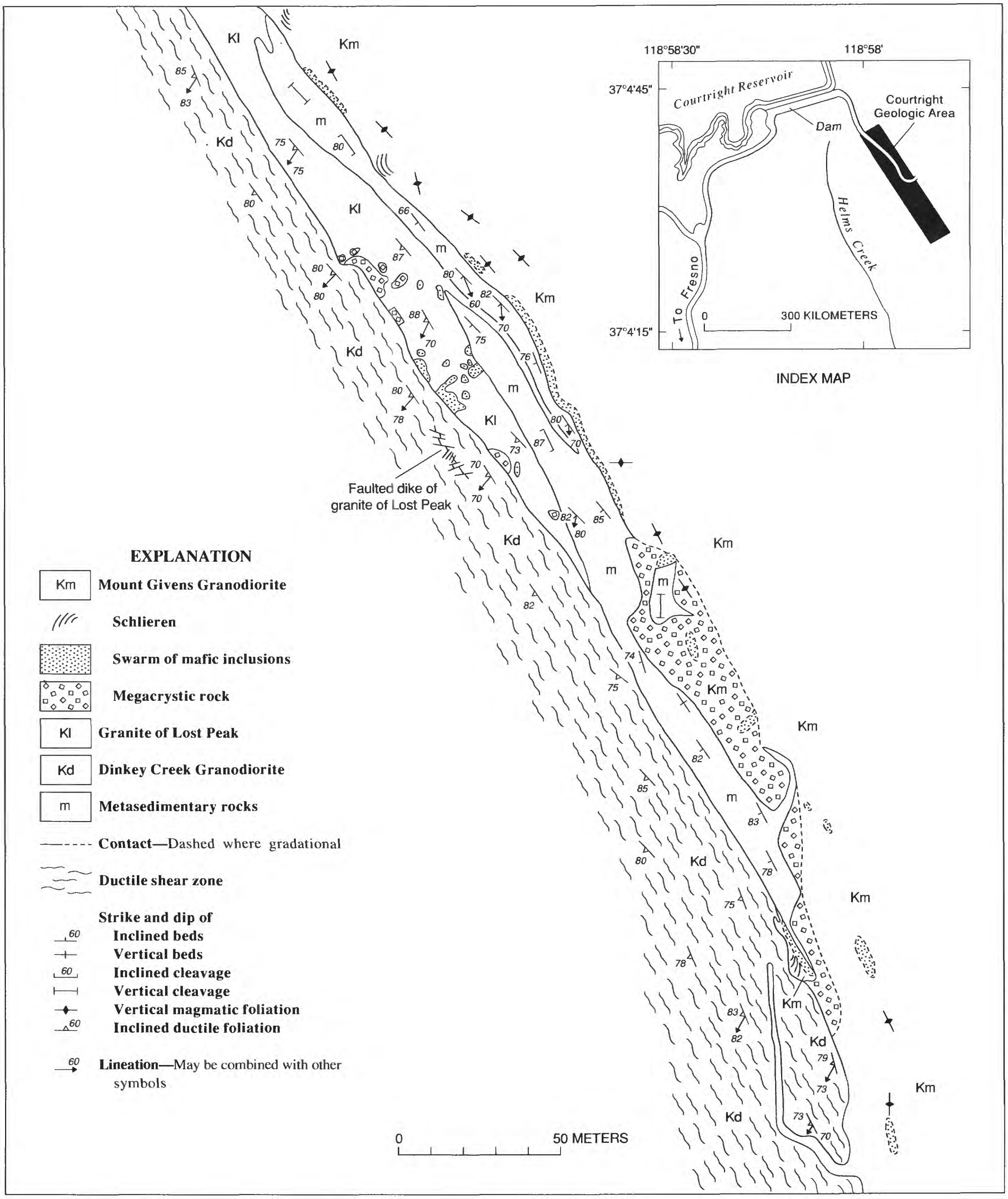


FIGURE 13.—Geology of Courtright Geologic Area at south end of Courtright Reservoir.

undeformed coarser grained dikes are offshoots of the Mount Givens Granodiorite.

Inclusions of sheared Dinkey Creek in the undeformed Mount Givens and undeformed dikes of Mount Givens in sheared Dinkey Creek show clearly that the shearing occurred before emplacement of the Mount Givens Granodiorite was completed and suggest that it represents faulting along a preexisting contact of the Dinkey Creek Granodiorite with metamorphic rocks before the Mount Givens Granodiorite was emplaced. However, planar foliation and an absence of lineation in the adjacent Mount Givens Granodiorite allow the possibility that ballooning of the Mount Givens magma caused the ductile deformation and that ballooning ceased before the Mount Givens magma crystallized sufficiently to shear itself. Continuation of the ductile structures in the Dinkey Creek Granodiorite into the granite of Lost Peak (fig. 13) and the nearby granite of Short Hair Creek shows that the deformation occurred after the Dinkey Creek Granodiorite was emplaced and had solidified and that it was not caused by the intrusion of the Dinkey Creek. The downdip elongation of inclusions and striae on the youngest shear surfaces indicate that the movement was dip-slip. However, the sense of movement—whether the southwest side moved relatively upward or downward—has not been determined.

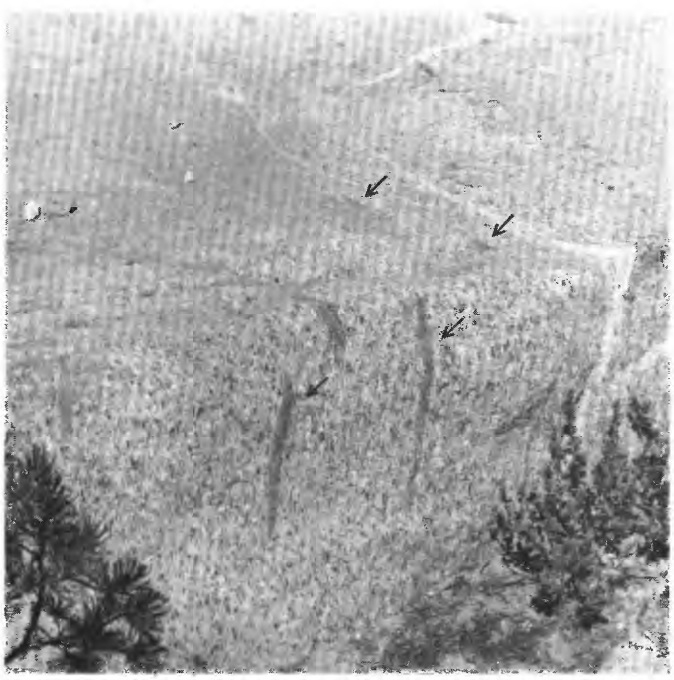


FIGURE 14.—Dinkey Creek Granodiorite showing mafic inclusions (arrows) drawn into bladelike forms as a result of ductile deformation. Courtright Geologic Area.

DEFORMATION ASSOCIATED WITH EMPLACEMENT OF INTRUSIONS

Some intrusions show no evidence of ductile deformation within themselves, whereas others show evidence of ductile deformation that is increasingly intense toward their margins. Intrusions that show no internal evidence of ductile deformation apparently stopped expanding before they had crystallized sufficiently to shear themselves, whereas those that show evidence of ductile deformation continued to expand after crystallization in the deformed parts was well advanced. Intrusions such as the Mount Givens Granodiorite and the Tuolumne Intrusive Suite, which show little or no internal evidence of ductile deformation, have nevertheless displaced their wallrocks; this forcible emplacement probably accounts for at least some of the cleavages and lineations within them. Apparently a magma too fluid to shear itself can nevertheless produce cleavages, lineations, and shears in adjacent solid rock. The structural effects of intrusion are readily apparent in and adjacent to rounded intrusions that obviously ballooned as they were emplaced but are much less obvious adjacent to elongate intrusions. Nevertheless, Tobisch and others (1986) suggested that cleavage and lineation in the Goddard septum, which is bordered on both sides by elongate intrusions, were caused by the emplacement of the intrusions.

Intrusions that apparently continued to move during late stages of crystallization and that have granoblastic fabrics include the megacrystic granites of Birch Creek and of Papoose Flat in the northern Inyo and White Mountains (Nelson and Sylvester, 1971; Nelson and others, 1978; Sylvester, Oertal, and others, 1978) and the Ward Mountain Trondhjemite of the Fine Gold Intrusive Suite. All these intrusions are composed of leucocratic rocks that crystallized in relatively low temperature ranges from viscous magma. Outward-dipping external contacts of all these intrusive bodies indicate that it is their upper parts that are exposed. Exposures of these intrusions may be localized in the margins of the batholith, where less rock has been removed by erosion and ambient temperatures were lower than in the central part of the batholith, where contacts are uniformly steep, erosion probably deeper, and ambient temperatures higher. The structural effects in the wallrocks of these intrusions are more intense than in the wallrocks of intrusions that were hotter and more fluid when they were emplaced. These wallrock effects, typified by conspicuous lineations, are described in more detail in the two following sections.

No studies have been made to determine whether the cleavage and lineation in country rocks bordering intrusions that were still fluid when they were emplaced were caused by the intrusions or are merely reoriented preexisting structures, but structures in older plutonic rocks marginal to such intrusions suggest that both may be present. Ballooning of the Tuolumne Intrusive Suite clearly caused ductile foliation and shearing in adjacent parts of the older granite of Ten Lakes and the Mount Hoffman pluton of the El Capitan Granite, whereas the tongue of Dinkey Creek Granodiorite that penetrates the Bass Lake Tonalite bowed foliations in the tonalite without producing ductile structures. Presumably the Dinkey Creek magma reheated the Bass Lake Tonalite above its solidus and allowed magmatic flow.

DEFORMATION ASSOCIATED WITH THE EMPLACEMENT OF THE WARD MOUNTAIN TRONDHJEMITE

A zone of deformation in the western foothills of the Sierra Nevada southwest of Oakhurst (called the Blackhawk deformation by Bateman, Busacca, and Sawka, 1983) is centered on the Ward Mountain Trondhjemite, which forms two adjacent plutons, one just south of the other (Bateman, Busacca, and Sawka, 1983). North, east, and south of these plutons for distances of 5 to 15 km, all the rocks older than the trondhjemite have been deformed and lineated (fig. 15). In contrast, the older rocks west of the plutons are largely undeformed except immediately adjacent to the Ward Mountain Trondhjemite. The deformed rocks include metasedimentary and metavolcanic strata, the Bass Lake Tonalite, several granite and granodiorite bodies, and the trondhjemite itself. The trondhjemite is largely gneissic, and foliation patterns are complex. Almost certainly the two plutons are continuous at depth and form a single body that is convex toward the east and much thickened in the middle. In cross section the trondhjemite is strongly asymmetric; the northeast, east, and southeast flanks of both plutons slope gently outward, whereas their west flanks are vertical or dip steeply eastward under the plutons.

The planar and linear structures in the rocks marginal to the trondhjemite are continuous across both the plutonic and stratified rocks and form coherent patterns (fig. 15). Planar structures generally strike parallel to contacts with the trondhjemite and dip outward, forming a concentric pattern around the northeast, east, and southeast sides of the plutons except between the plutons where foliations indicate a synform. Lineations generally plunge outward from the trondhjemite plutons and form a radial pattern.

The deformation is attributed to the emplacement of the trondhjemite at a time when crystallization was far advanced (Bateman, Busacca, and Sawka, 1983). Final movements were upward and westward. Heat and volatiles emanating from the trondhjemite probably increased the susceptibility of the rocks overlying the gently dipping east flanks of the trondhjemite to attenuate normal to the planar structures and to stretch in the direction of the linear structures. The boundaries of the deformed area are gradational and may represent a critical isotherm, possibly in the range of 300 to 400 °C, at which temperatures dislocation motion can occur in dry quartz but not in feldspar (Tullis and Yund, 1977).

Isotopic-age determinations show that all the deformed plutonic rocks were emplaced and deformed during a relatively brief interval. The isotopic age of the Bass Lake Tonalite is about 114 Ma, and that of the Ward Mountain Trondhjemite is about 115 Ma (Stern and others, 1981). The undeformed Knowles Granodiorite, which has an isotopic age of about 112 Ma, intrudes the deformed rocks and truncates the ductile structures in them.

DEFORMATION ASSOCIATED WITH INTRUSIONS IN THE WHITE AND NORTHERN INYO MOUNTAINS

Nelson and Sylvester (1971) described deformation associated with the emplacement of the granite of Birch Creek, and Nelson and others (1978) and Sylvester, Oertal, and others (1978) described in detail ductile deformation at the west end of the granite of Papoose Flat. Both plutons were emplaced in stratified rocks, so no older plutons were affected by their emplacement. However, the western margins of both plutons were deformed during intrusion, the grain size was reduced, and strong foliations and lineations developed. The granite of Papoose Flat is megacrystic, and relict alkali feldspar megacrysts are enclosed in a fine-grained foliated and lineated matrix composed chiefly of quartz, feldspar, and biotite. The megacrysts have been rotated to bring long diagonal axes into the plane of foliation, in contrast to the orientation of megacrysts in undeformed rocks where crystal faces commonly parallel magmatic-foliation planes. Survival of the megacrysts suggests that the temperature range at the time of deformation was below the range at which feldspar glides readily and close to the temperature on the solidus. Sylvester, Oertal, and others (1978) attributed the deformation to forcible emplacement of the deformed west end of the pluton when it was extremely viscous and almost completely crystalline. During

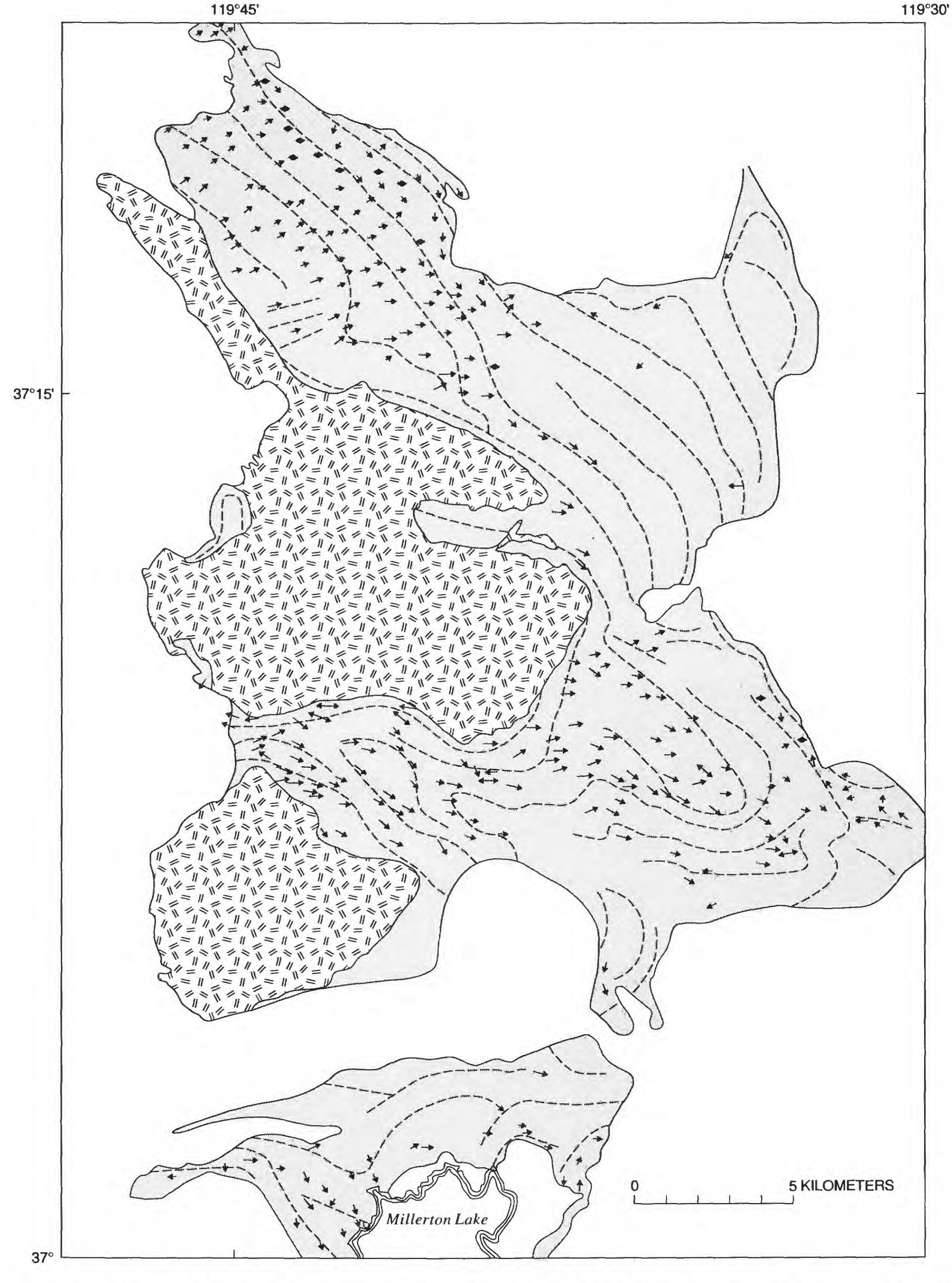


FIGURE 15.—Patterns of ductile foliation and lineation in the Bass Lake Tonalite and metamorphic and granitic rocks adjacent to the Ward Mountain Trondhjemite. Modified from Bateman, Busacca, and Sawka (1983).

emplacement, the principal axes of stress and of shortening strain were oriented perpendicular to the pluton walls and to layering in adjacent attenuated strata. The maximum elongation was in the direction of the long axis of the pluton and parallel to its upper surface, the plunge varying from place to place.

MAFIC INCLUSIONS

Dark, fine-grained inclusions with igneous textures are by far the most abundant extraneous material in the granitic rocks (fig. 16). They are distributed with great regularity in most granodiorites and tonalites but are sparse or absent in granitoids that contain little or no hornblende or in which almost all the hornblende and biotite crystals are euhedral. In reports on the Sierra Nevada published since 1960, these inclusions have been referred to as "mafic inclusions." Mafic inclusions have also been designated "basic concretions" (Grubenmann, 1896), "autoliths" (Holland, 1900; Pabst, 1928), "basic segregations" (Knopf and Thelen, 1905), and "microgranular enclaves" (Didier, 1973).

In compositionally zoned granitoid plutons, the abundance of mafic inclusions commonly varies with the mafic-mineral content of the host rock. For example, the number of mafic inclusions in the compositionally zoned granodiorite of Cartridge Pass decreases gradually from more than two per square meter in the margins, where the mafic-mineral content of the host rock is more than 15 percent, to 0 in the interior, where the mafic minerals constitute less than 6 percent of the host rock (Bateman and others, 1963).

The inclusions vary widely in size and shape but are generally of similar size, shape, and composition within any particular area. As seen in outcrops, they range in area from a few square centimeters to 1 m^2 or more. Doubly convex discoid shapes are most common, but irregularly rounded shapes prevail in the interiors of some plutons. Spindle shapes reported by Balk (1937) have been observed only in rocks that have undergone ductile deformation. Length-to-thickness ratios of the inclusions as seen in outcrops are greatest near external contacts of plutons and generally decrease with distance from the contact. Despite their apparent linearity in outcrops, the mafic inclusions in most outcrops are notably discoid; measurements on variously oriented joint surfaces do not indicate significant elongation. The boundaries of the inclusions with the enclosing granitoids appear sharp, but minerals interlock across these contacts.

The variably flattened mafic inclusions in different parts of intrusions indicate that their shapes have changed in accordance with movements in the enclosing granitic magma. An absence of evidence of ductile deformation implies that the inclusions were soft and contained interstitial melt when they attained their shapes; probably they were themselves magma, only a little less viscous than the surrounding granitic magma.

The mafic inclusions are composed of the same minerals as the enclosing granitoid but in different proportions. They contain much more hornblende, biotite, opaque minerals, and apatite, and much less quartz and alkali feldspar. Typically, the inclusions contain 35 to 60 percent plagioclase, 20 to 50 percent hornblende, 5 to 20 percent biotite, 0 to 15 percent quartz, little or no alkali feldspar, and

EXPLANATION

Ward Mountain Trondhjemite

Bass Lake Tonalite, metamorphic and granitic rocks, undivided

Boundary of ductilely deformed area

Form line showing trace of cleavage in stratified rocks and ductile foliation in granitoids

Lineation— Length of shaft of arrow is proportional to complement of angle of plunge

0°

45°

90°

FIGURE 15.—Continued

FIGURE 16.—Oriented double-convex discoid mafic inclusions in the Dinkey Creek Granodiorite. Southwest shore of Shaver Lake east of Shaver Lake Heights.

relatively abundant opaque minerals and acicular apatite. Studies by Bernard Barbarin (oral and written communs., 1985–1986) show that the Fe/Mg values for hornblende and biotite and the range of the anorthite content of plagioclase are the same in inclusions as in the enclosing granitic rock. The mineral compositions of the inclusions and the host rock change together from one place to another within intrusions and from one intrusion to another. The mafic inclusions also are in equilibrium with their enclosing host rock with respect to strontium isotopes; for example, $^{87}Sr/^{86}Sr$ values for mafic inclusions in the Bass Lake Tonalite (fig. 17) plot on the isochron for the enclosing Bass Lake Tonalite, whereas $^{87}Sr/^{86}Sr$ values for nearby mafic dikes and gabbroic intrusions plot well off the isochron (fig. 18).

The remarkably even distribution of the inclusions over wide areas indicates that they were in residence within the magma from early stages, but their source is still being debated. The compositional equilibrium between the same mineral species in the inclusions and the host granitoid and the equilibrium of the inclusions and the host granitoid with respect to $^{87}Sr/^{86}Sr$ mask the earlier history of the inclusions, which must account for their bulk compositions.

Some of the mafic minerals may have been derived from earlier crystallized anhydrous minerals that were present in the inclusions when they were first formed, as suggested by Burnham (1979). Likely early-crystallized minerals include pyroxene, calcic plagioclase, and possibly olivine. Sparse feldspar crystals and fragments of granitic rock present in some inclusions appear to have been incorporated mechanically during emplacement and solidification of the enclosing granitoid. Small amounts of interstitial quartz and alkali feldspar, especially where they are concentrated in the margins of the mafic inclusions, probably were introduced by diffusion through the melt phase. Euhedral crystals of hornblende and biotite that lie across the boundaries of mafic inclusions probably crystallized late, directly from the melt phase of the magma. Granitoids in which almost all the hornblende and biotite crystals are euhedral generally contain few mafic inclusions; for example, the mafic inclusions in the Half Dome Granodiorite are confined to swarms near its outer margins.

Relatively abundant needlelike apatite crystals are the most likely survivors from the original crystallization of the mafic inclusions. Experiments by Wyllie and others (1962) on the system CaO-$CaFe_2$-P_2O_5-H_2O-CO_2 indicate that slow cooling and crystallization yields stubby crystals of apatite, whereas sudden cooling yields acicular crystals. The acicular apatite crystals in some granitic rocks—for example,

FIGURE 17.—Mafic dike (left side of photograph) truncating foliation—represented by oriented, double-convex mafic inclusions—in the Bass Lake Tonalite (more leucocratic host, right side of photograph). Roadcut about 9 km southwest of North Fork. Strontium isotopic ratios for these rocks are plotted in figure 18.

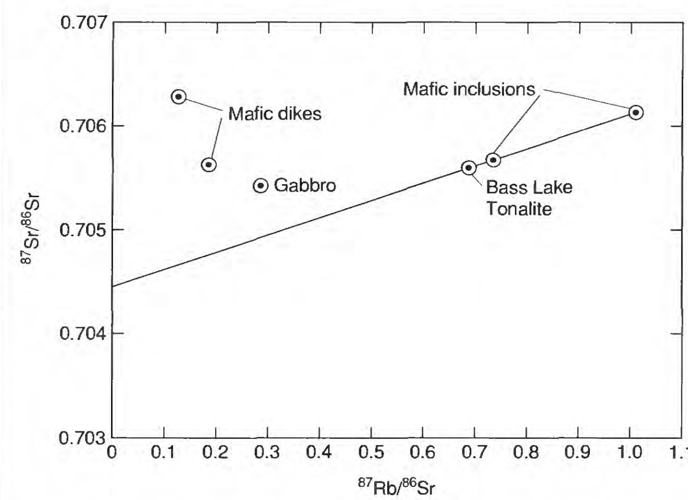


FIGURE 18.—Strontium isotopic ratios of the Bass Lake Tonalite, mafic inclusions, and mafic dike shown in figure 17 and of nearby body of gabbro. Isochron representing age of 114 Ma (from U-Pb data) is drawn through plots of mafic inclusions and tonalite.

the Tinemaha Granodiorite—may have been derived from mafic inclusions.

The fine grain size of the mafic inclusions and even finer grained margins of some inclusions are frequently cited as evidence that the inclusions were blobs of basaltic magma that were quenched in the cooler granitic magma. The assumption implicit in this interpretation is that the inclusions were formed with their present minerals and textures. The variety of shapes of the inclusions in different parts of an intrusion indicates that during their residence in granitoid magma they themselves were magma, though more viscous than the enclosing granitoid magma. Although the inclusions could have been fine grained when they originated, they were not necessarily so. If the inclusions originally were assemblages of anhydrous minerals, their grain size could have been reduced when they were converted to hydrous minerals or during the long period of continuing equilibration with their cooling and crystallizing host. In many granitic rocks, potassium metasomatism has transformed hornblende crystals to aggregates of tiny biotite and plagioclase grains, and similar recrystallization could have occurred in other minerals.

Possible sources for the inclusions are (1) refractory residue of old lower-crustal rocks from which the granitic magmas formed (restite), (2) early-crystallized cumulate or peripherally accreted minerals from granitoid magma, (3) mafic igneous or impure calcareous wallrocks, (4) blobs of basaltic, mantle-derived magma that mingled with anatectic crustal magma in forming the parent magmas for the granitoids, and (5) mafic dikes of mantle or lower-crustal origin that intruded the crystallizing but still-mobile granitoid magma and subsequently disrupted.

Reid and others (1983) and Vernon (1983) envisaged the mafic inclusions as being primary basaltic magma that mingled with and was quenched in granitic magma. In contrast, Noyes, Frey, and Wones (1983) considered wide variations in the modal proportions and major-element compositions of the mafic inclusions in the granodiorites of Red Lake and Eagle Peak to be inconsistent with an origin from crystallizing melt and consistent with their formation from entrained residual material (restite). Sawka (1985) considered high radiogenic-strontium content and higher REE concentrations in the mafic inclusions than in the host Tinemaha Granodiorite to be incompatible with either a primary magma source or a cumulate origin.

At this time the origin of mafic inclusions in granitic rocks remains an enigma, and it is possible that they were formed in more than one way. Current evidence favors most of them having originated as blobs of mantle-derived basalt that mixed and mingled with crustal anatectic magma to produce the magma from which the granitoids crystallized. Mingling probably involved the intrusion of dikes and their subsequent disruption. Some minerals (such as acicular apatite) could have crystallized in the basaltic magma when it first came in contact with cooler anatectic crustal magma, but the similar compositions of the minerals in the inclusions and their host granitoids and the subsequent mobility of the inclusions during final emplacement of the granitoids precludes the bulk of the minerals in the inclusions having crystallized at that time. They must have attained their present compositions concurrently with crystallization of the granitoids.

Combined modal plots of the granitoids and the mafic inclusions form an arcuate field that shows some interesting and probably genetic relations (fig. 19*A*). Mafic inclusions and mafic dikes plot in the part of the field that extends toward mafic and accessory minerals, and granitoids plot in the part that extends toward quartz and alkali feldspar.

A plot of CIPW norms (fig. 19*B*) of granitoids, mafic inclusions, and mafic dikes shows the same curved pattern as the modal plot. In the CIPW plot, the norms of country rocks within and east of the Sierra Nevada (table 2), graywackes of the Franciscan assemblage, and typical high-alumina basalts also are shown. The distribution of norms suggests that mixing of the basalt and country rocks could have produced the parent magmas of the granitoids, which then fractionated along the curved trend.

SCHLIEREN AND MULTIPLE LAYERING

Dark streaky layers common near the margins of many intrusions and present in a few dikes are called schlieren. In some studies mafic inclusions have been included with schlieren, but workers in the Sierra Nevada have followed the usage of Ernst Cloos (1936), who included only rock with layered or streaky structure in describing schlieren. Schlieren have many different forms, but most are composed of successions of layers in which the individual layers grade from dark fine-grained rock in one side to felsic coarser grained rock in the other (fig. 20).

Unconformities are common in layered sequences and clearly indicate that the coarser grained sides of the layers are younger than the finer grained sides. Unlike most graded beds in sedimentary rocks, the smallest minerals (composed mostly of dark minerals)

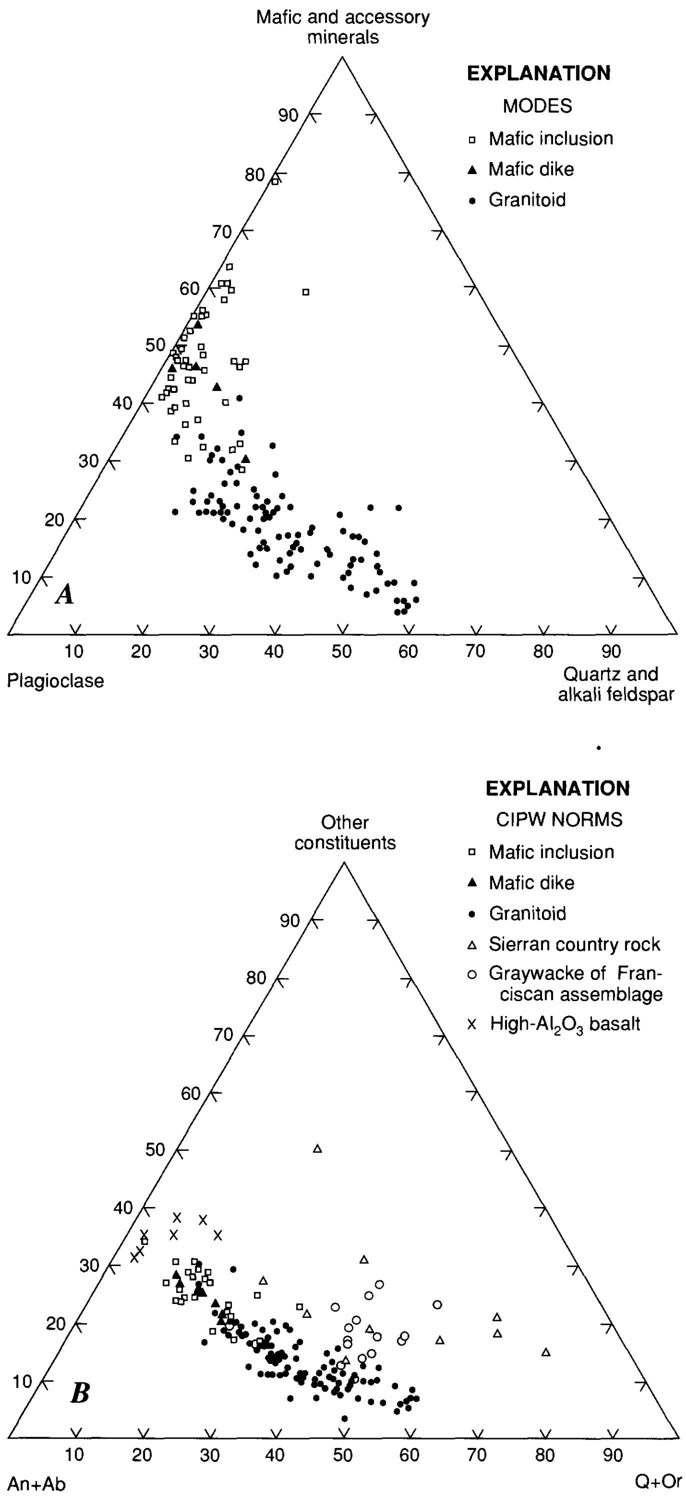


FIGURE 19.—Modes and norms of chemically analyzed samples of granitoids (Bass Lake Tonalite, Dinkey Creek Granodiorite, and Mount Givens Granodiorite) and of mafic inclusions and mafic dikes within those rocks. *A*, Modal analyses. *B*, CIPW norms, including data for country rocks within and east of Sierra Nevada (table 2), graywackes of the Franciscan assemblage (Bailey and others, 1964), and typical high-alumina basalt (Carmichael and others, 1974). An+Ab, normative plagioclase and albite; Q+Or, normative quartz and orthoclase.

are in the lower parts of the layers. Dips of layers range from horizontal to vertical, and highly complicated patterns that involve unconformities are common. Two mechanisms, operating either independently or concurrently, account for the schlieren: the sorting of grains by magmatic flow and the gravitational settling of crystals during agitation of crystals.

Ernst Cloos (1936), following an earlier interpretation by his brother, Hans Cloos (1922), attributed the size grading to the flow sorting of mineral grains. Bagnold (1954) showed that flow sorting is a mechanically viable process by immersing spherical grains in a fluid of the same density in the annular space between two coaxial drums. Rotation of one of the drums produced shear stress that caused the grains to collide with one another and to disperse in accordance with the squares of their diameters. The effect was to sort the grains by size, the largest grains being displaced into regions of lowest shear stress. More recently, Bhattacharji and Smith (1964), Komar (1972a, b; 1976), and Barrière (1976) further explored the mechanism of flow sorting. Shear stress reflects velocity differences (not velocity alone); a simple illustration of the pattern of shear stress resulting from convection along an interface is presented in figure 21.

Nevertheless, gravity settling probably also is effective in the formation of some schlieren. The mechanism envisaged is similar to the mechanism by which gold is concentrated by panning. With agitation, the smaller and heavier particles sift downward through the larger and lighter particles. Support for this settling mechanism comes from the observation that nowhere have the grains in schlieren been seen to increase in size downward from a horizontal or inclined contact and from a simple experiment performed by Coats (1936).

Coats introduced labradorite and hedenbergite grains of the same size into a cylindrical glass vial containing a mixture of bromoform and alcohol having a slightly lesser density than the labradorite. After agitation, the hedenbergite grains settled faster than the labradorite grains and began collecting at the bottom of the vial beneath an accumulating mesh of labradorite grains that were buoyed upward by a current created by the settling grains. Eventually the mesh of labradorite grains became too tight for further passage of the hedenbergite grains, and a second layer of hedenbergite began to collect above the labradorite mesh and beneath a second developing mesh of labradorite. The use of hedenbergite grains smaller than the labradorite grains would have approximated the relations in schlieren more closely and probably improved the definition of layers.

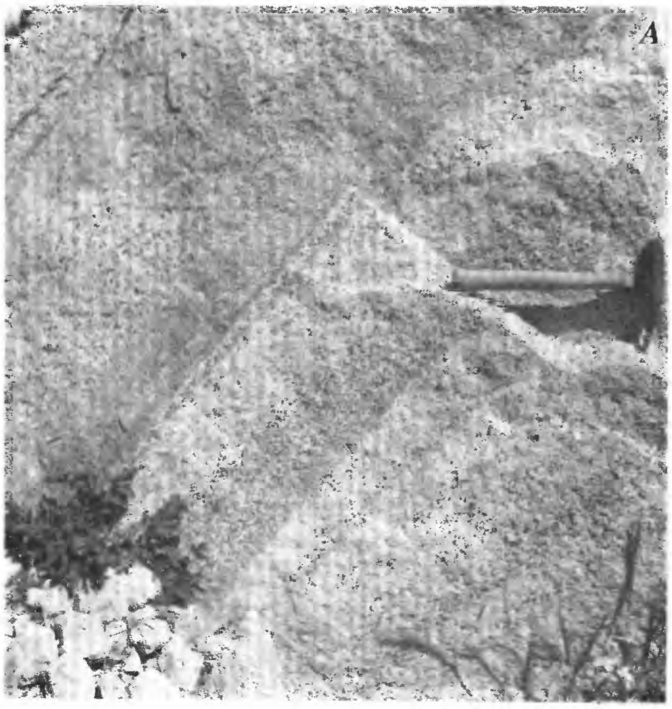
TABLE 2.—*Major-element oxide analyses of and isotopic data for country rocks within and east of the Sierra Nevada*

[All samples are chip samples that were collected across lithologic units or groups of units. Isotopic data from Kistler and Peterman (1973) and DePaolo (1981). T, 100 Ma; —, not applicable; n.d., no data or not determined; tr, <.05]

Sample	FD–32, 33, 34, 42[1]		FD–38[2]	FD–39, 46[3]		BM[4]	FD–56[5]	D–57 to FD–63[6]		PC–1 to PC–9[7]		PC–10 to PC–25[8]		PC–27[9]	PC–30a to PC–31[10]	
	Range	Average		Range	Average			Range	Average	Range	Average	Range	Average		Range	Average
Major oxides, in weight percent																
SiO_2	52.7 –66.1	60.5	56.5	74.6 – 74.9	74.8	72.7	40.0	33.3 –73.2	58.5	59.8 – 88.3	70.0	48.0 –77.2	60.0	66.9	60.9 –83.8	69.2
Al_2O_3	14.4 –15.4	15.0	15.0	—	11.6	6.4	7.1	10.6 –16.4	13.6	5.8 –17.5	14.2	4.1 –26.2	14.5	15.0	6.3 –19.1	14.2
Fe_2O_3	2.9 – 4.4	3.5	4.1	1.4 – 2.4	1.9	.63	1.0	.78 – 3.8	1.3	.65 – 1.9	1.1	.90 – 2.9	1.9	2.0	.60 – 3.6	2.1
FeO	2.5 – 5.5	3.7	4.8	1.3 – 2.0	1.7	1.7	.88	1.1 – 3.2	2.5	.68 – 7.0	3.7	2.2 –12.3	5.3	3.6	1.8 – 4.1	2.6
MgO	2.8 – 5.9	3.8	5.3	1.5 – 1.8	1.7	2.2	8.5	1.0 – 6.0	2.7	.26 – 4.6	1.7	2.6 – 5.5	4.2	1.3	.90 – 3.0	2.0
CaO	2.2 – 6.7	4.4	6.3	.76 – .94	.85	9.0	18.4	.55 –23.2	6.7	.20 – 1.9	.94	1.6 – 8.5	4.7	3.3	.58 – 1.5	1.1
Na_2O	2.1 – 3.4	2.6	2.2	.95 – 1.5	1.0	.32	.46	.71 – 2.2	1.6	.09 – 4.1	1.6	.6 – 3.6	1.8	3.2	.87 – 2.6	1.4
K_2O	.94 – 1.6	1.3	.96	2.6 – 2.7	2.7	2.0	1.3	2.2 – 3.9	3.3	2.2 – 4.6	3.9	.60 – 6.7	2.6	2.3	1.6 – 3.3	2.4
H_2O^+	2.2 – 3.4	2.7	2.3	1.6 – 2.9	2.3	.62	1.2	1.3 – 3.0	1.2	.83 – 3.6	1.9	.64 – 3.7	2.3	.72	.83 – 3.2	2.4
H_2O^-	.30 – .80	.62	.52	.21 – .47	.34	.04	.19	.15 – .24	.3	.06 – .18	.10	.12 – .70	.28	.08	.14 – .28	.21
TiO_2	.60 – .90	.64	.85	.53 – .55	.54	.36	.38	.46 – 1.3	.9	.27 – 1.4	.71	.44 – 1.6	1.0	.99	.49 – 1.1	.70
P_2O_5	.16 – .17	.17	.16	.09 – .14	.12	.15	.21	.08 – .20	.8	.04 – .21	.11	.08 – .28	.13	.15	.00 – .09	.04
MnO	.09 – .17	.12	.13	—	.06	.00	.07	.04 – .07	.06	.00 – .36	.10	.04 – .23	.11	.07	.04 – .08	.06
CO_2	tr – .08	tr	tr	tr	tr	3.5	20.5	tr –19.4	5.8	.06 – .65	.19	.01 – 1.9	.57	.02	.06 – .78	.31
Strontium and neodymium isotopic data																
$^{87}Sr/^{86}Sr_{(T)}$	0.7041– 0.7059	0.7047	0.7048	0.7120– 0.7135	n.d.	0.7132	n.d.	n.d	n.d.	n.d.	n.d.	n.d.	0.7150	0.7270	—	0.7360
$\varepsilon Sr_{(T)}$	n.d.	-9	n.d.	60 –131	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	450
$\varepsilon Nd_{(T)}$	n.d.	4.6	n.d.	-2.2 – 10.8	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	15.6

[1]Mesozoic volcanic and plutonic strata west of the Melones fault zone and along California State Highways 108 and 132 and Merced County Road 116.
[2]Greenstone of Bullion Mountain along California State Highway 140.
[3]Calaveras strata along California State Highways 120 and 140.
[4]Pennsylvanian and Permian(?) strata in Bloody Mountain block of Mount Morrison roof pendant.
[5]Paleozoic strata, ranging in age from Middle Cambrian to Permian, exposed in Mazurka Canyon, southern Inyo Mountains.
[6]Proterozoic and Early Cambrian strata in the White Mountains.
[7]Precambrian rocks from Pleasant Canyon, Panamint Range. PC–1 and PC–7 are sheared granite; other samples are sedimentary rocks of the Proterozoic Pahrump Group.
[8]Precambrian igneous and metamorphic rocks from Wildrose Canyon, Panamint Range.
[9]Gneiss from Funeral Peak, Death Valley.
[10]Precambrian strata from north end of Funeral Mountains. Only one initial strontium ratio was determined; result is shown under "Average."

In some plutons and intrusive suites, such as the Tuolumne Intrusive Suite, layered schlieren in both flanks dip outward from the core (figs. 20*A* and 22). This arrangement suggests that episodic resurgence in the core of the suite caused outward tilting and gravitational shearing on the partly solidified margins. Shearing parallel with dike walls during crystallization probably accounts for schlieren in dikes.



COMB LAYERING AND ORBICULES

Comb layering, like schlieren, generally is found near contacts and is composed of alternate felsic and mafic layers in which the long axes of elongate crystals lie across the layers. According to Moore and Lockwood (1973), comb layering forms chiefly in steeply inclined inverted troughs along overhanging walls of plutons or along the walls of dikes or pipes that cut country rocks adjoining the pluton. They report that flow layering along steeply dipping walls gives way to comb layering where the wall passes through the vertical plane and is overturned. Moore and Lockwood (1973) attributed comb layering to the upward migration of solute-rich aqueous fluids along pluton walls, especially in steeply inclined troughs in overhanging walls. The similarity of comb layering to layering in many pegmatites supports participation of an aqueous fluid.

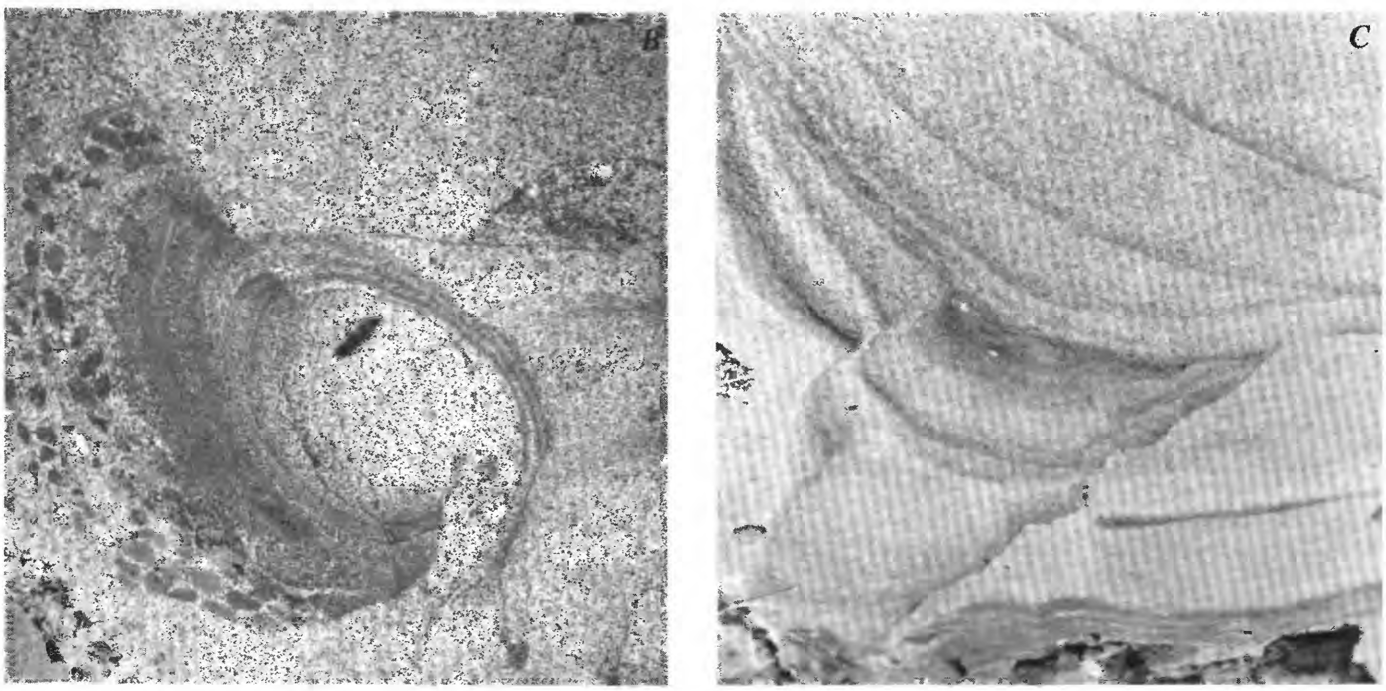


FIGURE 20.—Schlieren. *A*, Planar schlieren in the Half Dome Granodiorite at Tenaya Lake. *B* and *C*, tubular (faerie ring) schlieren and trough schlieren, respectively, in the Mount Givens Granodiorite at Courtright Geologic Area near Courtright Reservoir (same scale as preceding photographs).

Orbicules as much as 30 cm across are associated with comb layers in many places. The orbicules consist of a core composed of metamorphic rocks, the common granitic rock of the area, or comb-layered rock and surrounded by a series of comb layers identical to those in the adjacent comb-layered rock (fig. 23). Apparently they occur in pipelike structures. Moore and Lockwood (1973) postulated that the orbicules grew by precipitation around fragments that were suspended within an upward-moving aqueous fluid, in somewhat the same way as hailstones grow in the atmosphere or as calcareous ooliths form in agitated water.

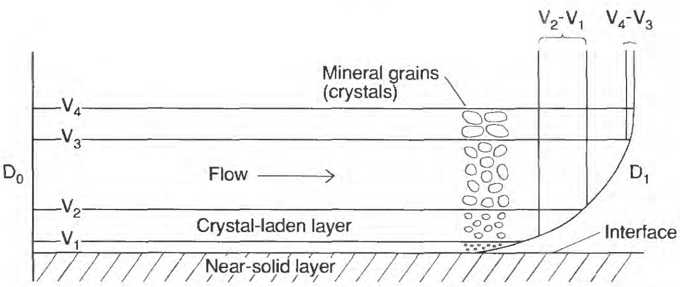


FIGURE 21.—Flow sorting of mineral grains in magma. Flow is assumed to be from line D_0 toward curve D_1 and parallel to interface that corresponds to boundary surface between two layers—one almost solid and one fluid but laden with crystals. Distance between line D_0 and curve D_1 represents distance grains would travel during an interval of time, and lines V_1 to V_4 represent velocities at different distances from interface. Because of drag, distance traveled in a unit of time and velocity decrease toward interface. Thus, velocity difference between paths same distance apart increases toward interface, and velocity difference between V_4 and V_3 is less than between V_2 and V_1. Velocity difference causes grains to collide and to be dispersed into regions where velocity difference is less. Larger grains collide with other grains more often than do small grains, and they are dispersed more effectively.

JOINTS AND MICROFAULTS

Joint sets can be seen in almost all exposures of granitoids, and lineaments extending great distances are conspicuous features on aerial photographs of the higher, better exposed parts of the Sierra Nevada (fig. 24). Typically, joints dip steeply and occur in conjugate sets. Those in a given set usually are almost parallel, but locally they intersect. Most are linear or gently curved; where abrupt changes in strike occur, joints of one strike interfinger those of another.

The joint patterns are regional in extent, and joints cross contacts between intrusions with little or no deflection. Joints related to individual intrusions such as those described by Balk (1937) have not been recognized. The spacing of joints is variable, and Bateman (1965b) observed that in the eastern escarpment of the Sierra Nevada west of Bishop the spacing is roughly proportional to the grain size of the rock in which they occur. Joints in medium- to coarse-grained rock may be spaced many tens of meters apart, whereas those in adjacent, fine-grained rock may be less than 1 cm apart. Erosion along widely spaced joints can develop deep slots in the rock, whereas rock with closely spaced joints breaks down to rubble.

Joints are not to be confused with gently dipping sheeting (fig. 25), such as that responsible for the magnificent domes of Yosemite. Sheeting strikes parallel to the topographic surface and dips more gently than the surface. Sheeting is the result of reduced load pressure caused by erosion and is a near-surface phenomenon. The action of ice and water facilitates the dislocation of rock slabs.

Segall and Pollard (1983) concluded that the joints formed earlier than sheeting and at greater

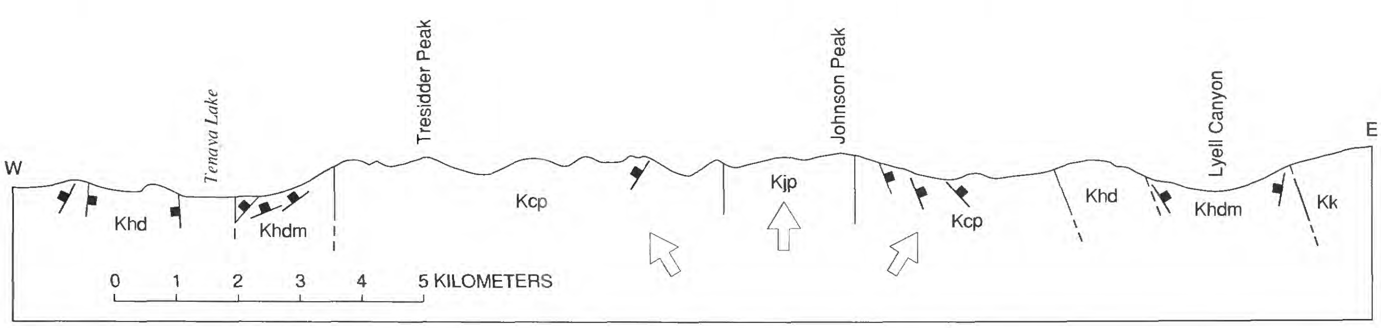


FIGURE 22.—Projection of schlieren on east-west profile across the Tuolumne Intrusive Suite. Square-flagged symbols are aligned with schlieren layers; inferred shear stress parallels symbols' long bars, and their square flags indicate direction in which grain size increases within individual schlieren layers. Arrows indicate direction of repeated intrusion, uplift, and expansion. Kjp, Johnson Granite Porphyry; Kcp, Cathedral Peak Granodiorite; Khdm, megacrystic facies of the Half Dome Granodiorite; Khd, equigranular facies of the Half Dome Granodiorite; Kk, granodiorite of Kuna Crest.

depths. The exact time when the joints were formed is not known. The regional distribution of the joints and their independence of pluton boundaries indicates that they are younger than the batholithic rocks. Bateman and Wahrhaftig (1966) reported that in the northern Sierra Nevada deep weathering beneath the Eocene auriferous gravels extends downward along joints into unaltered rock, indicating that the joints there are at least as old as Eocene; probably they formed during the Late Cretaceous as the Sierra Nevada was being uplifted, eroded, and cooled (see section "Westward Tilting and Uplift of the Sierra Nevada").

Segall and Pollard (1983) studied the joints within the Mount Givens Granodiorite in a small area near Florence Lake. Detailed field measurements and examination of thin sections show that displacements are normal to the joint surfaces and that the measured extensional strain in this area accommodated by joint dilation is on the order of 1×10^{-4} to 5×10^{-4}. They observed that individual joints actually consist of numerous subparallel, planar segments.

Many joints were filled with quartz, epidote, chlorite, or sulfides during or immediately after they were formed; some contain gouge and exhibit subhorizontal slickensides and offsets in their walls, indicating strike-slip displacement. Lockwood and Moore (1979) called the joints having strike-slip movement microfaults. Obviously the strike-slip movements occurred after the joints were formed by dilation; Lockwood and Moore (1979) stated that in one place they offset a late Miocene dike. They determined that most, and perhaps all, of the lineaments seen on aerial photographs are microfaults and concluded that far from being a monolithic block during the late Mesozoic and Tertiary, the Sierra Nevada was deformed along with the Basin and Range province to the east. However, the deformation in the Sierra Nevada is shear deformation dominated by strike-slip faulting along conjugate microfaults, whereas that of the Basin and Range province is dominated by normal faulting. Microfaults trending generally northeastward have sinistral movement, whereas those trending northward have dextral movement. Lockwood and Moore (1979) determined that extensional strain resulting from movement on the microfaults changed progressively from west-northwest at lat 38.5° N. to northwest at lat 36.5° N. A pure-shear constant-volume study based on detailed work at lat 37°20' N. indicated a maximum extension of 2.3 percent in a N. 61° W. direction.

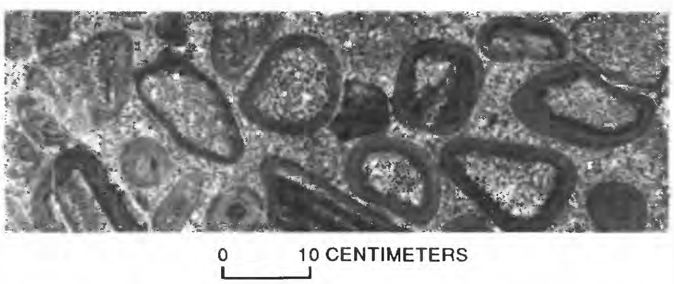


FIGURE 23.—Orbicules in mafic rocks of the Dinkey Creek Granodiorite.

FIGURE 24.—Joint sets in the leucogranite of Rawson Creek. View looking east across Owens Valley.

FIGURE 25.—Sheeting in the Sentinel Granodiorite due to unloading by erosion. Roadcut along California Highway 120 (Tioga Pass Road).

This direction is approximately parallel to the direction of tectonic extension in the Basin and Range province during the Cenozoic (Zoback and Zoback, 1980).

MAGMA EMPLACEMENT AND SOLIDIFICATION

DEPTH OF EROSION SINCE PLUTONISM

To make the best use of information derived from the exposed rocks, the amount of erosion that has taken place since the plutonic rocks were emplaced and the load pressure that then existed should be known. Unfortunately, there is no simple way to determine the thickness of rock stripped from the batholith. Bateman and Wahrhaftig (1966) attempted to estimate the thickness of the material that was removed by calculating the amount of eroded material deposited in adjacent areas since the batholith was emplaced, but numerous uncertainties made even a rough estimate by this method impossible.

The vertical or steeply dipping contacts of most plutons with metamorphic or older plutonic rocks, a general absence of miarolitic cavities (except in subvolcanic porphyries and a few fine-grained felsic rocks), and a paucity of fractures attributable to intrusion indicate depths generally greater than those at which brittle fracturing would occur. However, some field relations indicate that the batholith was not deeply eroded. Gently dipping felsic dikes marginal to such Sierran intrusions as the leucogranite of Bald Mountain and the Mono Creek Granite, which lifted the overlying rocks (Bateman, 1965b), the tongue of the Half Dome Granodiorite that splits the intrusive suite of Washburn Lake, horizontal and gently dipping foliations in the Lamarck and Tinemaha Granodiorites, and numerous small plutons of the Bass Lake Tonalite in the western metamorphic belt indicate proximity to the tops of these intrusions. Gently outward dipping contacts of the Ward Mountain Trondhjemite in the western part of the batholith and gently dipping contacts of some Cretaceous plutons in the White and northern Inyo Mountains and displacements and fractures in their wallrocks suggest that the eastern and western parts of the batholith may be less deeply eroded than the central part.

The presence of andalusite and sillimanite and the absence of kyanite in thermal aureoles indicate that the load pressure at the presently exposed level was less than that at the triple point for the aluminum silicates at the times of intrusion and contact metamorphism. Holdaway (1971) placed the triple point at 3.8 kbar and Robie and Hemingway (1984) at 4.0±0.5 kbar—equivalent to the weight of about 13 km of eroded rock. This is the maximum average thickness of rock that could have been eroded from the central Sierra Nevada since the batholith was emplaced. However, Ross (1983) reported hornblende-rich garnet- and hypersthene-bearing rocks in the southernmost Sierra Nevada; he attributed the presence of this mineral assemblage to a deeper level of emplacement than is indicated for the central Sierra Nevada.

The geobarometers that have been used to estimate emplacement pressures and depths in the Sierra Nevada have the limitation that they depend on magma saturation with water and indicate P_{load} only where it can be established that the required amount of water was present; otherwise, they indicate only the lower limiting pressure. These geobarometers indicate pressures in the general range of 1 to 2 kbar—equivalent to the weight of 3.5 to 7 km of rock.

Using the position of the temperature minimum in the salic tetrahedron and assuming water saturation (see fig. 26), Putnam and Alfors (1965) inferred that P_{H_2O} during crystallization of the Late Jurassic Rocky Hill stock (in the western foothills south of the map area) was 1.5 kbar. Using the same method, Noyes, Wones, and Frey (1983) deduced a pressure of about 1 kbar during the crystallization of late-stage aplites that had intruded the granodiorites of Eagle Peak and Red Lake. Sylvester and Nelson (1966) suggested that P_{H_2O} equals P_{total} and is 2 to 3 kbar for the Birch Creek pluton in the White Mountains on the basis of the presence of muscovite in the granite and the metamorphic grade attained in the aureole. However, after discovering that the muscovite is fluoromuscovite, which is stable over a wide range of temperatures and pressures, they concluded simply that the pressure and temperature were low when the pluton was emplaced (Nelson and Sylvester, 1971). Wones and others (1969) interpreted the assemblage biotite-alkali feldspar-magnetite in the granite of Dinkey Dome to indicate that P_{fluid} and P_{total} were about 2 kbar when the granite was emplaced. They cited the presence of miarolitic cavities in the granite as evidence that P_{fluid} equaled P_{total} before crystallization was completed. Brown (1980) interpreted the coexistence of biotite and iron-rich cordierite [Fe/(Fe+Mg)=0.7] together with alkali feldspar, muscovite, andalusite, and quartz to indicate pressures of 1.5 to 2 kbar in the country rocks at the Pine Creek tungsten mine near Bishop.

PHYSICAL CONSTITUTION OF SIERRAN MAGMAS

Four lines of evidence—involving deformation structures, petrography, metamorphic petrology, and experimental igneous petrology—no one of which is completely convincing by itself, indicate that most granitic magmas of the Sierra Nevada consisted of melt plus crystals and were undersaturated with water at the present levels of exposure until late stages of crystallization. Exceptions are aplite and a few small bodies of fine-grained granite that crystallized in a narrow, low-temperature range and may have been emplaced as melt or fluxed with volatiles.

The interpretations offered for the origin of magmatic foliation and lineation (see section "Primary Magmatic Foliation and Lineation") and for schlieren (see section "Schlieren and Multiple Layering") require that crystals were present in the magma when these structures were formed. However, the presence of magmatic foliation and lineation does not rule out the possibility that the magma was free of crystals earlier, when it was rising or during initial emplacement.

Mottled cores of plagioclase crystals and mafic inclusions, including small clots of hornblende, biotite, opaque minerals, titanite, zircon, and other accessory minerals and later alteration products, have been interpreted to be restite (Presnall and Bateman, 1973). If this interpretation is correct, the magma never was a complete melt.

During the progressive melting of a volume of rock, it is highly probable that the volume would become unstable and move buoyantly upward, following lines of weakness within the host rock, long before the temperature of the liquidus was reached. Van der Molen and Paterson (1979) determined that the strength of granite decreases rapidly with partial melting and that at 30 to 35 volume percent melt, granular-framework behavior changes to suspensionlike behavior. With this amount of melt a magma can move freely in response to gravity and tectonic forces.

The relatively low temperatures indicated by the mineral assemblages in thermal aureoles suggest that the temperatures of the adjacent magmas were substantially below their liquidus temperatures as inferred from laboratory experiments. Most Sierran hornfelses have been considered to be in the hornblende hornfels facies of thermal metamorphism. However, some assemblages in the interiors of large roof pendants are in the lower temperature albite-epidote hornfels facies, and some within a few meters of quartz diorite and tonalite intrusions are in the higher temperature alkali feldspar—cordierite hornfels (pyroxene hornfels) facies. The temperatures at the upper and lower boundaries of the hornblende hornfels facies have been estimated by Winkler (1979) to be 540±15 °C and 620±20 °C, respectively. Kerrick (1970) inferred that the formation temperatures of several different mineral assemblages in the Saddlebag Lake septum close to the Tuolumne Intrusive Suite were in the range of 530 to 630 °C, and Sylvester, Oertal, and others (1978) concluded that the temperature of metamorphism adjacent to the granite of Papoose Flat did not exceed 600 °C. Morgan (1975) estimated that the formation temperature of the garnet-pyroxene skarn in the Mount Morrison roof pendant ranged from 500 to 600 °C, and Newberry and Einaudi (1981) estimated the formation temperature of similar skarns to be between 550 and 650 °C.

Relating these temperatures of thermal aureoles to the temperatures of adjacent magmas is fraught with uncertainties because magmatic temperatures were undoubtedly considerably higher. Winkler (1979) stated that the temperature at the contact is higher than the sum of the ambient temperature of the wallrocks before intrusion and 60 percent of the temperature of the magma. If the temperature of the contact is assumed to be about 650 °C (the maximum temperature inferred from mineral assemblages in thermal aureoles) and the ambient temperature of the wallrocks is taken to be 150 °C, the temperature of the magma would be 830 °C, using Winkler's formula. A magmatic temperature of 1,000 °C would be reached if the temperature at the contact was 750 °C and the ambient temperature of the wallrocks 150 °C.

The water content of a magma is critical to the liquidus temperature and to the temperatures at which various minerals begin to crystallize. The liquidus temperature of completely dry magma has been determined experimentally to approach 1,200° at 2- and 8-kbar confining pressure but to decrease with increasing amounts of water (Robertson and Wyllie, 1971; Naney, 1977). The almost complete absence of hydrothermal effects, the restriction of miarolitic cavities to a few small, fine-grained masses of leucocratic granite, and a dearth of pegmatite dikes and quartz veins indicate that most Sierran magmas were undersaturated with water at levels presently exposed when they were emplaced and remained undersaturated during most of their periods of crystallization. The water content probably was greater at higher levels and may have exceeded saturation at the highest levels.

Naney (1977) determined experimentally that at 2-kbar total pressure (equivalent to a depth of about 7 km) hornblende is stable in magma of granodiorite composition only with at least 4 percent water. (A small amount of fluorine would reduce the required amount of water.) However, the hornblende crystals in many granodiorites and tonalites contain clinopyroxene cores, which must have crystallized in the presence of smaller amounts of water. An initial water content of about 2 percent or less for these Sierran magmas seems reasonable. The experimentally determined liquidus temperature of average granodiorite magma at 2-kbar total pressure and 2 weight percent water is about 1,000 °C (Naney, 1977), a temperature substantially higher than the magmatic temperature inferred from the mineral assemblages found in thermal aureoles.

MODE OF EMPLACEMENT (THE SPACE PROBLEM)

Space for the batholith was provided in a variety of ways: incorporation of crustal materials in the magmas, forcible displacement of wallrocks and roof rocks, stoping, extension across the area of the batholith, erosion, and expulsion of material during volcanism. Need for increased space is indicated by isotopic ratios. Initial $^{87}Sr/^{86}Sr$ values ranging from about 0.704 in the Fine Gold Intrusive Suite to greater than 0.710 in some intrusions in the White Mountains indicate that most of the granitoid magmas contained substantial amounts of crustal material. Conversely, they also indicate that most of the magmas contained material derived from the mantle and that extra space was required to accommodate the extra material. If the magmas had been derived entirely from crustal sources, a space problem would not have existed. Evidence for the different processes and mechanisms of space making is given in the following four sections.

FORCIBLE EMPLACEMENT

Many plutons have made space for themselves by crowding the rocks into which they are emplaced outward and upward, but unless the wallrocks on the two sides of the batholith spread apart, the displaced material must have ultimately moved upward, where it was removed by erosion, or downward, where it replaced rising magma. Thus, forcible emplacement can provide space for an individual pluton but is only a contributing factor in making space for the batholith as a whole.

Mushroom-shaped plutons appear to have formed at relatively shallow depths but within the zone of ductile flow during late stages of crystallization, when the difference between the viscosities of the magma and wallrocks was small and the magmas tended toward spherical forms with minimum surface areas. Most, and probably all, outward bulges of plutonic rocks at the surface are underlain at relatively shallow depths by older rocks. In most places where granitic magmas have crowded the adjacent country rocks aside, felsic magmas were involved; their lower temperatures and higher viscosities made them less subject to control by regional structures and permitted them to attain rounded shapes with minimum surface areas. Exceptions to this generalization are the lateral protrusion of the Bass Lake Tonalite into the southern part of the western metamorphic belt near Mariposa and of the Dinkey Creek Granodiorite into the Bass Lake Tonalite in the vicinity of Shaver Lake.

Much evidence for forcible emplacement is described in other parts of this report. Foliation patterns in older granitoids that were disturbed during the forcible emplacement of younger intrusions and ductile deformation of older plutons and metamorphic wallrocks during the emplacement of younger intrusions in late stages of crystallization are described in the section "Deformation Associated With Emplacement of Intrusions," whereas dislocations of the stratified country rocks during the forcible emplacement of intrusions are described in this section. Excellent examples of the deformation of wallrocks by intrusions are present in the western foothills, in the eastern Sierra Nevada, and in the White and northern Inyo Mountains. A few particularly noteworthy examples are cited in the following paragraphs.

In the western part of the map area, south of Mariposa, a lens of serpentinite that lies within the south end of the Melones fault zone bends from its regional southeasterly trend due west along the north side of a lobe of Bass Lake Tonalite for a distance of more than 2 km before being cut off by the tonalite. The distribution of rocks in the vicinity indicates that a lobe of tonalitic magma moved westward, and probably also upward, crowding the country rocks aside and dragging them along its margins.

In the eastern Sierra Nevada, strata in the Pine Creek and Bishop Creek septa and Mount Morrison roof pendant have been disturbed by several intrusions (Rinehart and Ross, 1964; Bateman, 1965b, pl. 8). The largest displacement is indicated by the Wheeler Crest septum, which extends northward discontinuously from the north end of the Pine Creek septum (fig. 2), bending around the east side of the Round Valley Peak Granodiorite.

The strata in the septum are continuous, except for an interruption by a small diorite pluton, with strata in the Pine Creek septum and are correlative with Pennsylvanian and Permian(?) strata in the Mount Morrison roof pendant to the north. The spatial relations suggest that before the Round Valley Peak Granodiorite was emplaced, the strata in the Wheeler Crest septum continued northward on strike with strata in the Pine Creek septum to the correlative strata in the Mount Morrison roof pendant. The strata in the Wheeler Crest septum were pushed eastward 13 km when the Round Valley Peak Granodiorite and cogenetic Mono Creek Granite were emplaced.

In the southern part of the Mount Morrison roof pendant, metavolcanic strata follow the southern contact of the northwesternmost pluton of Round Valley Peak Granodiorite (called the Lee Lake mass by Rinehart and Ross, 1964) through a 45° arc, indicating forcible dislocation by the Round Valley Peak Granodiorite. Farther south, in the southern part of the Pine Creek septum and in a ring of metamorphic rocks that encircles Mount Humphreys, metavolcanic rocks bulge eastward, bending and truncating metasedimentary strata that continue to the north and south. Although this eastward bulge of metavolcanic rocks into metasedimentary rocks could conceivably have resulted from regional thrusting, the coincidence of the bulge with a protrusion of leucocratic-facies rocks of the Lake Edison Granodiorite strongly suggests that it was caused by forcible eastward expansion of the crystallizing viscous magma. In the northern part of the Bishop Creek septum, protrusions of the Tungsten Hills Granite dislocated an anticline and a syncline that had been formed earlier in response to regional stress. In the southern part of the septum, a tongue of the leucogranite of Rawson Creek has dragged adjacent strata westward with it.

In the White and northern Inyo Mountains, the wallrocks adjacent to the granites of Papoose Flat, Birch Creek, and Pellisier Flats were deformed concurrently with the ductile deformation of peripheral parts of these intrusions during their emplacement (see section "Deformation Associated With Emplacement of Intrusions"). The sedimentary strata around the west side of the Papoose Flat pluton have been thinned to about 10 percent of their regional thickness (Sylvester, Oertal, and others, 1978), and the strata around the west side of the Birch Creek pluton have been faulted and bowed outward for a distance approximately equal to the width of the pluton (Nelson and Sylvester, 1971).

STOPING

Stoping of blocks of country rocks can make space for intrusions by exchanging places with magma; the stoped blocks either sink to lower levels or rise to higher levels where they may be eroded away. However, no evidence has been recognized within the map area (pl. 1) for either the sinking or rising of stoped blocks. Where visible they generally are relatively small and confined to the margins of intrusions; major stoping can only be inferred from the flat or gently undulating tops of a few intrusions. Perhaps the most convincing evidence of stoping is found in the granodiorite of Jackass Meadows, which intrudes the volcanogenic Minarets sequence. Here, large blocks of the Minarets sequence occupy almost half of what is undoubtedly the top of the intrusion, and smaller fragments are widely distributed between the larger blocks. Near-horizontal foliations in the interior of the Tinemaha Granodiorite—indicating a flat top—and gentle antiforms and synforms in the foliation of the Lamarck Granodiorite—indicating an undulating top—suggest that large blocks have subsided or pushed upward to provide space for these intrusions. However, neither intrusion contains identifiable remnants of roof rocks. The presence of numerous plutons of the Bass Lake Tonalite within the country rocks near Mariposa suggests that the upper surface of the tonalite plunges gently northwestward beneath the adjacent country rocks and projects upward irregularly, but the method by which space was made for the tonalite is uncertain.

Prior to erosion, additional evidence of stoping probably would have been found at higher levels of exposure where brittle fracture is more common. Pitcher and Cobbing (1985) cited flat roofs, steep sidewalls, and absence of evidence that the wallrocks were disturbed during intrusion as evidence that subsiding crustal blocks provided space for the plutons in the coastal batholith of Peru, where the exposed level is within the zone of brittle fracture and apparently shallower than the exposed level in the Sierra Nevada.

REGIONAL EXTENSION

A dearth of evidence that any intrusion was emplaced during regional tectonism, although some intrusions obviously have deformed themselves or were subsequently deformed, suggests that most intrusions were emplaced during inactive or extensional regimes and not during tectonic compressive regimes. However, the amount of possible extension appears to have been limited. The dissimilar compositions of the Jurassic granitoids on the two sides of the batholith precludes their having been emplaced in adjacent positions and spread apart the distance between them during intrusion of the Cretaceous granitoids. Furthermore, the apparent absence of granitoids associated with the Independence dike swarm of mafic dikes, which extends from the latitude of Bishop southward beyond the south end of the Sierra Nevada (Moore and Hopson, 1961), suggests that significant extension can allow mafic magma generated in the mantle or lower crust to pass upward through the crust without ponding and mixing with crustal materials to produce granitic magmas.

The long, narrow shapes of intrusive suites and of individual intrusions, especially those of the John Muir Intrusive Suite, and the absence of convincing evidence that the deformation in the wallrocks was caused by the intrusions permit the interpretation that the wallrocks were dilated. The Lamarck and Lake Edison Granodiorites and the Mono Creek Granite have lengths of 70, 50, and 55 km, respectively, and maximum widths of 10 to 15 km. The Evolution Basin Alaskite is 30 km long, but its maximum width is only about 6 km. Some maximum widths are in areas where foliation patterns show that the magma locally bulged outward and pushed the adjacent rocks aside forcibly during late stages of crystallization when the viscosity of the magma was highest. The most obviously swollen area is the lobe of felsic rock that projects eastward from the Mono Creek Granite. Before this lobe was protruded, the Round Valley Peak Granodiorite probably was elongate and parallel to other intrusions of the John Muir Intrusive Suite. Excluding these bulges, the maximum widths of most plutons of this group are generally less than 8 km.

According to Tobisch and others (1986), homoclinal sequences of metavolcanic rocks in the Ritter Range roof pendant and Goddard septum have been tilted westward on convex-upward listric faults that flatten at depth. Such faulting requires significant regional extension.

EROSION AND VOLCANISM

Erosion that accompanied and closely followed uplift during the emplacement of the batholith and expulsion of material through volcanism provided substantial amounts of space for the Sierra Nevada batholith. Geobarometers indicate that about 3.5 to 7 km of rock has been removed from the area underlain by the batholith since the plutonic rocks were emplaced. Clasts of granitic rocks reported by Fiske and Tobisch (1978) in conglomerate at the base of the ~100-Ma Minarets sequence show that plutonic rocks were already exposed by mid-Cretaceous time, and by Eocene time very low relief characterized the region of the Sierra Nevada (the so-called Eocene peneplain). Much of the eroded rock is contained in the Great Valley sequence to the west (Dickinson and Rich, 1972). Expulsion of volcanic materials, chiefly in the form of easily eroded ash-flow tuff and airborne ash, also provided space. Ash-flow tuff would have been quickly eroded and carried westward, whereas prevailing winds would have carried the airborne ash eastward.

SEQUENCE OF CRYSTALLIZATION OF MINERALS

The order in which minerals crystallized has been partly determined in the compositionally zoned Tuolumne Intrusive Suite (Bateman and Chappell, 1979) and the bulbous north end of the Mount Givens Granodiorite of the John Muir Intrusive Suite (Bateman and Nokleberg, 1978). The order of crystallization in these suites is generally representative of the order in most granitoids in the axial parts of the Sierra Nevada batholith. The order was determined by observing the first appearance of discrete well-formed grains inward from the margins of the intrusions. Thin, interstitial grains were interpreted to have crystallized from intergranular melt.

In the Tuolumne Intrusive Suite, samples collected close to the margins contain equant and subequant grains of plagioclase, hornblende (some with clinopyroxene cores), biotite, magnetite, titanite, and apatite, but only thin elongate grains of quartz and alkali feldspar. The equant and subequant grains are interpreted to have been present in the magma when it was emplaced or to have precipitated shortly thereafter, whereas the anhedral (interstitial) phases crystallized from intergranular melt. Inward from the margins, the first subequant grains encountered are first quartz and then alkali feldspar, showing that quartz began to crystallize first and was closely followed by alkali feldspar. A similar pattern is present in the Mount Givens Granodiorite, except that tiny subequant grains of quartz in rock closest to the margin indicate that quartz had already begun to crystallize when the magma was emplaced. As in the Tuolumne Intrusive Suite, alkali feldspar began to crystallize shortly after quartz.

No evidence has been found to indicate that the Tuolumne and Mount Givens magmas ever were heated to liquidus temperatures. However, experimental studies indicate that if they were originally melts, plagioclase would have formed at the liquidus temperature, followed by clinopyroxene, which occurs as cores in hornblende (Naney, 1977); hornblende and biotite would have begun to crystallize next, but their relative order of crystallization is uncertain. In both the Tuolumne Intrusive Suite and the Mount Givens Granodiorite, a rapid inward decrease in the amount of hornblende and a slow decrease in the amount of biotite reflect progressive depletion of MgO, CaO, FeO, and Fe_2O_3 in the melt phase of the magma (see figs. 34, 35). The constant abundances of magnetite and of titanite from margin to core of the Tuolumne Intrusive Suite suggests that these minerals were formed at fairly constant rates throughout crystallization.

The sequence of crystallization in the low-potassium rocks of the western foothills is probably similar to that in the Tuolumne Intrusive Suite and Mount Givens Granodiorite, except for an extended time and temperature gap between the beginning of crystallization of quartz and the beginning of crystallization of alkali feldspar. In these rocks, quartz occurs in medium-sized subequant grains, whereas the small amounts of alkali feldspar in the tonalite, trondhjemite, and most bodies of granodiorite are almost all interstitial.

In contrast to the low-potassium rocks of the western foothills are the high-potassium Jurassic rocks of the White and northern Inyo Mountains. The small amounts of quartz in the monzonite of Joshua Flat are interstitial to the other minerals, whereas alkali feldspar in undeformed rocks forms large poikilitic crystals that enclose most of the other minerals that are present in the rock. The reverse order of crystallization of quartz and alkali feldspar and the resultant contrasting textures of the high-potassium rocks of the White and northern Inyo Mountains and the low-potassium rocks of the western foothills of the Sierra Nevada are consistent with the relations predicted by the salic tetrahedron (see fig. 26).

CRYSTALLIZATION IN RELATION TO THE SALIC TETRAHEDRON

The salic tetrahedron (fig. 26) portrays the experimentally determined relations in the system $CaAl_2Si_2O_8$-$NaAlSi_3O_8$-$KAlSi_3O_8$-SiO_2. For convenience, these components will be referred to in the ensuing discussion as An, Ab, Or, and Qz, respectively. Because the mafic constituents are omitted, the salic tetrahedron does not accurately represent the compositions of real rocks. Nevertheless, it is a useful approximation and provides a framework for understanding and following the sequence of crystallization of the felsic minerals within intrusive suites and for distinguishing between suites. The relations in this system were determined chiefly by workers at the Geophysical Laboratory of the Carnegie Institution who have experimented with both binary and ternary systems—for example, Bowen (1928), Schairer and Bowen (1947), Schairer (1950), Franco and Schairer (1951), Yoder and others (1957), Tuttle and Bowen (1958), and Yoder (1968). Some dry systems were investigated initially, but because reaction rates are extremely slow in dry systems, water was added in most experiments. The addition of water increases reaction rates and significantly lowers the temperatures of crystallization. Most studies were conducted at confining pressures of 2 or 5 kbar, but a few experiments were conducted with P_{H_2O} as low as 0.5 kbar and as high as 10 kbar.

The most realistic conditions for the Sierra Nevada batholith would be melt undersaturated with water at confining pressures of about 2 kbar (equivalent to a depth of 7 km). However, experimental data for these conditions are incomplete, and the salic tetrahedron shown in figure 26 is an approximation of relations at P_{H_2O} of 2 kbar. Changes in P_{H_2O} cause temperature differences and shifts in the positions of the two internal phase-boundary surfaces. With increasing P_{H_2O}, crystallization temperatures decrease, the quartz-saturation surface shifts toward Qz, and the temperature minimum is shifted away from Ab. However, these temperature differences and position shifts are not significant in the following discussion, which is aimed at illustrating principles rather than at duplicating the precise conditions of crystallization.

The salic tetrahedron contains three liquidus phase volumes, which are separated by two internal phase-boundary surfaces. The volume that includes the An and Ab corners of the tetrahedron is the plagioclase liquidus phase volume. Similarly, the volume that includes the Qz corner is the quartz liquidus phase volume, and the volume that includes the Or corner is the alkali feldspar liquidus phase volume. The quartz-saturation surface separates the plagioclase and quartz liquidus phase volumes, and the two-feldspar surface separates the plagioclase and alkali feldspar liquidus phase volumes. The intersection of these two surfaces is the ternary cotectic. Temperatures are highest at the corners of the tetrahedron and decrease toward the point where the ternary cotectic intersects the base of the tetrahedron.

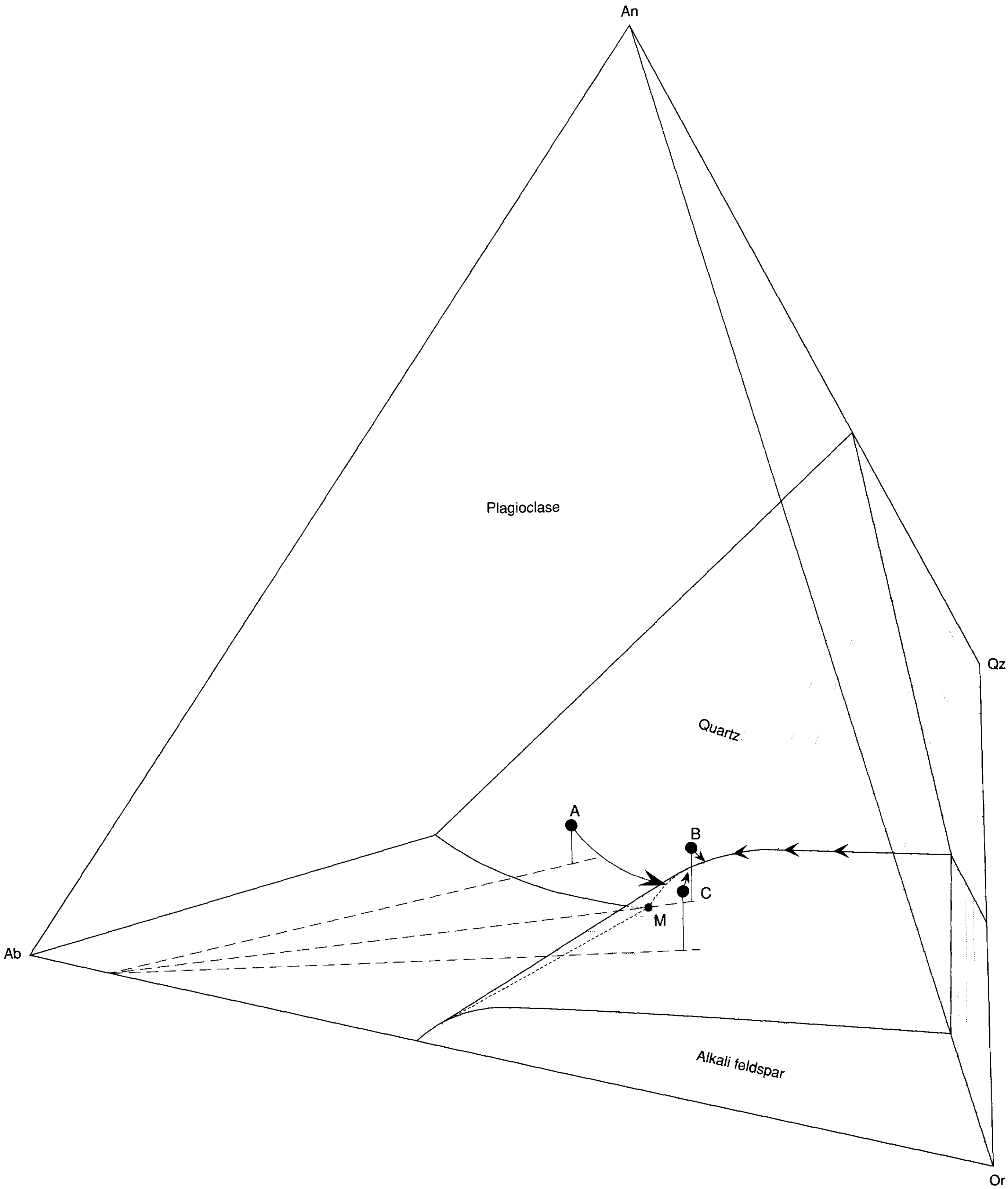


FIGURE 26.—Salic tetrahedron showing liquidus phase volumes and crystallization paths. Ab and An, albite and anorthite end-member plagioclase; Or, orthoclase; Qz, quartz. A, B, and C are approximate positions where, with falling temperature and crystallization of plagioclase, granitic melts with low, medium, and high potassium contents would intersect surfaces bounding plagioclase liquidus phase volume. Dashed lines are projected traces of these melts on base of tetrahedron. Arrows show subsequent paths of melt on bounding surfaces. M, temperature minimum.

In a melt whose composition plots within the quartz liquidus phase volume, quartz is the first mineral to crystallize with falling temperature, whereas alkali feldspar is the first mineral to crystallize in a melt whose composition plots within the alkali feldspar volume, and plagioclase is the first mineral to crystallize in a melt whose composition plots within the plagioclase volume. The norms of all Sierran granitoids plot within the plagioclase liquidus phase volume or on its internal phase-boundary surfaces. Therefore, if these rocks solidified from melts or partial melts, only the paths of crystallization within the plagioclase liquidus phase volume and on its bounding surfaces need be considered.

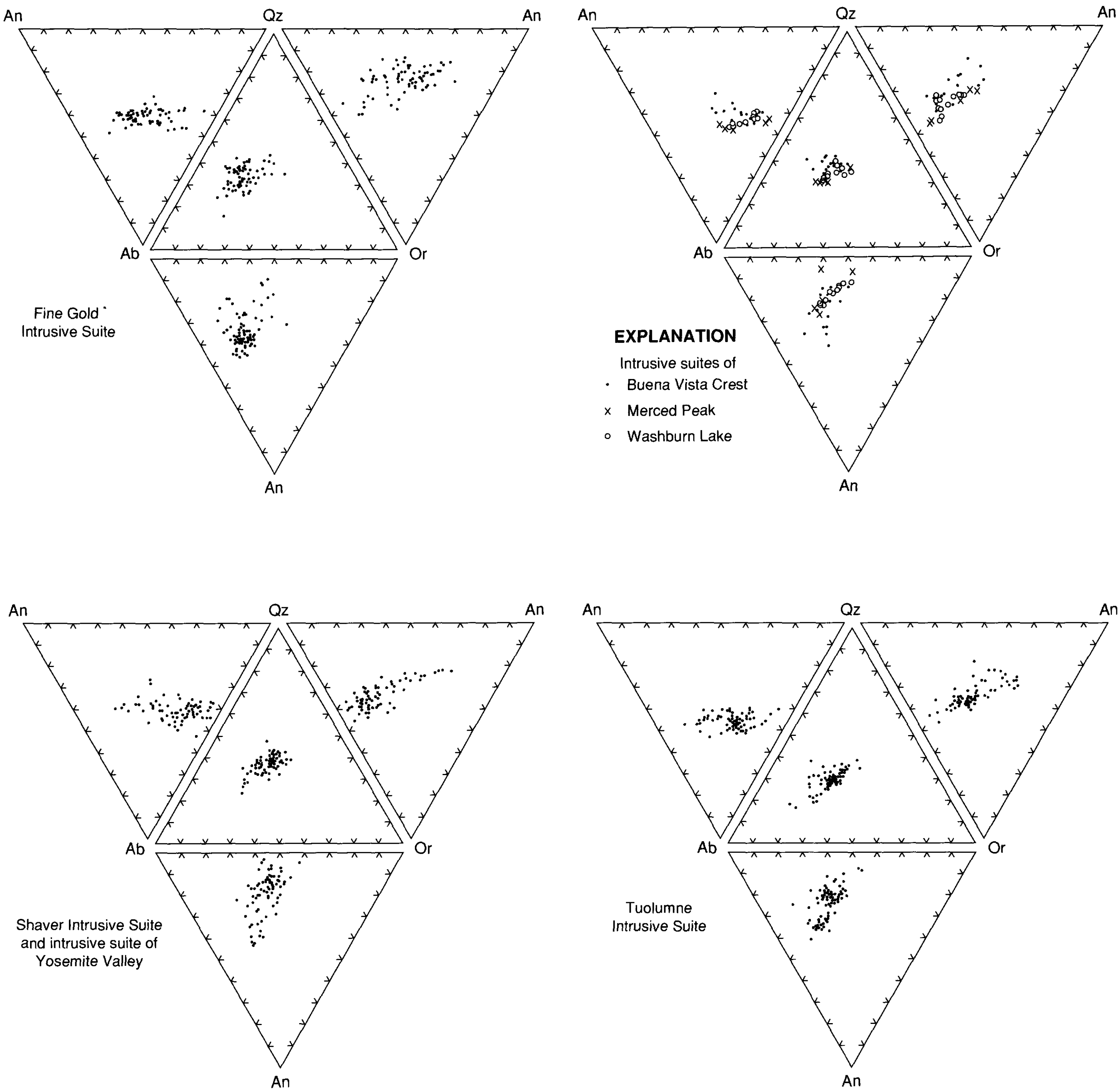


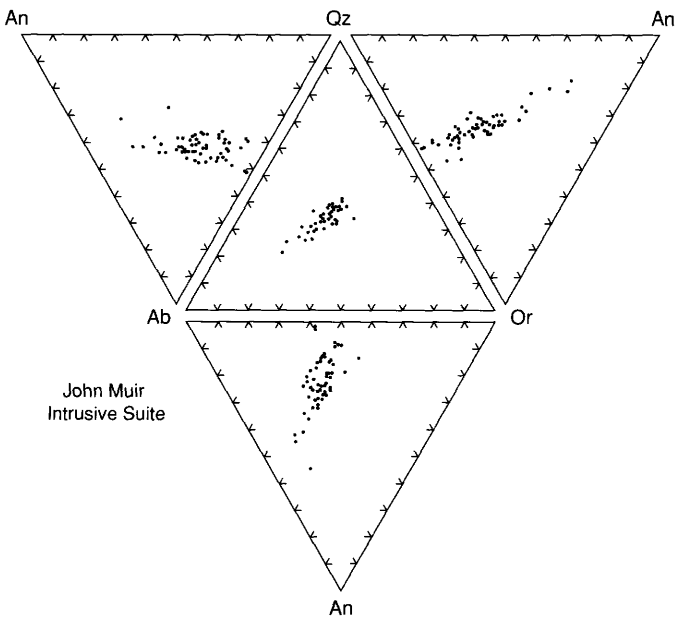
FIGURE 27.—Projections of CIPW norms of intrusive suites on faces of salic tetrahedron. Ab and An, albite and anorthite end-member plagioclase; Or, orthoclase; Qz, quartz.

Plots of the normalized compositions of samples from several of the better studied and more representative intrusive suites are given in figure 27. The true positions of the norms within the salic tetrahedron are within the plagioclase liquidus phase volume, although their projections on the faces of the tetrahedron may suggest otherwise. Their position on each external face is the position in which they would appear if the eye viewed the samples in their proper positions within the tetrahedron from the corner opposite the face.

The plots on the Qz-Ab-Or face show that most of the intrusive suites form coherent plots elongate in the general direction of Ab, and most of the plots on the Qz-An-Or face show that the plots are elongate toward An. Together, these plots show that each suite lies close to a planar surface that radiates from a line on the Ab-Or-An face, close to the An-Ab edge. If the samples had been composed entirely of the four constituents represented in the tetrahedron, the surfaces would have radiated exactly from the An-Ab edge.

The distribution of samples within the planes is shown by the projections on the Ab-An-Q and the Ab-An-Or faces. The distribution within suites from the western Sierra Nevada, which contain little potassium, is shown best on the Ab-An-Q face, whereas the potassium-rich rocks from the White Mountains are best represented on the Ab-An-Or face. In fact, the planes that contain the plots fan out across the tetrahedron, the low-potassium Fine Gold Intrusive Suite lying closest to the Ab-An-Q face and the high-potassium White Mountains rocks lying closest to the Ab-An-Or face.

To show the compositional differences among the suites, normative plots for the Fine Gold and John Muir Intrusive Suites in the Sierra Nevada and the Soldier Pass Intrusive Suite and other Jurassic intrusive rocks of the White Mountains are shown on a



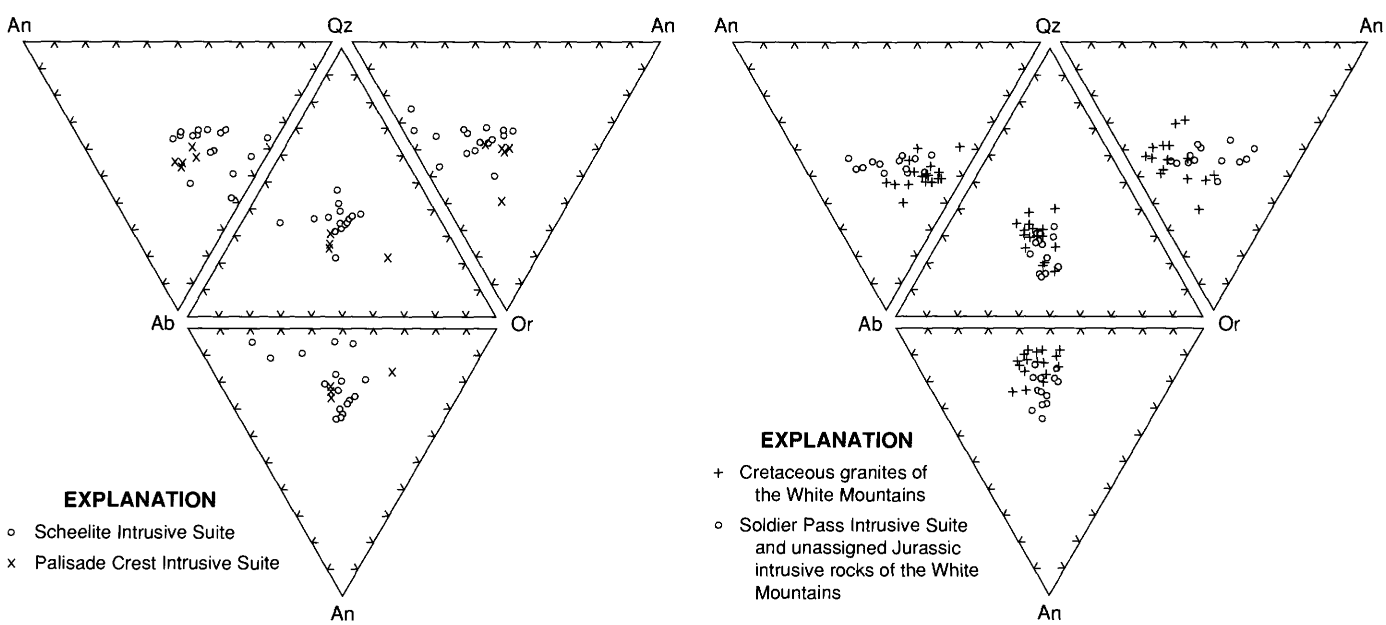


FIGURE 27.—Continued

diagram (fig. 28) in which Ab and An are combined. In effect, the tetrahedron is collapsed with a hinge on the Qz-Or sideline. The fanning out of suites across the tetrahedron reflects a regional increase in K_2O eastward across the Sierra Nevada. Obviously, compositional differences within suites are distinct from those between suites.

The path of crystallization for a suite can be traced within the salic tetrahedron. To illustrate, crystallization within three types of granitic suites—the low-potassium Fine Gold Intrusive Suite in the western foothills, the medium-potassium John Muir and Tuolumne Intrusive Suites farther east, and the high-potassium Soldier Pass Intrusive Suite in the White and northern Inyo Mountains—is discussed in the following three sections.

LOW-POTASSIUM FINE GOLD INTRUSIVE SUITE

A granitoid parent magma such as that which crystallized to form the low-potassium Fine Gold Intrusive Suite plots within the plagioclase liquidus phase volume close to the Ab-An-Qz face. As temperature decreases, the liquidus is reached, and plagioclase richer in An than the magma begins to crystallize, causing the composition of the melt phase to be displaced away from the Ab-An edge of the tetrahedron. As temperature continues to decrease, plagioclase that contains progressively smaller proportions of An crystallizes, causing the melt to follow a curved path within the tetrahedron until it intersects the quartz-saturation surface (at A in fig. 26). If the melt contains no K_2O, crystallization of plagioclase and quartz in constant proportions and at a constant temperature continues until all the melt has solidified. If, however, the melt contains even a little Or, it moves across the quartz-saturation surface as quartz and plagioclase crystallize, becoming progressively enriched in Or until the cotectic is reached; at this stage, quartz, plagioclase, and alkali feldspar crystallize together as the melt follows the cotectic toward the temperature minimum until the melt is totally depleted. Because of its bulk composition, most of the crystallization in a suite such as the Fine Gold Intrusive Suite takes place while the melt phase is moving toward and then across the quartz-saturation surface; only a small amount of crystallization occurs after it reaches the cotectic. Consequently, the sparse alkali feldspar is interstitial to the other minerals.

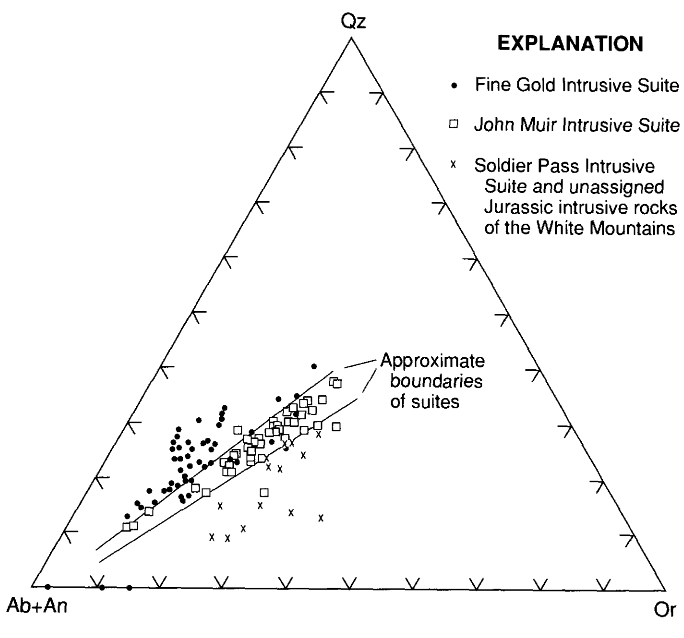


FIGURE 28.—CIPW norms of representative intrusive suites on diagram representing salic tetrahedron collapsed on Qz-Or hingeline so that Ab and An coincide. Ab and An, albite and anorthite end-member plagioclase; Or, orthoclase; Qz, quartz.

MEDIUM-POTASSIUM JOHN MUIR AND TUOLUMNE INTRUSIVE SUITES

A melt with medium amounts of potassium such as the John Muir or Tuolumne Intrusive Suites (fig. 28) will also begin to crystallize plagioclase first, then quartz, and finally alkali feldspar. However, crystallization of plagioclase causes the melt to intersect the quartz-saturation surface close to the cotectic and to move onto the cotectic after only a small temperature decrease (at B in fig. 26). Consequently, alkali feldspar begins to crystallize shortly after quartz and forms equant grains. In the Tuolumne Intrusive Suite, textural relations (Bateman and Chappell, 1979) indicate that quartz began to crystallize when the composition of the crystallizing plagioclase was An_{35-32} and that alkali feldspar began to crystallize only slightly later when the composition of the crystallizing plagioclase was An_{32-29}. If this suite had been a little richer in Or, the cotectic would have been intersected directly and quartz and alkali feldspar would have begun to crystallize simultaneously.

HIGH-POTASSIUM SOLDIER PASS INTRUSIVE SUITE

In an alkali-calcic suite such as the Soldier Pass Intrusive Suite, as in low-potassium and medium potassium suites, the first mineral to crystallize is plagioclase, but, because of its bulk composition, the melt moves toward the two-feldspar surface in the salic tetrahedron, where plagioclase and alkali feldspar crystallize together (see C in fig. 26). If the melt contains no quartz, plagioclase and alkali feldspar will crystallize together in a fixed ratio until crystallization is complete. However, if the melt contains even a small amount of quartz, as the temperature decreases it will move across the two-feldspar surface to the cotectic, where plagioclase, alkali feldspar, and quartz will crystallize together until crystallization is complete.

The final product of crystallization will have the same composition as the initial melt—regardless of the composition of the melt—if there is no separation of crystals. However, crystal separation or accumulation could explain the complete range of compositions found in the Sierra Nevada batholith. There is a general agreement between the succession of rock compositions found in the different intrusive suites and those expected to result from partial separation of crystals from the melt during crystallization; this is one of the strongest arguments supporting fractional crystallization as one of the principal mechanisms responsible for the lithologic variety within suites. In contrast, compositional differences between suites are caused by different compositions of the parent magmas.

INTERPRETATION OF MODAL PLOTS

The salic tetrahedron provides a basis for interpreting plots of modal data on the standard Q-A-P triangular diagram used in the classification of the granitoids (fig. 6; Streckeisen, 1973). As with the plots of norms in the salic tetrahedron, plots of modes do not take the mafic minerals into account. Modal plots (fig. 29) of the Fine Gold, Tuolumne, and Soldier Pass Intrusive Suites illustrate the changing compositional patterns across the batholith. The modal plots show wide scatter that is partly the result of analytical inaccuracy but that also must reflect such processes as melt segregation and crystal accumulation, turbulent convective mixing, surges of new magma, and other processes that would be expected to occur in a magma chamber. Nevertheless, most plots are in general agreement with patterns expectable from crystal fractionation. Linear plots also can result from the mixing of mafic with felsic magmas. Isotopic data are needed to determine which process dominated.

FINE GOLD INTRUSIVE SUITE

The modal plot for the Fine Gold Intrusive Suite is curved (fig. 29). The greatest concentration of modes is in the tonalite field close to the Q-P sideline, between 25 and 35 percent Q. Plotted modes extend from this concentration along and close to the Q-P sideline toward P and into the interior of the diagram, across the granodiorite field, and into the granite field. The modes that plot in the tonalite field represent rocks that plot chiefly on the quartz-saturation surface in the salic tetrahedron (fig. 26). The modes that extend from the area of greatest concentration along and close to the Q-P sideline toward P generally represent rocks that have high color indices and plagioclase with high maximum and minimum An contents. The abundant mafic minerals and plagioclase in these rocks probably crystallized within the plagioclase liquidus phase volume of the salic tetrahedron before the quartz-saturation surface was reached (fig. 26). The modes that plot in the granodiorite and granite fields represent rocks that have relatively low color indices for the suite and higher-than-normal contents of alkali feldspar and represent substantial crystallization on the cotectic.

TUOLUMNE INTRUSIVE SUITE

Most of the modes for the Tuolumne Intrusive Suite plot in the granite and granodiorite fields closer to A than the modes for the Fine Gold Intrusive Suite and extend in a curved path toward the Q-P sideline in the quartz diorite field (fig. 29). The slight curve in the trend is consistent with crystal fractionation; however, a small but regular inward increase of initial $^{87}Sr/^{86}Sr$ values (fig. 36) for samples collected near the margins of the suite and that plot close to the Q-P sideline in the tonalite, quartz diorite, and adjacent parts of the quartz monzodiorite and granodiorite fields indicate the mixing of source materials (see section "Tuolumne Intrusive Suite" under "Patterns Within Intrusive Units"). The trend in figure 29 indicates that plagioclase was first to crystallize among the felsic minerals, followed closely by quartz and shortly thereafter by alkali feldspar. A linear trend directly away from P would indicate either mixing of end-member magmas or simultaneous crystallization of quartz and alkali feldspar, after crystallization of plagioclase. The abundance of modes in the granodiorite field and the adjacent part of the granite field indicates that much of the crystallization was on the cotectic.

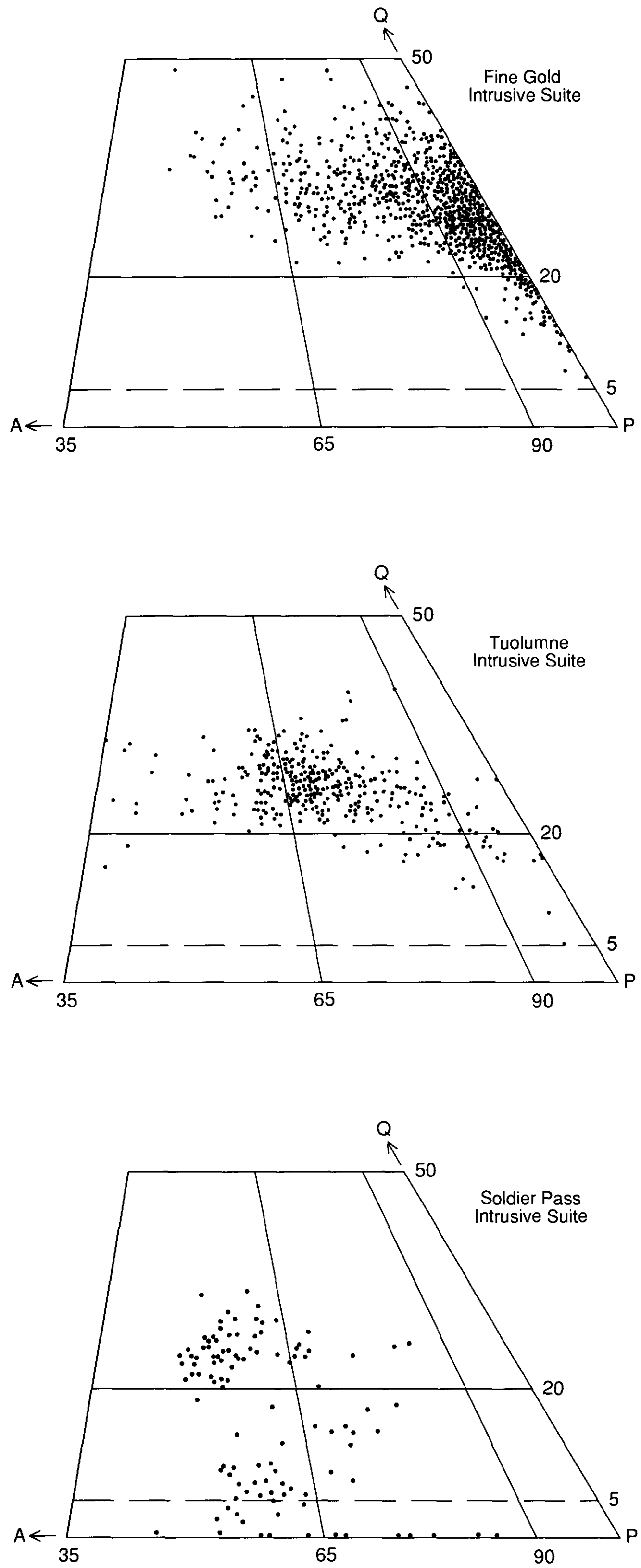


FIGURE 29.—Modes of representative calcic, calc-alkalic, and alkali-calcic intrusive suites on Q-A-P diagram. See figure 6 for Q-A-P classification diagram.

SOLDIER PASS INTRUSIVE SUITE

The modal plot for the Soldier Pass Intrusive Suite is a somewhat-distorted mirror image of the Fine Gold Intrusive Suite (fig. 29). Modes extend along the A-P sideline within the monzodiorite and monzonite fields, then leave the sideline in the monzonite field and move into the quartz monzonite, quartz monzodiorite, granodiorite, and granite fields, becoming more felsic. Modes for the granodiorite of Beer Creek are widely scattered in the granite, granodiorite, quartz monzonite, and quartz monzodiorite fields (see fig. 88). They complicate an otherwise simple pattern of crystal fractionation. Probably they reflect the introduction of extraneous materials into the parent magma. The modes that plot along and close to the A-P sideline represent minerals that crystallized early; they would plot on the two-feldspar surface within the salic tetrahedron. Modes that plot in the quartz monzodiorite, quartz monzonite, granodiorite, and granite fields represent later crystallization; they would plot along the cotectic within the salic tetrahedron.

MAGMA MIXING

Although crystal fractionation is highly effective in producing compositional variations within a suite of rocks, mixing of two or more magmas or contamination of a magma with crustal material can also cause compositional variations within an intrusive suite. Mixing probably is a dominant process early in the history of a magmatic system, whereas fractionation is most effective later, during crystallization. Almost certainly, mantle-derived basalt is the principal carrier of heat into the crust, and if it continues to rise into the crust it undoubtedly assimilates crustal material and mixes and mingles with anatectic magmas that it generates within the crust. Even if the basaltic magma ponds at the base of the crust and acts only as a carrier of heat, it can produce anatectic magmas in the overlying rocks, which can mix with one another and assimilate other crustal materials.

Physical evidence of magma mixing at the present level of exposures in the Sierra Nevada is scarce, but mixtures of more mafic with less mafic magmas can be seen along the Tioga Pass Road about 2 km west of Tenaya Lake in the equigranular facies of the Half Dome Granodiorite. If mafic inclusions, almost ubiquitous in tonalite and granodiorite, are blobs of mantle-derived basalt, as seems likely, their mere presence indicates magma mixing.

At this time the most convincing evidence of magma mixing is found in radioactive isotopic ratios because they can be mixed but not unmixed. An initial $^{87}Sr/^{86}Sr$ of 0.708 probably indicates a larger proportion of crustal material in the parent magma of an intrusion or intrusive suite than a ratio of 0.704. Constant initial $^{87}Sr/^{86}Sr$ throughout an intrusion indicates that unless the mixed components had the same isotopic ratios, any mixing that may have taken place earlier terminated before emplacement, and that any compositional variations within the intrusion are the result of differentiation. On the other hand, variations in initial $^{87}Sr/^{86}Sr$ within an intrusion indicate incomplete homogenization or continued mixing. For example, a fairly regular increase of initial $^{87}Sr/^{86}Sr$ inward from the west margin of the Tuolumne Intrusive Suite from about 0.7057 to 0.7064 must reflect continued addition of crustal material to the magma.

Linear-variation diagrams or plots of oxides or elements against one another have been cited as evidence of magma mixing but should be used with caution. If more than two magmas were involved, nonlinear plots can result. Also, fractionation is largely a process of unmixing and under some circumstances can lead to linear plots.

REGIONAL AGE PATTERNS

The local order of intrusion has been established in many places by determining the relative ages of plutons in contact with one another. However, gaps in the chronology occur because many intrusions are not in contact with either the next older or next younger intrusion and prevent relating the different local intrusive successions to one another. Isotopic ages are required to relate them in a single chronology.

HISTORY OF ISOTOPIC DATING IN THE CENTRAL SIERRA NEVADA

The first radiometric age determinations for rocks from the central Sierra Nevada were for granitoids in the Bishop area using the nonisotopic lead-alpha method (Larsen and others, 1958). These determinations yielded Mesozoic ages, but the reported ages showed no relation to the sequence of intrusion as established from field relations. The next determinations were K-Ar ages on biotite from samples of granitoids collected in Yosemite Valley and eastward to the crest of the Sierra Nevada (Evernden and others, 1957; Curtis and others, 1958). These ages decreased eastward from 95.2 to 82.4 Ma in accordance with the relative ages of the rocks as established in the field by intrusive relations. Isotopic ages by other methods have since shown that the ages of all but the youngest granitoid were reduced as a result of reheating during the west-to-east emplacement of intrusions.

In the 1960's, the K-Ar ages on hornblende as well as biotite were determined for samples collected along a traverse across the Sierra Nevada south of Yosemite at about the latitude of Bishop (Kistler and others, 1965). The biotite and hornblende ages of samples of the Mount Givens, Lamarck, and Round Valley Peak Granodiorites from the central part of the traverse are concordant in the range of 90 to 80 Ma, indicating rapid cooling to below the blocking temperatures of both hornblende and biotite. East and west from this central segment, both the hornblende and the biotite ages increase, but hornblende ages increase more than the biotite ages, producing increasing discordance. The concordant ages for rocks from the central segment were interpreted to be crystallization ages and the discordant ages to be reduced from older crystallization ages as a result of reheating by younger plutons (Kistler and others, 1965). Because the blocking temperature of hornblende is higher than that of biotite, the hornblende ages were assumed to approach the true ages of crystallization more closely than the biotite ages. The ages of the Lamarck and Mount Givens Granodiorites, the Evolution Basin Alaskite, and the granite of Dinkey Dome were interpreted to be 90±10 Ma and the ages of most of the other rocks to be considerably older. K-Ar ages of granitoids in the White and Inyo Mountains (McKee and Nash, 1967; Crowder and others, 1973) and selected granitoids in the western foothills (Bateman and Wones, 1972b; Kistler, 1974; Bateman and Lockwood, 1976; Bateman and Sawka, 1981) were subsequently published.

Evernden and Kistler (1970) published a summary of all the K-Ar ages that had been determined for rocks from the Sierra Nevada, including many new ages for samples chiefly from north and south of the map area. These K-Ar ages show that within the map area (pl. 1) a dense belt of Cretaceous intrusions striking about N. 20° W. crosses a belt of scattered Jurassic plutons striking about N. 40° W. (fig. 30). Evernden and Kistler (1970) also proposed five intrusive epochs that occurred periodically over a timespan that extended from 215 to 81 Ma. Each epoch was postulated to have lasted 11 to 20 m.y. and to have been separated by nonmagmatic intervals of 11 to 15 m.y.

U-Pb isotopic ages of granitoids of the White Mountains were first published in the late 70's (Sylvester, Miller, and Nelson, 1978; Gillespie, 1979). In 1980, these ages were supplemented by U-Pb ages for granitoids in the western foothills south of the map area

(Saleeby and Sharp, 1980). Stern and others (1981) published 62 U-Pb ages for 48 different granitoid units within the map area, and shortly thereafter Chen and Moore (1982) published 132 additional U-Pb ages on zircon and 7 on titanite from 82 rock samples, collected mostly from the eastern Sierra Nevada south of the map area), but a few were from the eastern part of the map area. Most recently, Dodge and Calk (1986) published eight U-Pb ages for granitoids in the northwestern part of the map area and Hanson and others (1987) published a U-Pb age for the granite of McAfee Creek in the White Mountains. The U-Pb ages published prior to 1986 are summarized graphically in figure 31.

To date, relatively few U-Pb ages have been determined for the metavolcanic rocks, but a few have been published by Fiske and Tobisch (1978), Tobisch and others (1986), and Hanson and others (1987). A single U-Pb age for a sample of rhyolite tuff from the White Mountains is given in table 1.

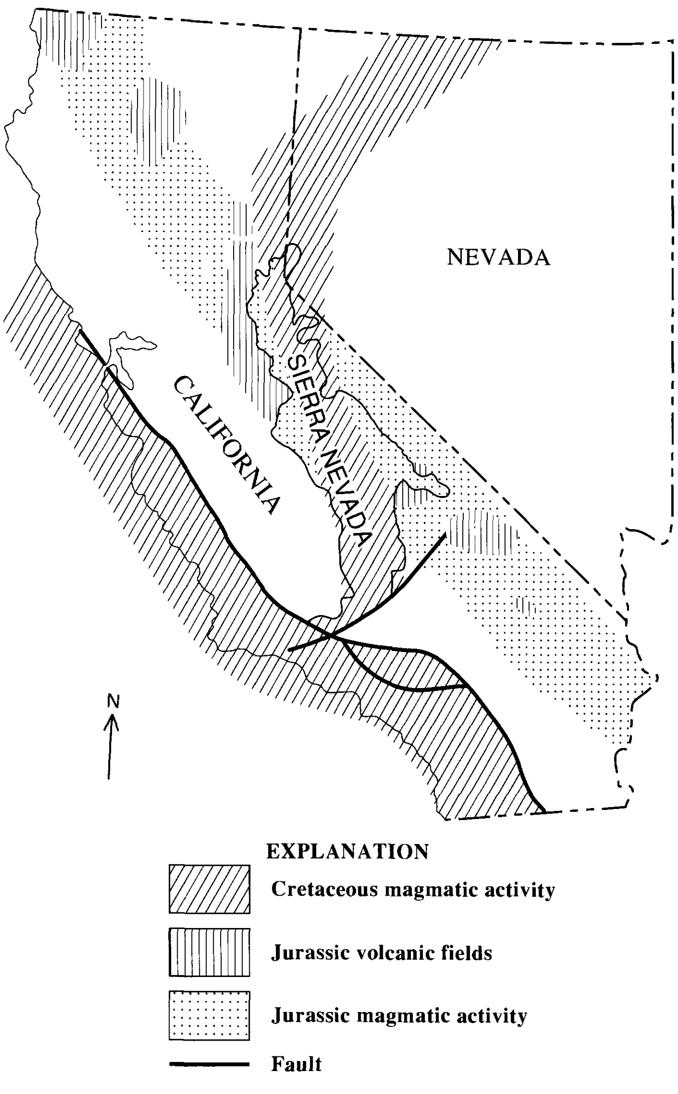


FIGURE 30.—Areas of Jurassic and Cretaceous plutonism in California and Nevada. Modified from Kistler and others (1971).

The U-Pb ages represent higher blocking temperatures than K-Ar ages on either biotite or hornblende and are less affected by thermal and tectonic disturbances. For several different intrusions, ages determined in different laboratories show good agreement, effectively indicating the intervals of plutonism and suggesting that the distribution of ages within the map area is representative of the entire Sierra Nevada batholith (fig. 31). Inherited zircon, a complication in many terranes, apparently is not a serious problem in the Sierra Nevada. The U-Pb ages generally confirm the gross distribution of ages drawn from K-Ar dating by Evernden and Kistler (1970) but cast serious doubt on the validity of their proposed five intrusive epochs of plutonism and consequently on interpretations premised on the reality and supposed periodicity of these epochs (Kistler and others, 1971; Shaw and others, 1971). The U-Pb ages for samples collected from within the map area indicate markedly increased plutonism from approximately 214 to 201 Ma (latest Triassic to earliest Jurassic), 172 to 148 Ma (late Middle to Late Jurassic), and 119 to 86 Ma (late Early to early Late Cretaceous). General agreement of the isotopic ages for rocks collected south of the map area with these periods of accelerated plutonism suggests that they apply over a large area.

AGES OF INTRUSIVE SUITES AND UNASSIGNED UNITS

Throughout this report, the Decade of North American Geology 1983 Geologic Time Scale (Palmer, 1983) is used in assigning isotopic ages to geologic periods and episodes. This time scale shows the boundary between Early and Late Cretaceous to be at 97.5 Ma; because some isotopic ages of both volcanic and plutonic rocks are close to this boundary, ages that are within the range 101 to 97 Ma are referred to in this report as "mid-Cretaceous."

The following observations and conclusions are based largely on the U-Pb ages, and the ages cited are 206Pb-238U ages unless otherwise indicated. K-Ar ages, especially hornblende ages, are used where U-Pb ages are dubious or lacking. All these ages are taken from the publications cited in the preceding sections. A few as yet unpublished Rb-Sr ages that were determined by R.W. Kistler (written communs., 1982–85) are also included.

The isotopic ages of the plutonic rocks indicate the following distribution patterns within the map area (pl. 2): (1) Triassic granitoids occur in the eastern

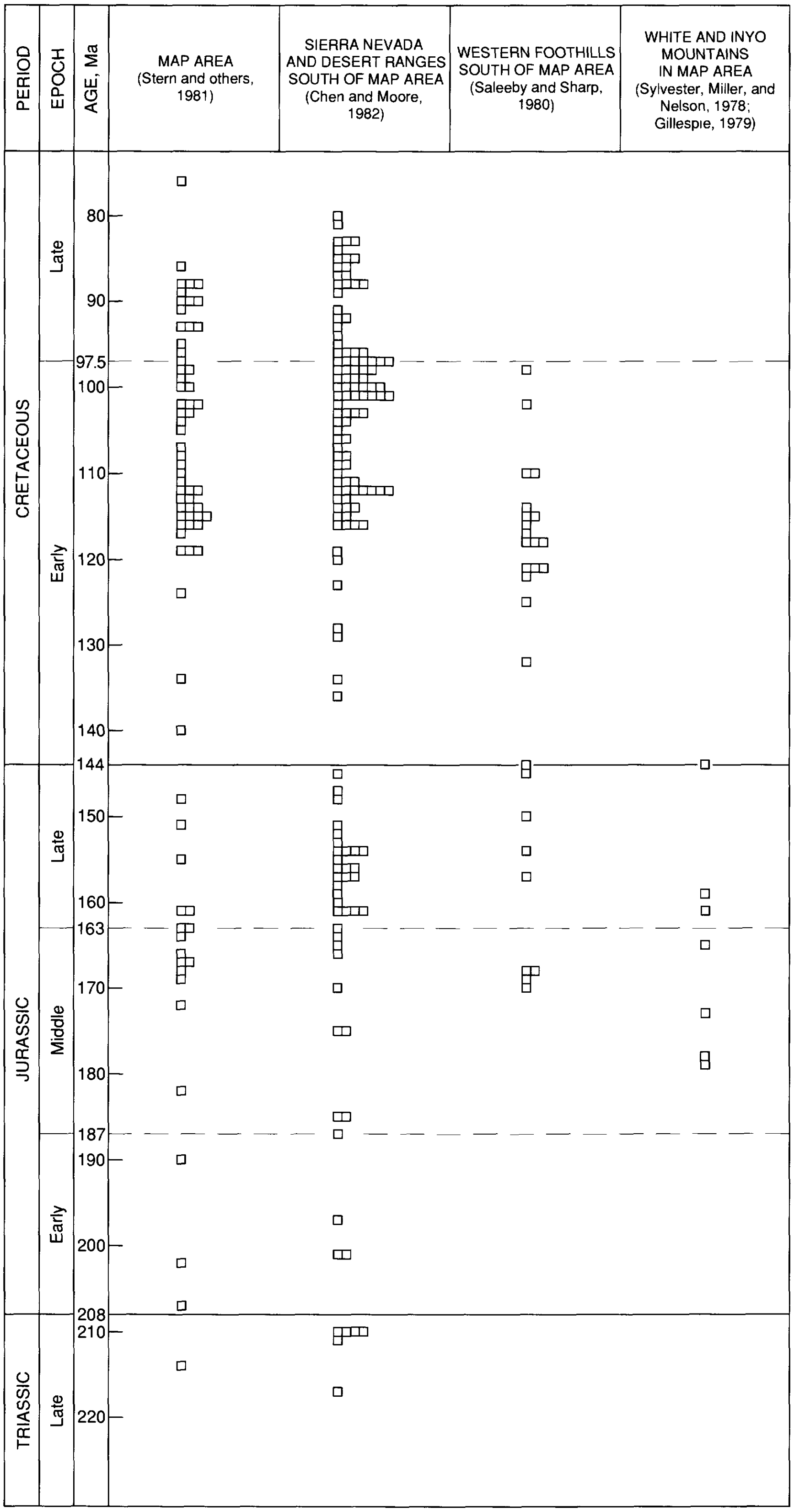


FIGURE 31.—Summary of U-Pb ages for Sierran granitoids published prior to 1986. Period and epoch boundaries from Palmer (1983). Map area referred to is Mariposa 1° by 2° quadrangle (pl. 1).

escarpment of the Sierra Nevada north of Big Pine and extend north and east into the Benton Range and other exposures of bedrock between the Sierra Nevada and the White Mountains but do not extend into the White and Inyo Mountains. (2) Jurassic plutonic rocks are present in the eastern escarpment of the Sierra south of Big Pine and in the White and Inyo Mountains, are sparsely represented in the interior of the batholith, and are present locally in the western foothills of the Sierra Nevada. (3) Cretaceous granitoids occupy most of the Sierra Nevada west of the range crest, where they become progressively younger eastward, and are also present locally in the eastern Sierra Nevada and in the northern White Mountains. The age groups form parallel belts that extend to the north and south beyond the limits of the map area, probably for the full length of the batholith. Nevertheless, Jurassic granitoids are present on both sides of the batholith, reflecting regional-scale crossing of the Jurassic by the Cretaceous granitoids (fig. 30). The following is a more detailed description of the distribution and ages of the granitic rocks.

TRIASSIC INTRUSIVE ROCKS

The Triassic rocks compose the Scheelite Intrusive Suite, which consists of the Wheeler Crest Granodiorite, the granite of Lee Vining Canyon, and the Tungsten Hills Granite. The U-Pb age determinations for these rocks range from 217 to 123 Ma, but all determinations of less than 190 Ma were for rocks that also yielded ages older than 190 Ma for other samples or for other fractions of the same sample. The Wheeler Crest Granodiorite and the granite of Lee Vining Canyon yielded U-Pb ages that are consistent with a true age of about 214 Ma, whereas two U-Pb ages for the Tungsten Hills Granite (202 and 198 Ma) are significantly younger (Stern and others, 1981; Chen and Moore, 1982). Nevertheless, this granite is included in the suite because of its close association with the Wheeler Crest Granodiorite and because some of the isotopic ages for both the Wheeler Crest Granodiorite and the granite of Lee Vining Canyon are younger than 200 Ma.

JURASSIC INTRUSIVE ROCKS

The Jurassic intrusive rocks on the west side of the batholith include the tonalite of Granite Creek, with a U-Pb age of about 163 Ma, the granite of Woods Ridge, with a U-Pb age of 151 Ma, and six plutons west of the map area. Two of the six plutons west of the map area yielded ages very close to that of the tonalite of Granite Creek and were included in the Jawbone granitoid sequence by Stern and others (1981). The other four plutons are of two age groups. Two small plutons west of the Melones fault zone and also west of the map area have U-Pb ages of 190 and 182 Ma and intrude the Peñon Blanco Volcanics, to which they may be related; the other two plutons range in age from 151 to 140 Ma and belong to a group of intrusions in the western foothills, including the granite of Woods Ridge, which postdate the Nevadan orogeny. U-Pb ages for rocks south of the map area and K-Ar hornblende ages for rocks north of the map area indicate that most of these foothills plutons were emplaced within the timespan of about 154 to 134 Ma but that some may be as young as 120 Ma and transitional to the voluminous Cretaceous intrusive suites (Evernden and Kistler, 1970; Saleeby and Sharp, 1980).

The isotopically dated Jurassic intrusive rocks in the eastern part of the batholith include the Palisade Crest Intrusive Suite (chiefly in the eastern escarpment of the Sierra Nevada but also in Owens Valley and the Inyo Mountains) with Rb-Sr isotopic ages of about 170 Ma; the Soldier Pass Intrusive Suite in the White and northern Inyo Mountains, also with an average isotopic age of about 170 Ma; the unassigned quartz monzonite of Mount Barcroft in the central White Mountains with U-Pb ages of 165 and 161 Ma; the granite of Sage Hen Flat in the southern White Mountains with a U-Pb age of 144 Ma; the leucogranite of Casa Diablo Mountain in the Benton Range with a U-Pb age of 161 Ma; and quartz diorite from the Pine Creek tungsten mine in the eastern Sierra Nevada with a U-Pb age of 169 Ma. The granodiorite of Cabin Creek (in the White Mountains) is undated, but it occupies the same structure as the quartz monzonite of Mount Barcroft and probably is of about the same age.

In the Goddard septum, a mass of sheared, fine-grained quartz syenite that has a reported U-Pb age of 157 Ma (Chen and Moore, 1982) may be cogenetic with adjacent metavolcanic rocks. This and other as yet undated intrusions within the septum are considered to be Jurassic or older because they are cut by mafic dikes thought to belong to the Independence dike swarm, which were determined by Chen and Moore (1982) to have a U-Pb age of 148 Ma.

CRETACEOUS INTRUSIVE ROCKS

The Cretaceous intrusive rocks range in age from the beginning of the Cretaceous to early Late Cretaceous. The oldest isotopically dated Cretaceous intrusions are scattered, relatively small plutons in the western foothills; they were emplaced beginning in latest Jurassic time after the Nevadan orogeny (see section "Jurassic Intrusive Rocks"). Cretaceous magmatism accounts for the great bulk of the batholith; it began with intrusion of the Fine Gold Intrusive Suite in the western part of the batholith, adjacent to the western metamorphic belt. This suite has an average age of about 114 Ma but may have been emplaced over a timespan of several million years. It was succeeded eastward by the Shaver Intrusive Suite and the coeval intrusive suite of Yosemite Valley (both ~103 Ma); the approximately coeval intrusive suites of Buena Vista Crest, Merced Peak, and Washburn Lake (all ~98 Ma); and the approximately coeval Tuolumne and John Muir Intrusive Suites (both ~90 Ma). The ages of the unassigned leucogranites of Graveyard Peak (99 Ma) and of Rawson Creek (95 Ma) in the eastern Sierra Nevada fit fairly well into this pattern of eastward-decreasing ages.

In the White Mountains, a U-Pb age of 100 Ma has been reported for the granite of McAfee Creek (Hanson and others, 1987), and the granite of Pellisier Flats has a discordant U-Pb age of 90 Ma and K-Ar ages of 161 Ma on biotite and of 100 and 92 Ma on hornblende. The ages of the other Cretaceous granites, indicated only by K-Ar determinations on biotite, range from 87 to 74 Ma (Crowder and others, 1973).

RATE OF EASTWARD MIGRATION OF PLUTONISM DURING THE CRETACEOUS

Chen and Moore (1982) calculated the rate of eastward migration of plutonism during the Cretaceous along two traverses. One crosses the map area and is based chiefly on the U-Pb ages published by Stern and others (1981), and the other is farther south (between lats 36° and 37° N.) and is based on their own U-Pb age data. They determined the rate of migration along both traverses to be 2.7 mm/yr (2.7 km/m.y.). The positions and ages of intermediate-composition suites indicate that plutonism was discontinuous and episodic, even though the rate of migration was fairly constant.

DURATION OF MAGMATISM IN THE FORMATION OF INTRUSIVE SUITES

The close grouping of the U-Pb ages for individual intrusive suites and intrusive relations with other suites suggest that the main phase of plutonism associated with a single magmatic episode persisted for no more than a few million years, although minor activity may have continued for longer periods. The Fine Gold Intrusive Suite has yielded ages ranging from 123 to 105 Ma, but all the concordant ages except one age of 105 Ma range from 116 to 112 Ma. Seven U-Pb ages for the John Muir Intrusive Suite range from 93 to 88 Ma, and three for the Tuolumne Intrusive Suite range from 91 to 86 Ma (Stern and others, 1981).

Chen and Moore (1982) addressed this problem by determining the U-Pb ages on 19 zircon and 2 titanite size fractions of 11 samples from a nested and compositionally zoned suite of rocks—the Sequoia Intrusive Suite of Moore and Sisson (1987)—in Sequoia National Park. The four units that compose this suite range from granodiorite in the margins to leucogranite in the core. Most of the U-Pb ages are concordant within 0.6 m.y. and range from 102.3 to 96.3 Ma. They concluded that the entire suite was emplaced in a relatively short timespan (1 to 2 m.y.) and that late-stage magmatic activity continued for about 3 m.y. while the suite cooled inward from the margins.

COMPOSITIONAL PATTERNS

PATTERNS WITHIN INTRUSIVE UNITS

The concept underlying the designation "intrusive suite" is that all the units in a particular suite have common characteristics and are in some manner cogenetic and originated during the same magmatic episode. The most firmly established suites are those that have distinctive modal, chemical, isotopic, and textural characteristics and in which the different units that make up the suite are transitional to one another. Such suites are compositionally zoned, and some exhibit partially or completely nested patterns in which more mafic rock in the margins gives way inward to younger, more felsic rock in the interior. The Tuolumne Intrusive Suite, originally called the Tuolumne Intrusive Series (Calkins, 1930), is a splendid example of this kind of intrusive suite.

However, some of the intrusive suites shown on plate 1 are composed of rocks that occur in the same area, were emplaced at about the same time, and generally were emplaced in order from mafic to felsic but which have few common characteristics and lack compositional transitions between units. In the following discussion, the compositional and textural variations within several suites and lithodemes, especially ones that show transitions and common characteristics, are reviewed to provide a background for developing a viable explanation for the diversity of rock compositions within intrusive suites.

TUOLUMNE INTRUSIVE SUITE

The Tuolumne Intrusive Suite consists of a group of nested units that are progressively younger and more leucocratic inward (fig. 32). At the outer margins are the granodiorites of Kuna Crest and Grayling Lake and the tonalites of Glen Aulin and Glacier Point. Probably these units were emplaced as a single mass and disrupted during intrusion of the interior units, but it is possible that more than one mass existed. Successively interior to these rocks are the equigranular facies of the Half Dome Granodiorite, the megacrystic facies of the Half Dome Granodiorite, the Cathedral Peak Granodiorite, and the Johnson Granite Porphyry. Contacts between these units are sharp except for the contact between the equigranular and megacrystic facies of the Half Dome Granodiorite, which is gradational everywhere except for a short segment at the north end of the west contact. Compositional changes occur both gradually across units and abruptly at contacts. The different units were emplaced as magmatic surges (fig. 33). In places where a younger surge rose alongside the next older surge without intruding it deeply, compositional and textural changes are largely within the units, but where a surge cut across or penetrated deeply into earlier surges, abrupt compositional and textural changes also occur at the contact.

Bateman and Chappell (1979) made a detailed study of samples collected along a traverse that crosses the Tuolumne Intrusive Suite (traverse A–B, fig. 32) and concluded that the compositional zoning is the result of crystal fractionation. They suggested that crystals with which the magma was saturated accreted at the margins of the inward cooling and solidifying magma and (or) settled from the hotter and more fluid interior magma. Since this study was made, Kistler and others (1986) published isotopic data on these same samples, which require that source materials with different isotopic properties mixed to produce inward increase of initial $^{87}Sr/^{86}Sr$ in the outer part of the suite and a small additional increase in the innermost unit, the Johnson Granite Porphyry. They also inferred from the relations between isotopes, major elements, and minor elements that crystal fractionation accounts for part of the compositional variations.

The traverse of Bateman and Chappell's (1979) follows California State Highway 120 except at the west end where a 5-km segment runs cross-country to the suite's external contact on the west side, north of May Lake (fig. 32). Where this traverse crosses contacts, compositional and textural differences between the two sides are small. The largest changes take place gradually across units rather than abruptly at contacts. A major exception is the abrupt contact of the Johnson Granite Porphyry with the Cathedral Peak Granodiorite.

The following discussion largely pertains to the west half of the traverse where exposures are virtually continuous; limited exposures show that the east half is essentially a mirror image of the west half. Compositional changes are greatest in the outer 1 km, whereas textural changes occur throughout the suite. Dark, fine-grained, strongly foliated rock in the outer part of the tonalite of Glen Aulin gives way inward to progressively lighter colored, coarser grained, more weakly foliated rock in the Half Dome and Cathedral Peak Granodiorites. Changes in modal compositions, specific gravity, major elements, and initial $^{87}Sr/^{86}Sr$ of the rocks along this traverse are shown graphically in figures 34 through 36.

Quartz and alkali feldspar, sparse in the margins of the suite, increase in abundance and grain size inward well into the equigranular facies of the Half Dome Granodiorite, then remain constant across the megacrystic facies of the Half Dome and the Cathedral Peak Granodiorites (fig. 34). Corresponding to these variations are decreases in specific gravity and in the abundance of plagioclase, biotite, and hornblende (some with clinopyroxene cores) into the equigranular facies of the Half Dome Granodiorite. Farther inward, hornblende decreases more moderately and is absent in the inner part of the Cathedral Peak Granodiorite and in the Johnson Granite Porphyry, whereas biotite remains fairly constant inward to the Johnson Granite Porphyry, where it decreases. Specific gravity decreases sharply into the outer part of the equigranular facies of the Half Dome Granodiorite, then more gradually farther inward.

Of the major oxides, SiO_2 and K_2O increase sharply, and Al_2O_3, Fe_2O_3, FeO, MgO, and CaO decrease sharply inward from the margins into the equigranular facies of the Half Dome Granodiorite (fig. 35). Farther inward, SiO_2 remains fairly constant to the Johnson Granite Porphyry, where it rises slightly, and Fe_2O_3, FeO, MgO, and CaO decrease gradually by small amounts. Na_2O increases inward into the inner part of the Cathedral Peak

Granodiorite, then decreases into the Johnson Granite Porphyry.

Plagioclase crystals are zoned, and their maximum, minimum, and average anorthite contents decrease

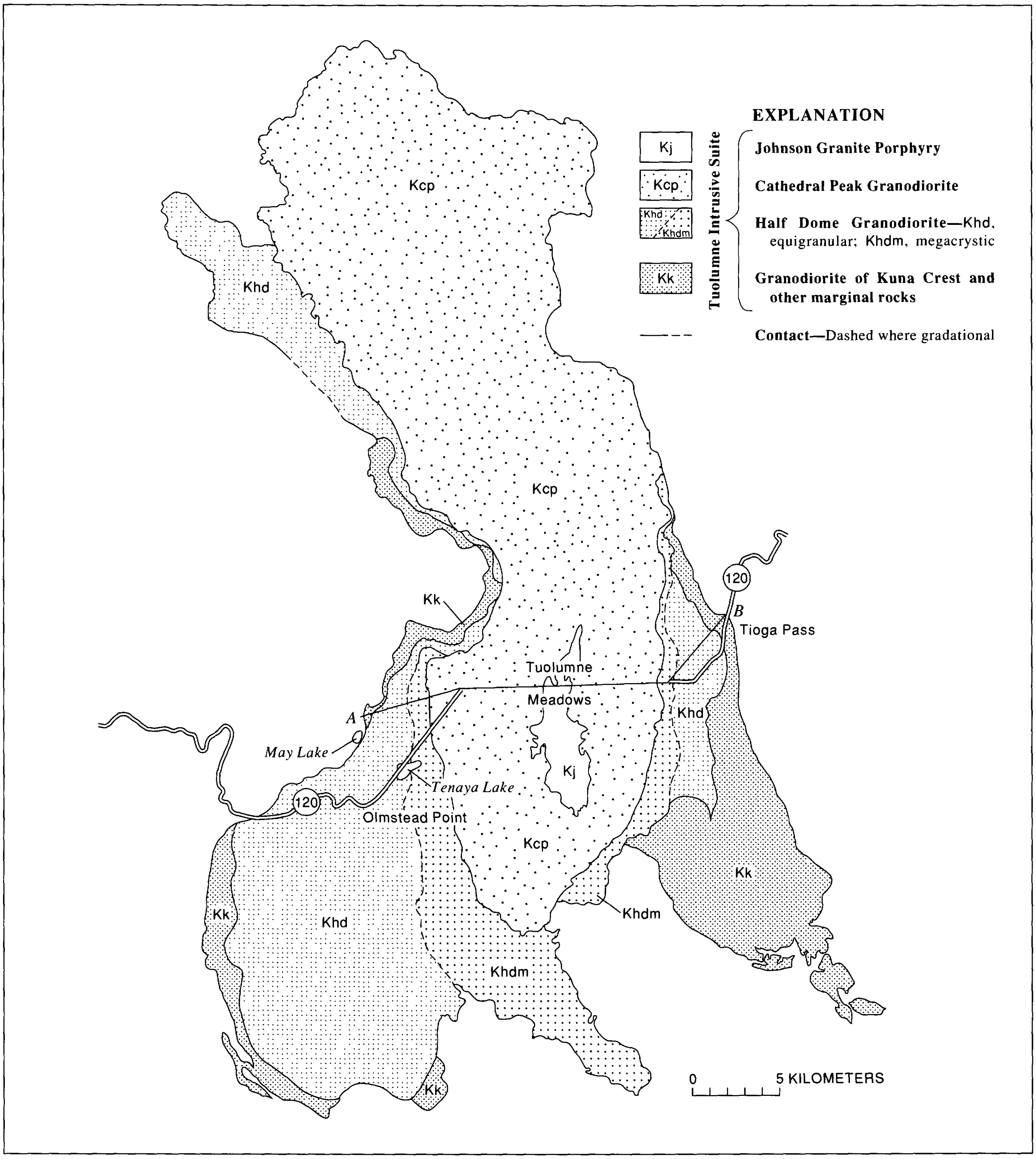


FIGURE 32.—Generalized geology of the Tuolumne Intrusive Suite and location of traverse A–B referred to in figures 34 through 36. Modified from Bateman and Chappell (1979).

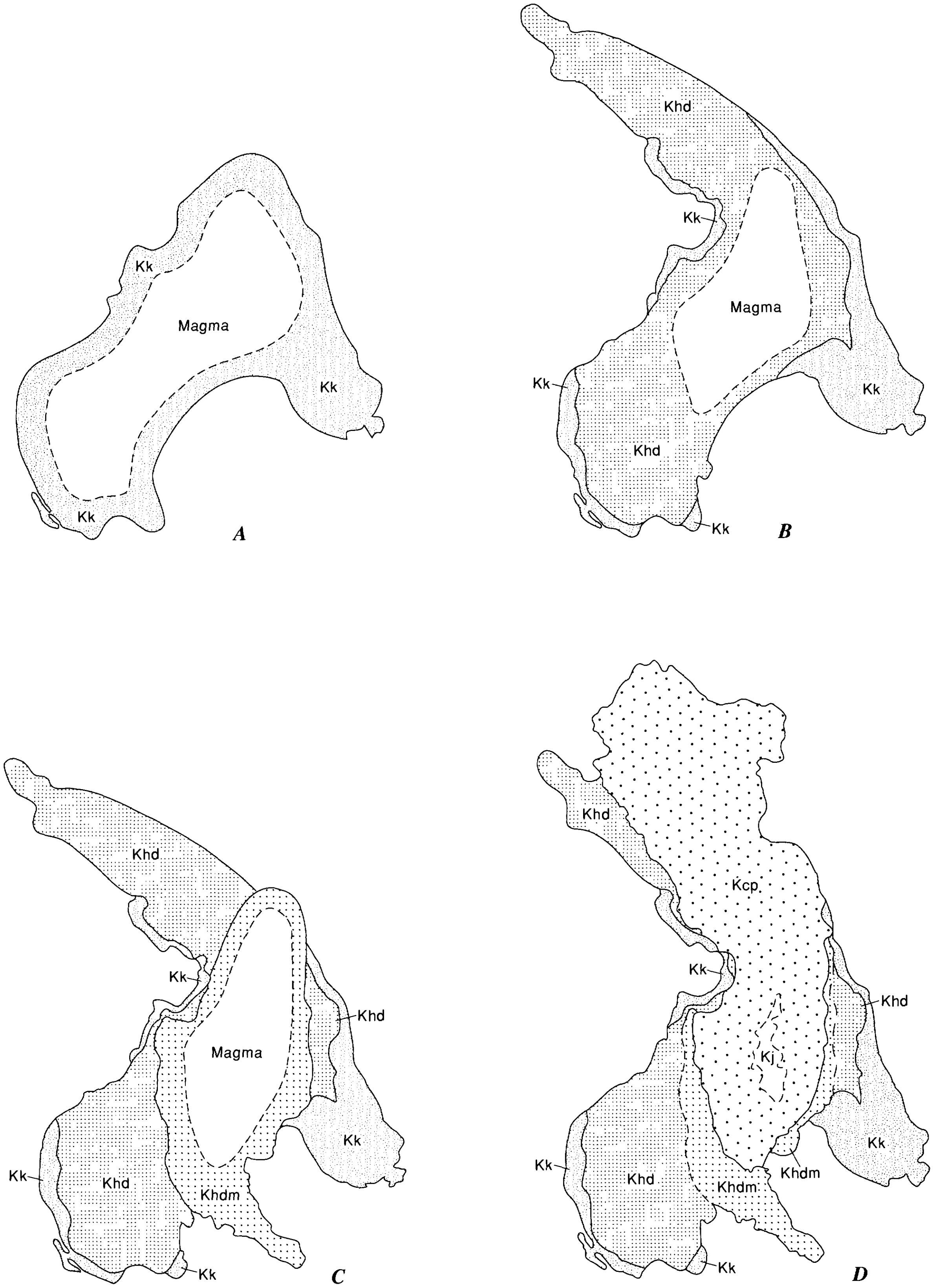


FIGURE 33.—Stages in emplacement history of the Tuolumne Intrusive Suite (Bateman and Chappell, 1979). *A*, Initial intrusion and solidification of granodiorite of Kuna Crest and other peripheral granitoids (Kk). *B*, New surge of magma followed by solidification of equigranular facies of the Half Dome Granodiorite (Khd). *C*, Second surge of magma followed by solidification of megacrystic facies of the Half Dome Granodiorite (Khdm). *D*, Third surge of magma followed by solidification of the Cathedral Peak Granodiorite (Kcp). Eruption and solidification of the Johnson Granite Porphyry (Kj) was final stage of emplacement. Dashed lines are gradational contacts.

gradually inward from the margins to the core of the suite (fig. 34). These progressive changes indicate inwardly decreasing temperatures of crystallization or increasing water, or both. Overall, plagioclase in this suite contains distinctly less anorthite than other Sierran intrusive suites having the same general compositional range; these lower anorthite contents suggest a relatively high water content in the magma, though less than the amount required for saturation until the Johnson Granite Porphyry was emplaced. The occurrence of hornblende and biotite in discrete well-formed crystals and their increasing size inward into the equigranular facies of the Half Dome Granodiorite as their abundances decrease support this interpretation. A fine-grained groundmass and the presence of miarolitic cavities indicate that the magma was saturated with water when the Johnson Granite Porphyry was emplaced.

Variations in initial $^{87}Sr/^{86}Sr$ along traverse A–B (fig. 32) are shown graphically in figure 36 together with variations in SiO_2 for comparison. Both values increase inward from the margin of the suite into the equigranular facies of the Half Dome Granodiorite, remain nearly constant inward to the Johnson Granite Porphyry, then increase slightly in the Johnson. The total range of initial $^{87}Sr/^{86}Sr$ is 0.0009, from 0.7057 to 0.7066. Although the range of replicate analyses of individual samples is as much as 0.0004, the plot in figure 36 clearly shows that initial $^{87}Sr/^{86}Sr$ increases from about 0.7057 in the outer part of the tonalite of Glen Aulin to about 0.7064 in the middle of the equigranular facies of the Half Dome Granodiorite and that no significant changes take place farther inward to the Johnson Granite Porphyry, where it probably rises slightly to 0.7066.

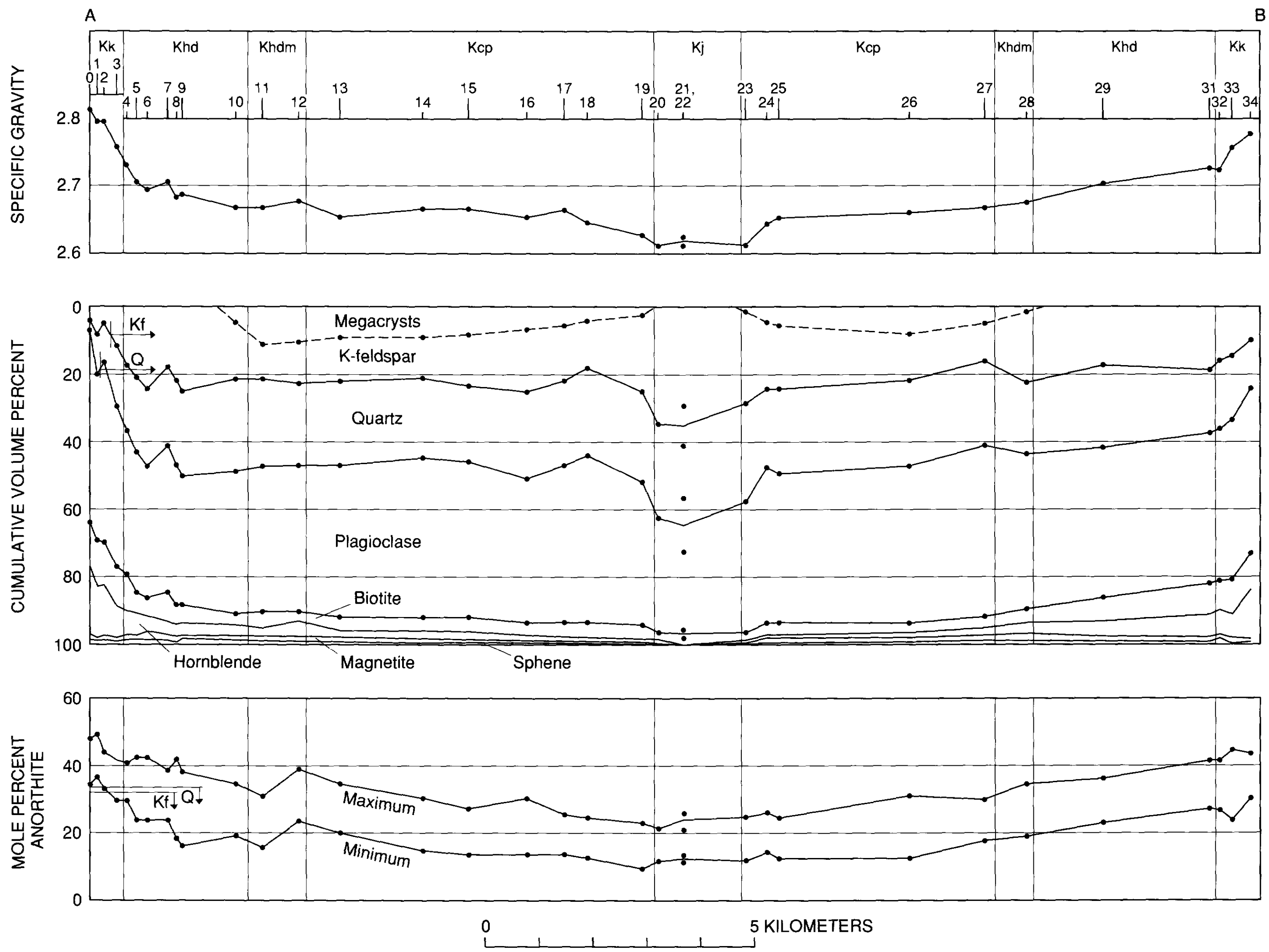


FIGURE 34.—Specific gravity, modes, and plagioclase compositions along traverse A–B, figure 32. Kk, granodiorite of Kuna Crest and other peripheral granitoids; Khd, equigranular facies of the Half Dome Granodiorite; Khdm, megacrystic facies of the Half Dome Granodiorite; Kcp, Cathedral Peak Granodiorite; Kj, Johnson Granite Porphyry. Q and Kf indicate composition of plagioclase crystallizing at beginning of crystallization of quartz and alkali feldspar. Numbers along traverse line are sample localities. Line plotted between data points indicates average values. From Bateman and Chappell (1979).

Because strontium isotopes can mix to give an average value but cannot unmix during crystal fractionation, the progressive inward increase of initial $^{87}Sr/^{86}Sr$ in the outer part of the suite requires that two source materials with different isotopic properties mixed, whereas the constant values across the megacrystic facies of the Half Dome Granodiorite and the Cathedral Peak Granodiorite imply that no further mixing occurred until the emplacement of the Johnson Granite Porphyry, which introduced slightly but distinctly increased $^{87}Sr/^{86}Sr$ (fig. 36).

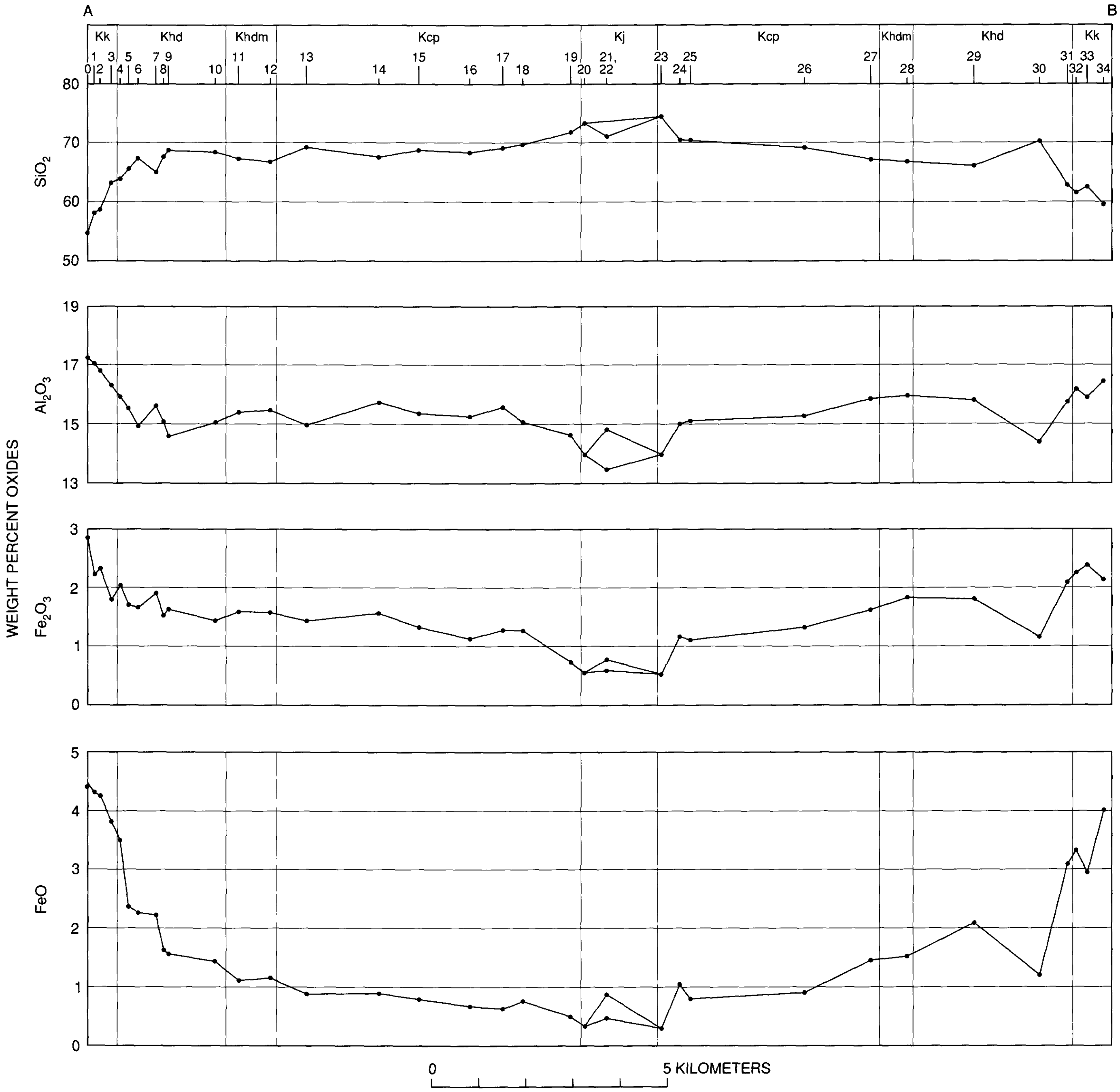


FIGURE 35.—Variations in major oxides along traverse A–B (fig. 32) across the Tuolumne Intrusive Suite. Kk, granodiorite of Kuna Crest and other granitoids in margins of batholith; Khd, equigranular facies of the Half Dome Granodiorite; Khdm, megacrystic facies of the Half Dome Granodiorite; Kcp, Cathedral Peak Granodiorite; Kj, Johnson Granite Porphyry. Numbers along traverse line are sample localities. From Bateman and others (1988).

The close correspondence of variations in SiO_2 with variations in initial $^{87}Sr/^{86}Sr$ suggests that mixing of source materials can explain all the compositional variations within the Tuolumne Intrusive Suite and that no other mechanism of differentiation is required. However, systematic variations in other oxides in the inner part of the suite, where SiO_2 and initial $^{87}Sr/^{86}Sr$ are constant, indicate that another

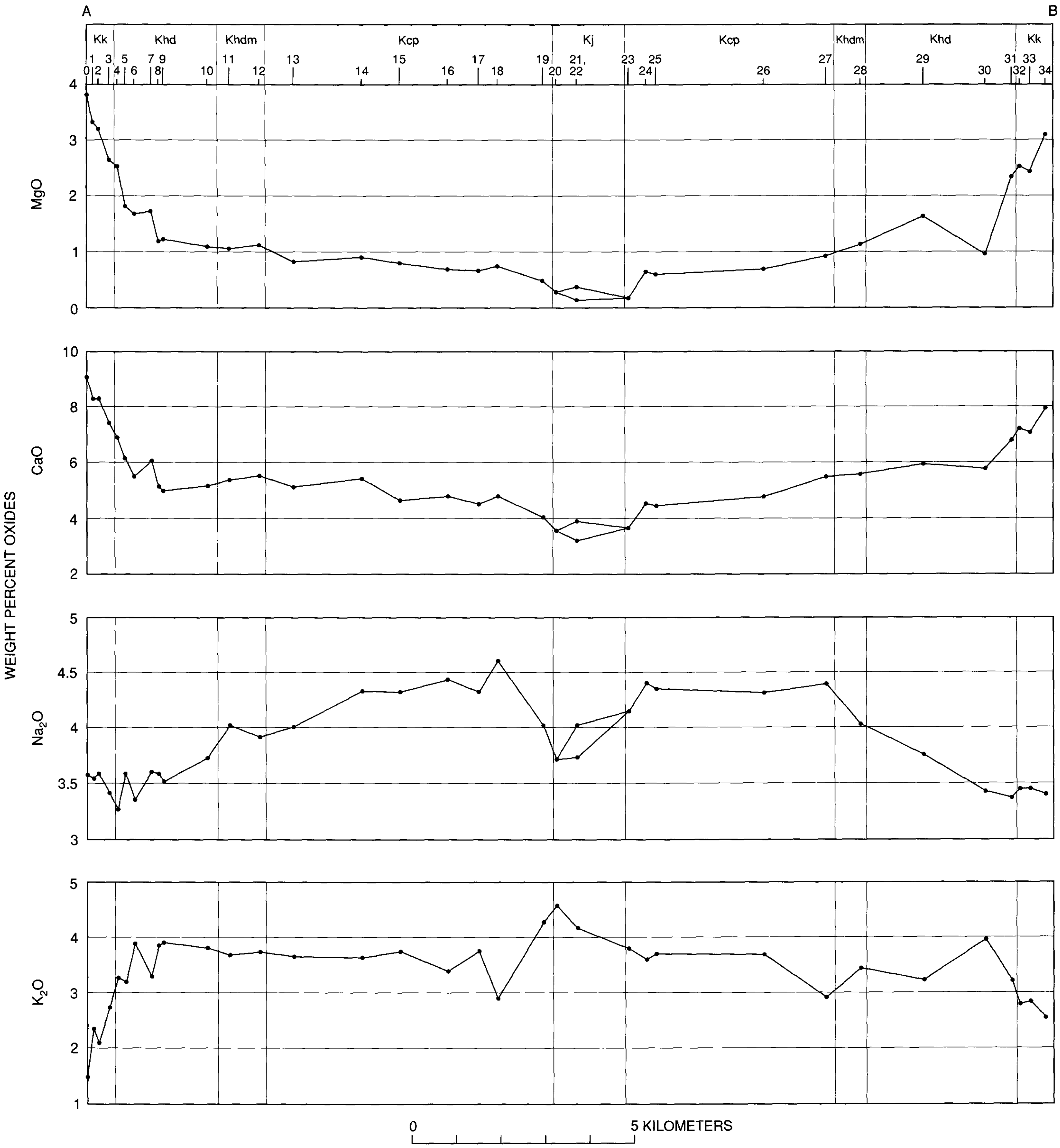


FIGURE 35.—Continued

process also was operating. Na_2O increases steadily from the margins of the suite inward to the Johnson Granite Porphyry where it decreases sharply, and FeO, Fe_2O_3, MgO, and CaO continue their inward decrease into the interior of the suite though by smaller amounts than in the outer parts (fig. 35). These changes reflect decreasing amounts of hornblende and biotite inward in the suite and decreasing anorthite in plagioclase. They are the result of inward-decreasing temperatures of crystallization and (or) increasing water, which control the solubility of these chemical constituents in the melt phase of the magma. The plagioclase contains more CaO and less Na_2O and the biotite and hornblende contain more FeO, Fe_2O_3, and MgO than the coexisting melt. CaO, Fe_2O_3, and FeO were constantly replenished in the coexisting melt phase by convection of interior magma and by diffusion in the melt phase. Consequently, FeO, Fe_2O_3, and MgO were progressively depleted and Na_2O enriched in the interior of the magma.

The explanation that I presently prefer for the compositional and isotopic patterns in the Tuolumne Intrusive Suite is that basaltic magma generated in the mantle mixed and mingled with materials in the lower crust before it rose into the upper crust and that it incorporated additional crustal material as it rose to form the compositionally and isotopically strongly zoned outer part of the Tuolumne Intrusive Suite. Discontinuous, indefinite contacts between more mafic and less mafic rocks readily visible in the Half Dome Granodiorite west of Tenaya Lake along California State Highway 120 show that mixing was far from perfect. Meanwhile, lower-crustal material that contained little or no mantle material was melting as a result of heat rising from the mantle or of underplating by mantle-derived basalt and eventually rose to form the inner parts of the Tuolumne Intrusive Suite. Crystal-liquid fractionation produced compositional changes in the inner part of the suite without changing the initial $^{87}Sr/^{86}Sr$. Although all the compositional variations in the outer part of the suite could be attributed to magma mixing and mingling, crystal-liquid fractionation should operate wherever a thermal gradient exists in a granitic magma. Unlikely to have been confined to the inner

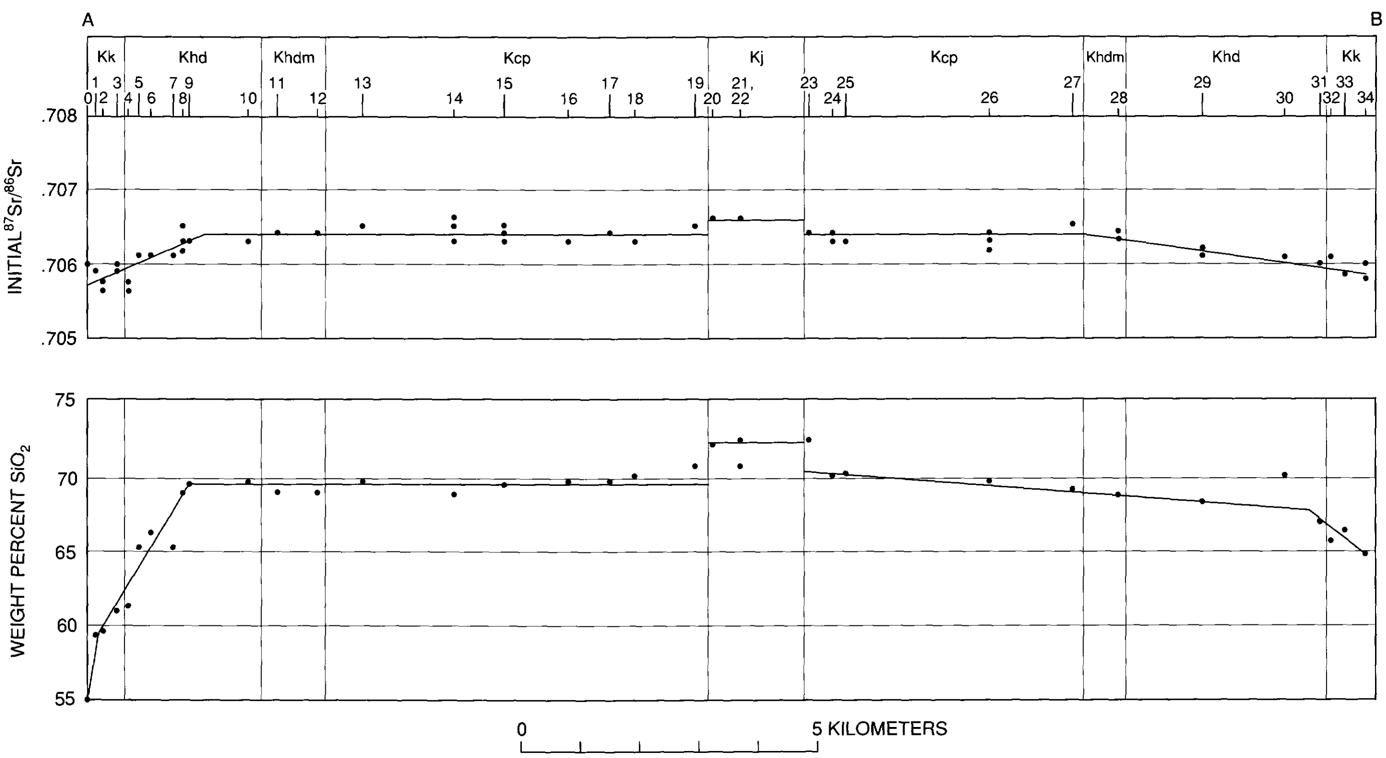


FIGURE 36.—Variations in initial $^{87}Sr/^{86}Sr$ (assuming age of 87 Ma) and SiO_2 along traverse A–B (fig. 32) across the Tuolumne Intrusive Suite. Kk, granodiorite of Kuna Crest and other granitoids in margins of batholith; Khd, equigranular facies of the Half Dome Granodiorite; Khdm, megacrystic facies of the Half Dome Granodiorite; Kcp, Cathedral Peak Granodiorite; Kj, Johnson Granite Porphyry. Lines through points indicate general variation trends. Numbers along traverse lines are sample localities. $^{87}Sr/^{86}Sr$ data from Kistler and others (1986).

part of the suite, crystal-liquid fractionation probably played a role in producing the compositional changes in the outer part of the suite as well.

SHAVER INTRUSIVE SUITE

Compositional variations within the Shaver Intrusive Suite are conspicuous but have not been studied in detail. The Dinkey Creek Granodiorite, the oldest and most extensive intrusion in the suite, ranges from equigranular tonalite and granodiorite containing abundant mafic inclusions to equigranular and megacrystic granite containing almost no mafic inclusions. The compositional variations are complex, but the higher elevation of the northern, more leucocratic part relative to the southern, more mafic part suggests that the variations are chiefly vertical. Units of the Shaver Intrusive Suite that intruded the Dinkey Creek Granodiorite include the megacrystic granodiorite of McKinley Grove, several younger masses of equigranular granite, and the granodiorite of Whisky Ridge (fig. 37).

Isopleths for specific-gravity values 2.70 and 2.75 based on determinations for samples collected from the Dinkey Creek Granodiorite reflect variations in bulk rock composition (fig. 37). In most places changes in specific gravity from one part of the intrusion to another are smooth, but the pattern is complex in the central part and in the tongue that penetrates the Fine Gold Intrusive Suite to the southwest. The tongue is of particular interest because of a repetition of specific-gravity fields. The arcuate pattern of foliation (pl. 1) and of the specific-gravity fields (fig. 37) shows that the tongue was forcibly protruded, and it seems reasonable to infer that it has a shallow bottom. Section *A–A'* (fig. 37), drawn along the long axis of the tongue, portrays the specific-gravity variations as having resulted from folding of crystallizing magma that was progressively denser downward. Section *B–B'*, drawn parallel to the long axis of the Dinkey Creek Granodiorite, is consistent with the interpretation that the compositional variations within the Dinkey Creek are vertical rather than horizontal.

The granodiorite of McKinley Grove and the megacrystic facies of the Dinkey Creek Granodiorite are similar in composition, texture, and specific gravity (<2.70). Nevertheless, the granodiorite of McKinley Grove has sharp contacts with the Dinkey Creek Granodiorite and was emplaced after magmatic foliation had been established in the Dinkey Creek. The granodiorite of McKinley Grove was intruded by still younger and more felsic rocks such as the granites of Dinkey Dome and north of Snow Corral Meadow.

The relations shown in conjectural sections *A–A'* and *B–B'* fig. 37) imply that the Dinkey Creek magma was density-stratified before the tongue of Dinkey Creek Granodiorite was protruded toward the southwest and folded and before steeply dipping foliations were imposed. After the Dinkey Creek magma was generated, it is unlikely that country rocks remained unmelted at depth and later provided a source for the granodiorite of McKinley Grove and the younger, less dense, and more felsic rocks of the Shaver Intrusive Suite. It seems far more likely that fractionation processes concentrated salic magma in the core as well as at the top of the cooling and crystallizing Dinkey Creek magma.

MOUNT GIVENS GRANODIORITE

The Mount Givens Granodiorite consists of a large elongate pluton that terminates in a bulbous head at its northwest end. Most of the rock is equigranular granodiorite, but the core and concentric zones within the bulbous head are megacrystic (fig. 38). The equigranular granodiorite facies contains many lensoid mafic inclusions; well-formed crystals of biotite and hornblende are scarce. The alkali feldspar megacrysts in the megacrystic facies are smaller than those in the Tuolumne Intrusive Suite, and their crystal faces generally are irregular. Most contacts between megacrystic and equigranular rock are gradational; however, the outer contact of the horseshoe-shaped inner zone of dark equigranular granodiorite is sharp, and the dark equigranular rock intrudes the megacrystic rock. A distinctly higher average anorthite content in the plagioclase of the Mount Givens Granodiorite than in that of the Tuolumne Intrusive Suite and the presence of ilmenite (absent in the Tuolumne) indicate a lower water content in the Mount Givens magma. The occurrence of most of the hornblende and biotite as euhedral crystals in the Half Dome and Cathedral Peak Granodiorites of the Tuolumne Intrusive Suite and a dearth of such crystals in the Mount Givens Granodiorite may also reflect the lower water content of the Mount Givens magma.

The spatial relations of the different facies of the Mount Givens Granodiorite were interpreted in various ways in earlier publications (Huber, 1968; Bateman and others, 1971; Bateman and Nokleberg, 1978) but in this report are interpreted to indicate that the core area, including the horseshoe-shaped body of equigranular granodiorite and

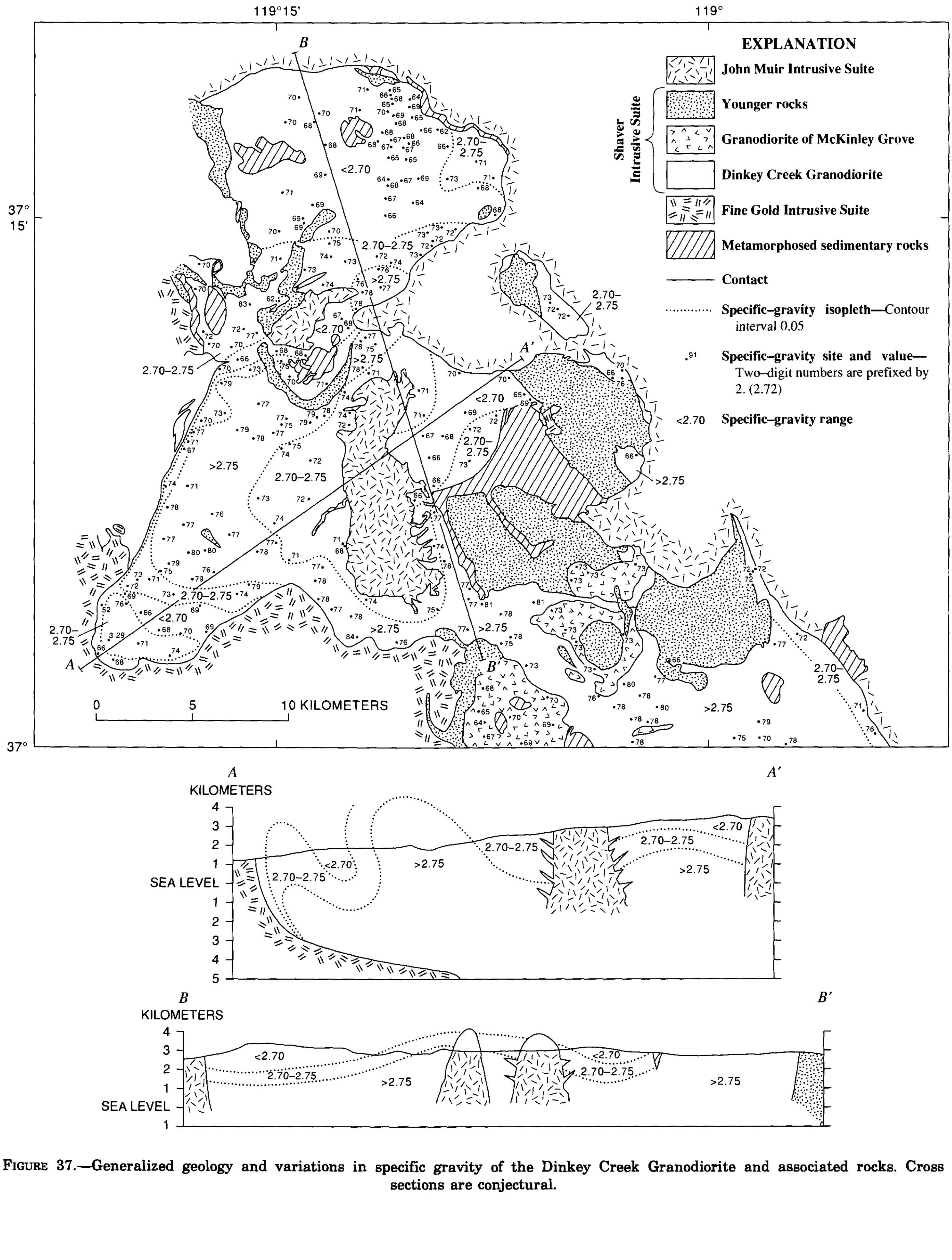


FIGURE 37.—Generalized geology and variations in specific gravity of the Dinkey Creek Granodiorite and associated rocks. Cross sections are conjectural.

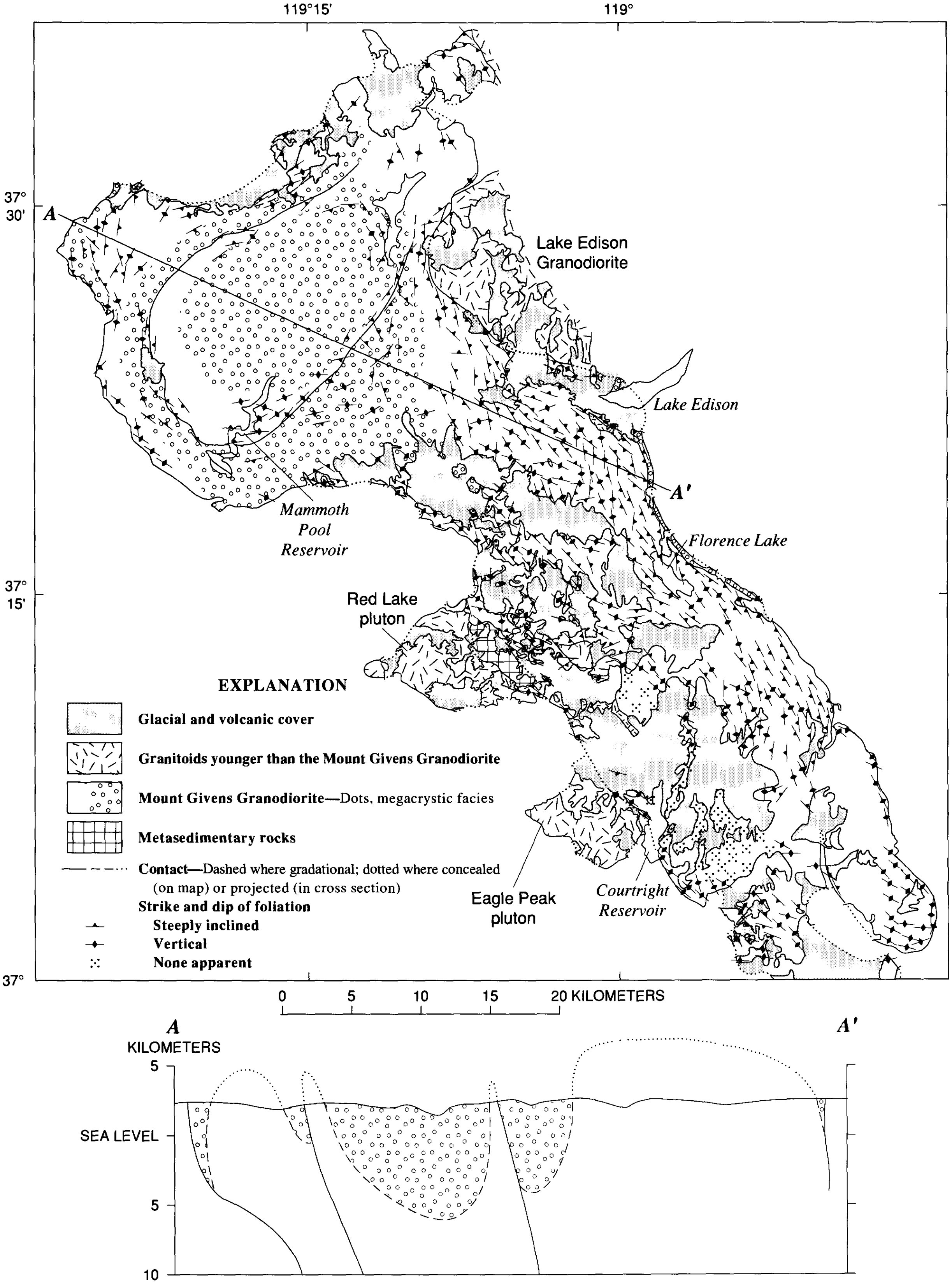


FIGURE 38.—Generalized geology of the Mount Givens Granodiorite showing foliation patterns. Cross section is conjectural. Rocks bordering the Mount Givens Granodiorite are older granitoids or metamorphic rocks except where other units are shown.

the enclosed megacrystic core rock, represents an upwardly resurgent intrusion. The bimodal distribution of compositions in the core area is similar to that in the granodiorite of McMurry Meadows of the Palisade Crest Intrusive Suite, which Sawka (1985) attributed to sidewall crystallization made extraordinarily effective by slow cooling of surrounding heated rock.

Apparently, at the currently exposed level, the bulbous head of the pluton expanded as it was emplaced, and it seems likely that the alternating pattern of megacrystic and equigranular rock in the margins was caused by folding during the expansion of originally horizontal or gently dipping layering in which megacrystic magma overlaid denser, equigranular magma.

Bateman and Nokleberg (1978) examined a series of samples that were collected along a traverse that extends from the southwest margin to the core of the bulge (fig. 39, traverse A–B). Although the grain size appears to increase inward along this traverse, only quartz and potassium feldspar grains increase in size; biotite and hornblende decrease inward in both size and abundance. Decrease in the grain size of the mafic minerals precludes chilling as the cause of the apparent finer grain size of the marginal rock. These relations are compatible with longer growth times or increased water content in the magma during inward crystallization and with progressive impoverishment of the magma in iron and magnesium for the mafic minerals.

The general pattern of abundance of minerals along the traverse investigated by Bateman and Nokleberg (1978) shows that quartz and alkali feldspar increase inward, plagioclase changes little, and hornblende and biotite decrease (fig. 40). These changes are interrupted at the intrusive contact of the horseshoe-shaped mass of granodiorite and are partly repeated farther inward. The compositional change is greatest in the outer 1 km, as in the Tuolumne Intrusive Suite, but continues inward to the porphyritic facies.

Bateman and Nokleberg (1978) attributed the compositional changes along their traverse across the Mount Givens Granodiorite to crystal fractionation, just as Bateman and Chappell (1979) later did for the Tuolumne Intrusive Suite. Mixing of source materials, as is indicated by isotopic data for the outer part of the Tuolumne Intrusive Suite, is a possible explanation for at least part of the zoning in the Mount Givens Granodiorite but cannot be evaluated because of an absence of detailed isotopic data. However, similar compositional zoning in the Palisade Crest Intrusive Suite occurs without isotopic change.

PALISADE CREST INTRUSIVE SUITE

Sawka (written and oral communs., 1982–1985; 1985) investigated compositional variations within the northwestern part of the Palisade Crest Intrusive Suite in the eastern escarpment of the Sierra Nevada southwest of Big Pine. The area that he studied includes the western lobe of the Tinemaha Granodiorite and the granodiorite of McMurry Meadows, a roughly circular body that is enclosed within and intrudes the Tinemaha Granodiorite (fig. 41). These intrusions are of interest because the Tinemaha Granodiorite offers an opportunity to investigate vertical as well as horizontal compositional variations in the field and because initial $^{87}Sr/^{86}Sr$, though different for the two intrusions, is constant within each intrusion (Sawka, 1985). Thus, the compositional variations within the

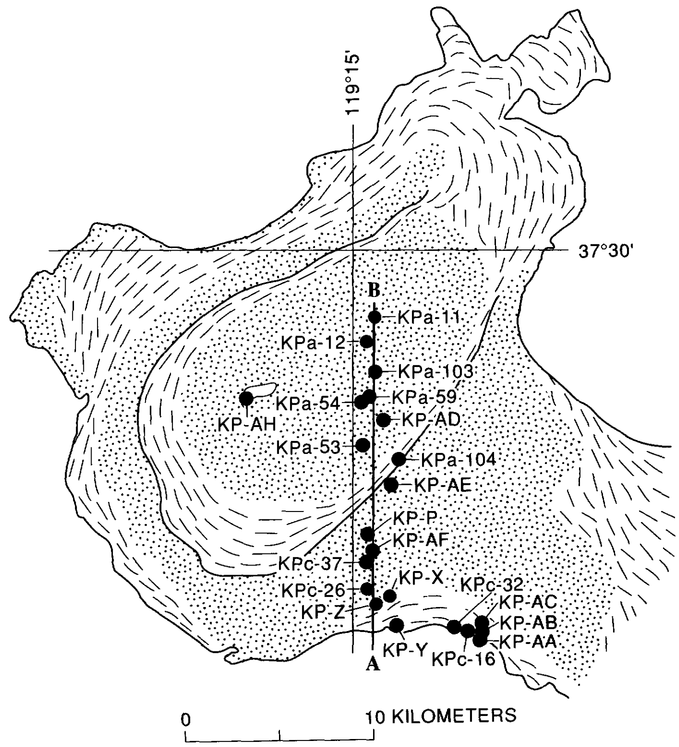


FIGURE 39.—North end of the Mount Givens Granodiorite showing traverse A–B and sample localities (dots) referred to in figure 40 and in text. Megacrystic facies is shown by stipple, and foliation trends in equigranular facies by dash pattern. Solid internal contact indicates sharp boundary between facies. Unpatterned area at sample locality KP-AH is aplitic facies of the Mount Givens at Jackass Rock. From Bateman and Nokleberg (1978).

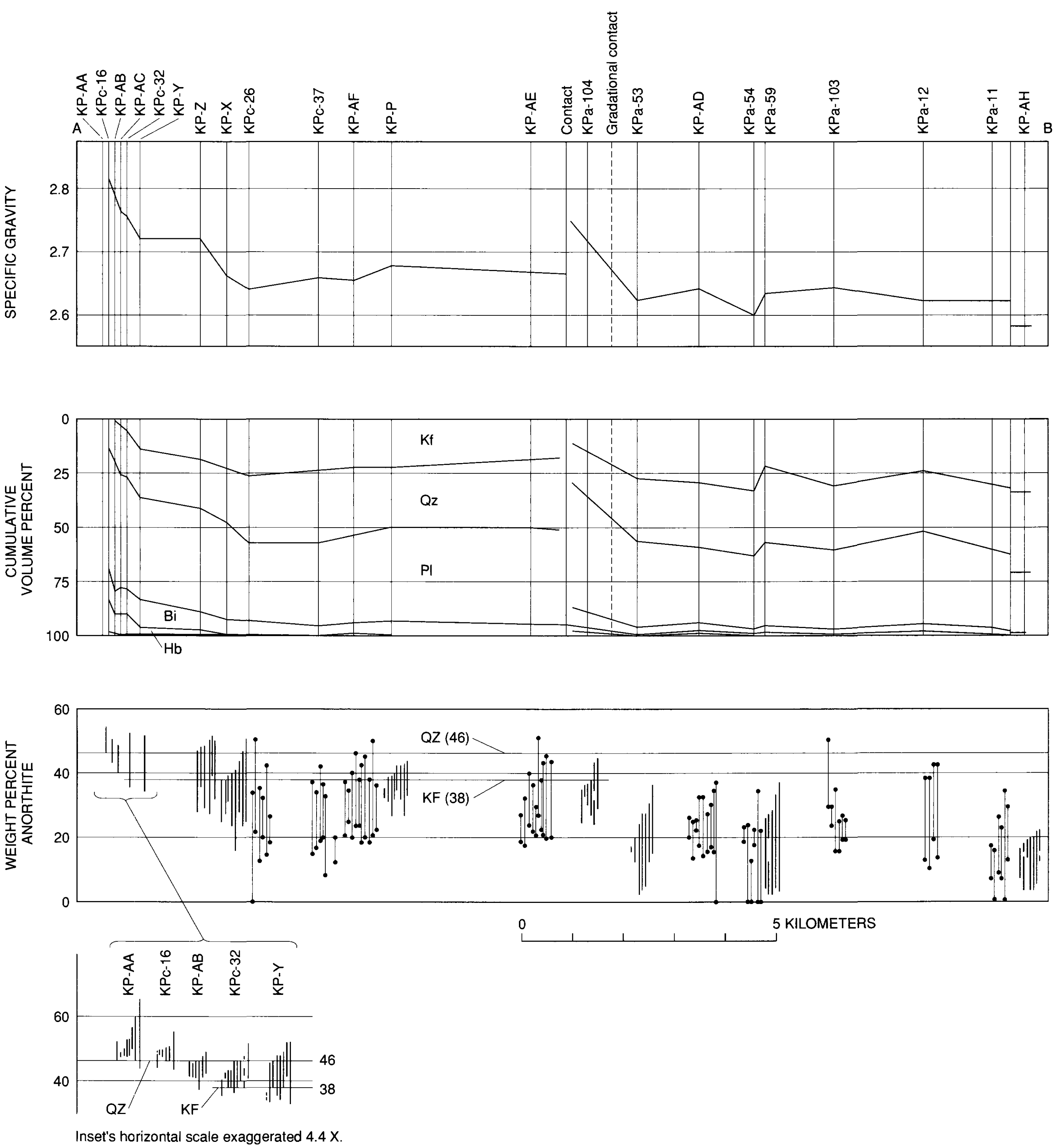


FIGURE 40.—Specific-gravity values, modes, and plagioclase compositions of samples collected from the Mount Givens Granodiorite along traverse A–B, figure 39. Kf, alkali feldspar; Qz, quartz; Pl, plagioclase; Bi, biotite; Hb, hornblende. Microprobe determinations of plagioclase compositions are shown by lines and optical determinations by lines terminating in dots. Length of lines shows range of An content of zoned plagioclase crystals. Horizontal lines QZ and KF show composition of plagioclase crystallizing at beginning of crystallization of quartz and alkali feldspar. From Bateman and Nokleberg (1978).

Palisade Crest Intrusive Suite are the result of some mechanism other than the mixing of source materials with different isotopic properties.

TINEMAHA GRANODIORITE

The western lobe of the Tinemaha Granodiorite crops out at altitudes ranging from 2,073 m (14,058 ft) in the southeast corner to 4,285 m (6,800 ft) at Split Mountain. Part of this large range in altitude is the result of faulting, but the altitudinal range within unbroken blocks is sufficient to permit study of vertical as well as horizontal variations.

On the west, the Tinemaha Granodiorite is in contact with the younger Inconsolable Quartz Monzodiorite. Discontinuous septa of metamorphic rocks along the contact show that little or none of the original margin of the Tinemaha was eliminated when the Inconsolable was emplaced, but the presence of extremely flattened and attenuated mafic inclusions close to this contact in both intrusions suggests that the peripheral part of the Tinemaha may have been remobilized by the intrusion of the Inconsolable Quartz Monzodiorite. The foliation in both rock units is steep or vertical close to the contact, but the foliation in the Tinemaha Granodiorite flattens toward the east and is horizontal or gently dipping in the interior of the western lobe. As magmatic foliation generally parallels or diverges only slightly from contacts with older rocks, the gently dipping and horizontal foliation is assumed to indicate that before erosion the east-central part of the Tinemaha Granodiorite was overlain by a roof of older rocks resting on a generally horizontal or gently undulating surface, probably not far above the highest levels now exposed. A roof of older rocks is also suggested by the presence of several remnants of metasedimentary rocks nearby.

The Tinemaha Granodiorite ranges widely in composition from quartz monzodiorite to granite, and some modes plot in the granodiorite and quartz monzonite fields on a Q-A-P diagram (fig. 86). Sawka (1985) collected samples along an equal-altitude traverse at 3,300 m, which extends inward from the western margin of the Tinemaha Granodiorite, across the zone of steeply dipping foliations and into the zone of horizontal foliations, and then across the granodiorite of McMurry Meadows (traverse A–A' in fig. 41). He also collected samples upward from this traverse through a vertical distance of about 900 m to an altitude of 4,200 m, chiefly along traverses B–B' and C–C' (fig. 41). The samples collected along the equal-altitude traverse (A–A') show that inward from the west contact for about 0.4 km, across the belt of steeply dipping foliations, the rocks become more felsic and that the composition is relatively constant within the zone of horizontal foliations (fig. 42). Samples collected adjacent to the Inconsolable Quartz Monzodiorite, within the zone of steeply dipping foliations and, to a lesser extent, close to the contact with the granodiorite of McMurry Meadows, contain much more plagioclase and biotite and less quartz and alkali feldspar than the samples from the interior of the western lobe of the Tinemaha Granodiorite. Clinopyroxene is present in the western margin of the Tinemaha Granodiorite but is scarce in the interior of the intrusion.

Samples collected within the zone of horizontal or gently dipping foliations from 4,200 m down to 3,300 m (the level of equal-altitude traverse A–A') show the following changes with decreasing altitude along traverses B–B' and C–C' (figs. 41 and 43): (1) Specific gravity increases from about 2.70 to about 2.76, (2) plagioclase increases from about 32 to 45 percent, (3) hornblende increases from about 7 to 15 percent, (4) quartz decreases from about 23 to 13 percent, and (5) alkali feldspar decreases from about 29 to 17 percent. Chemical analyses of the bulk rock samples show chemical variations that are consistent with the modal variations—CaO, MgO, Al_2O_3, total iron as Fe_2O_3, TiO_2, P_2O_5, and MnO increase and SiO_2 and K_2O decrease downward.

The model offered by Sawka (1985) to explain the vertical distribution of compositions is that crystals of mafic and calcic minerals accreted to the sidewalls of the magma chamber and lowered the density of the adjacent magma, which then flowed buoyantly to the top of the magma chamber. More rapid loss of femic and calcic constituents to the crystallizing margins than gain by diffusion from the interior of the magma chamber caused the magma adjacent to the solidifying margin to become progressively more felsic and less dense. Consequently, later magma flowed past earlier magma to the top of the chamber and produced the vertical compositional gradation within the zone of flat-lying foliations. The observed or inferred distribution of rock compositions within the lobe is shown in figure 44. This model has been demonstrated in laboratory experiments using saline solutions (McBirney, 1980; Turner, 1980).

GRANODIORITE OF MCMURRY MEADOWS

The rock in the margins of the granodiorite of McMurry Meadows has almost the same compositions as the rock in the western margin of the Tinemaha Granodiorite and, like the Tinemaha Granodiorite, is increasingly felsic inward (fig. 41

and 42). However, in contrast to the gradational compositional changes inward in the Tinemaha Granodiorite, the granodiorite of McMurry Meadows is bimodal. A narrow transitional zone separates the equigranular quartz monzodiorite facies in the margins from megacrystic granite in the interior. Plagioclase crystals in both the peripheral and interior rock have core compositions as calcic as An_{75}, but rim compositions average about An_{25} in the margins and about An_{15} in the interior. Biotite is uniformly more abundant than hornblende. Some euhedral to subhedral hornblende crystals have augite cores, especially within the rock in the margins.

Sawka (1985) postulated that when the granodiorite of McMurry Meadows was emplaced, the Tinemaha Granodiorite was still hot and formed a thermal blanket that prevented rapid loss of heat and volatiles at the sidewalls of the McMurry Meadows. The slow rate of inward crystallization and solidification fostered a balance between the loss of

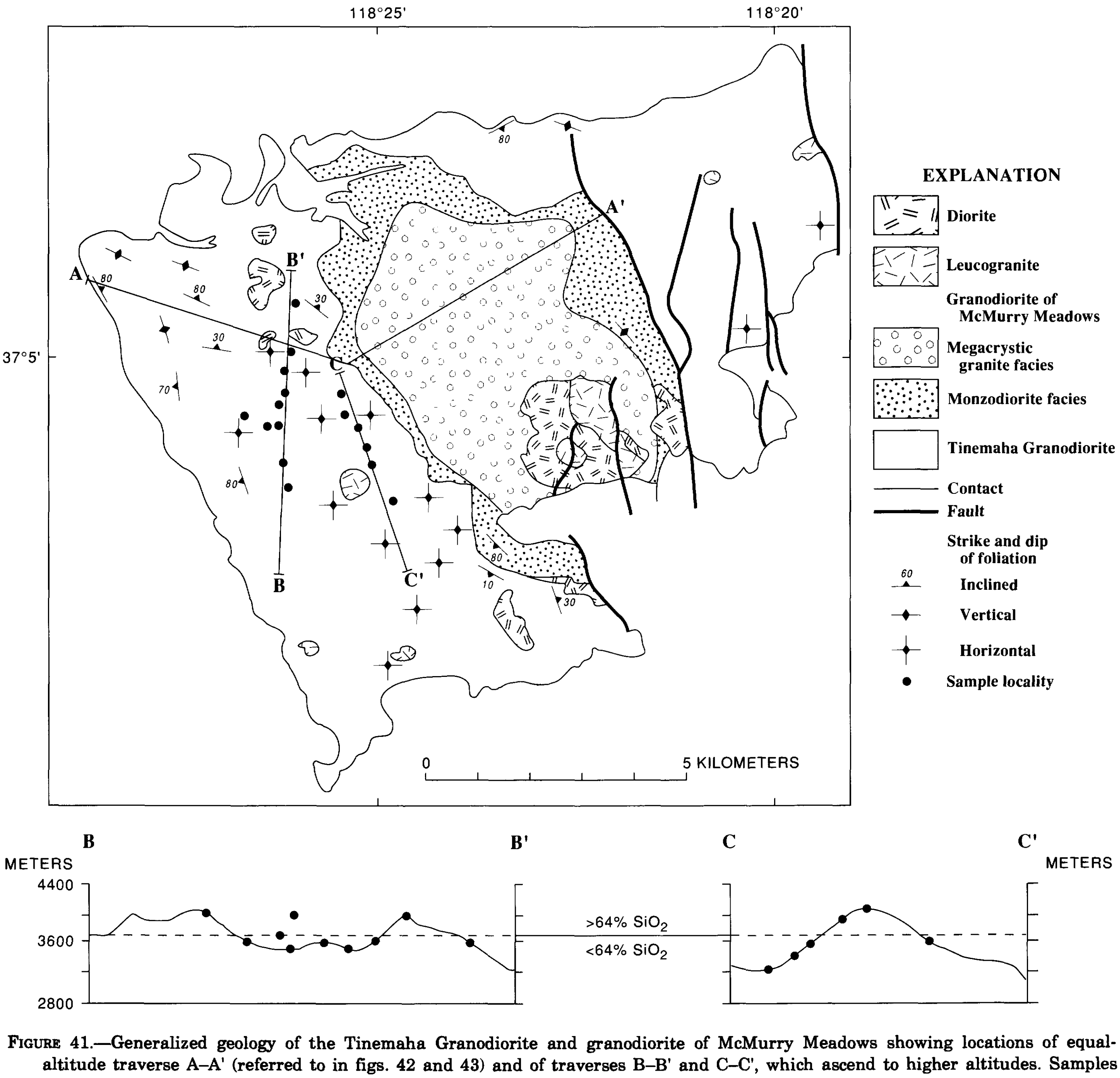


FIGURE 41.—Generalized geology of the Tinemaha Granodiorite and granodiorite of McMurry Meadows showing locations of equal-altitude traverse A–A' (referred to in figs. 42 and 43) and of traverses B–B' and C–C', which ascend to higher altitudes. Samples collected along traverses B–B' and C–C' are also shown on profiles. From Sawka (1985).

constituents from the boundary layer by crystallization and accretion at the margin and the renewal of constituents by diffusion from the interior magma. Slow but efficient precipitation and accretion of crystals at the sidewalls were almost equaled by diffusion of material from the core magma. Consequently, the compositional change and density decrease within the boundary magma were small and the core magma did not rise as buoyantly as the boundary magma within the Tinemaha Granodiorite (fig. 44). Thus, the minerals that accreted at the margins represent one peak in a bimodal pattern, and the depleted magma, the other peak.

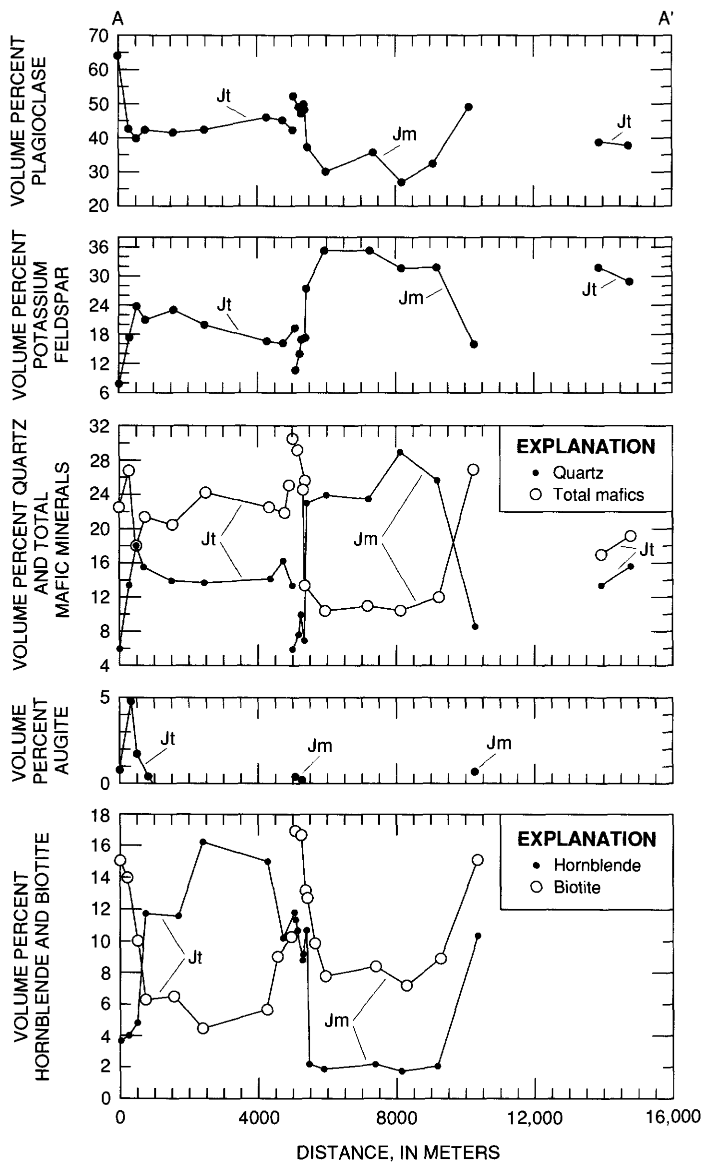


FIGURE 42.—Variations in modal abundance of minerals in the Tinemaha Granodiorite (Jt) and granodiorite of McMurry Meadows (Jm) along equal-altitude traverse A–A' of figure 41. Modified from Sawka (1985).

REGIONAL PATTERNS

Progressive change in the average composition of the granitoids across the Sierra Nevada has been known since Lindgren (1915) wrote that the earliest intrusions in the Cordilleran region are peridotite, pyroxenite, and gabbro in the west and that the intrusions gradually spread eastward and became progressively more felsic. Buddington (1927) showed that the intrusive rocks of the Coast Ranges of Alaska are arranged in belts that are progressively richer in SiO_2 and K_2O eastward and pointed out the similarity of this pattern to the pattern in the Sierra Nevada. Since these two early reports, Moore (1959), Wollenburg and Smith (1968, 1970), Bateman and Dodge (1970), Dodge (1972), Bateman (1979), and Dodge and others (1982) described west-to-east compositional changes in terms of a variety of parameters.

In the field it is obvious that the rocks in the western foothills have a high color index and a low alkali feldspar content and that the color index decreases and the alkali feldspar content gradually increases eastward. These changes are indicative of systematic variations in the modal, major-element, minor-element, and isotopic compositions of the rocks.

MODAL VARIATIONS

Modal variations across the batholith have already been described in connection with the interpretation of modal plots on Q-A-P diagrams (fig. 29). The east-to-west modal variations are also shown by projections of modal mineral contents onto a line that extends from the southwest corner of the map area (pl. 1) through Bishop to the east boundary (line A–B, fig. 45). The projection (fig. 46) shows that alkali feldspar increases eastward from an average of about 5 percent in the west to about 30 percent in the east and that plagioclase decreases from an average of about 60 percent in the west to less than 40 percent in the east. The alkali feldspar content remains constant for about 50 km from the west end of line A–B (fig. 45) and then increases eastward to an average of about 30 percent. Plagioclase decreases irregularly from an average of a little less than 60 percent in the west to about 35 percent in the east but increases slightly at the extreme east end (in the White Mountains). The content of mafic minerals increases for a short distance from 10 percent in the west to about 20 percent, decreases to about 5 percent at the east base of the Sierra Nevada, and increases to about 10 percent in the White Mountains. Quartz varies irregularly from 20 to 30 percent except in the Tungsten Hills near Bishop, where it increases to about 34 percent.

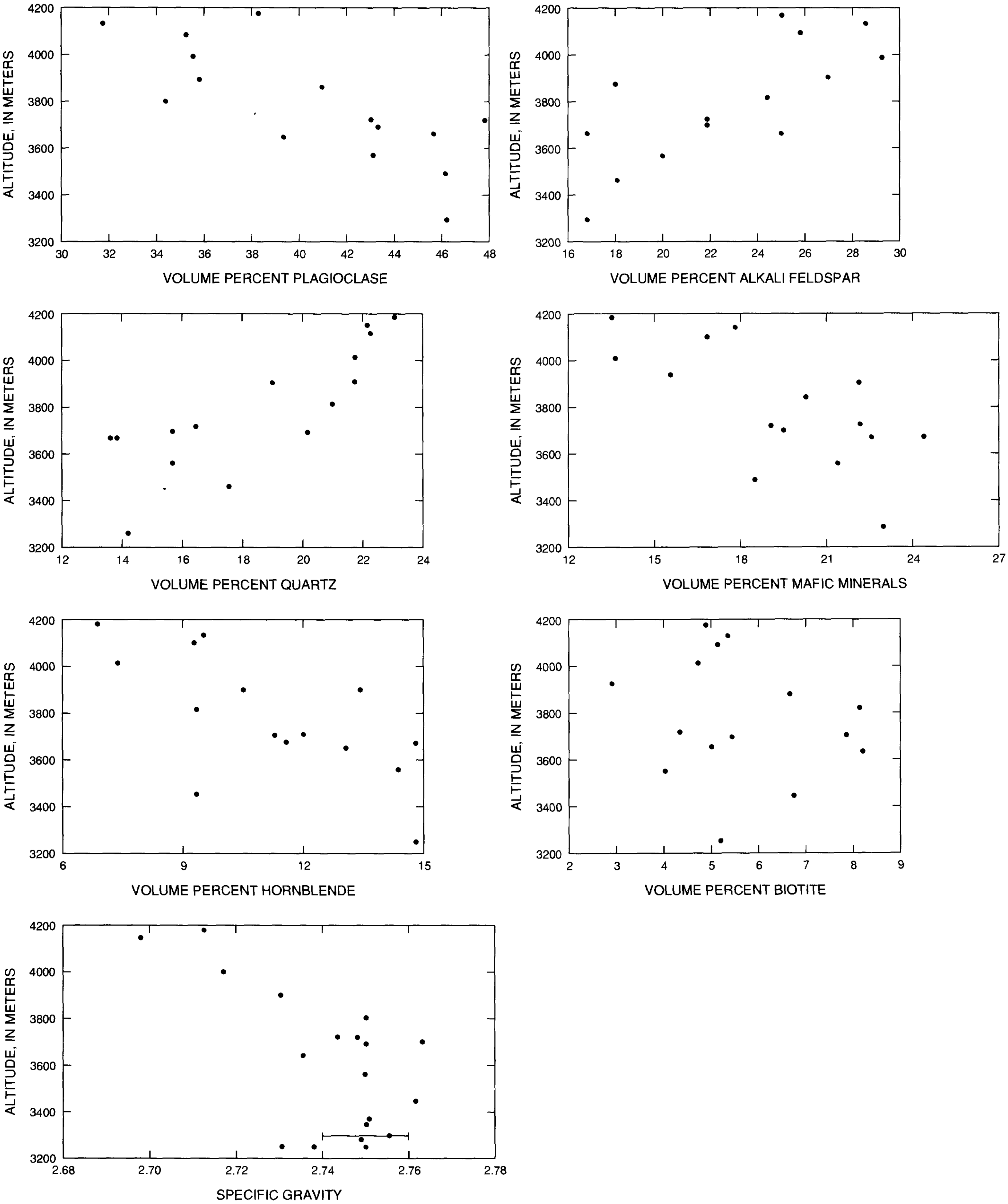


FIGURE 43.—Variations in modal abundance of minerals and in specific gravity with altitude within interior of western lobe of the Tinemaha Granodiorite. Bar shows range of specific gravity along equal-altitude traverse A–A' (figs. 41, 42). Modified from Sawka (1985).

VARIATIONS IN MAJOR OXIDES

The variations in major-oxide contents reflect the regional modal variations. In figure 47, the major-oxide contents (in weight percent) of most chemically analyzed samples from within the map area (pl. 1) have been projected onto line A–B (fig. 45), which is perpendicular to the long axes of the

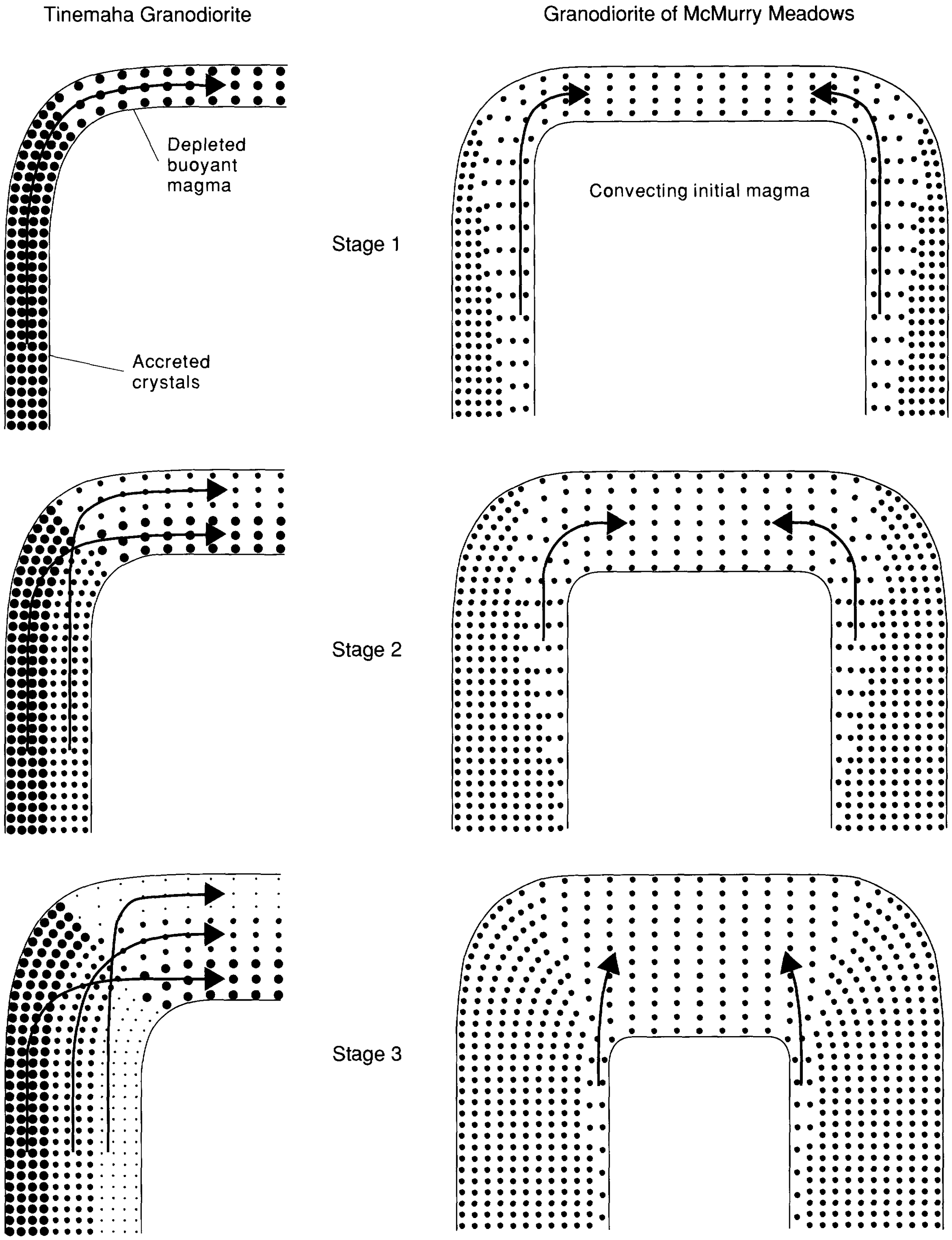
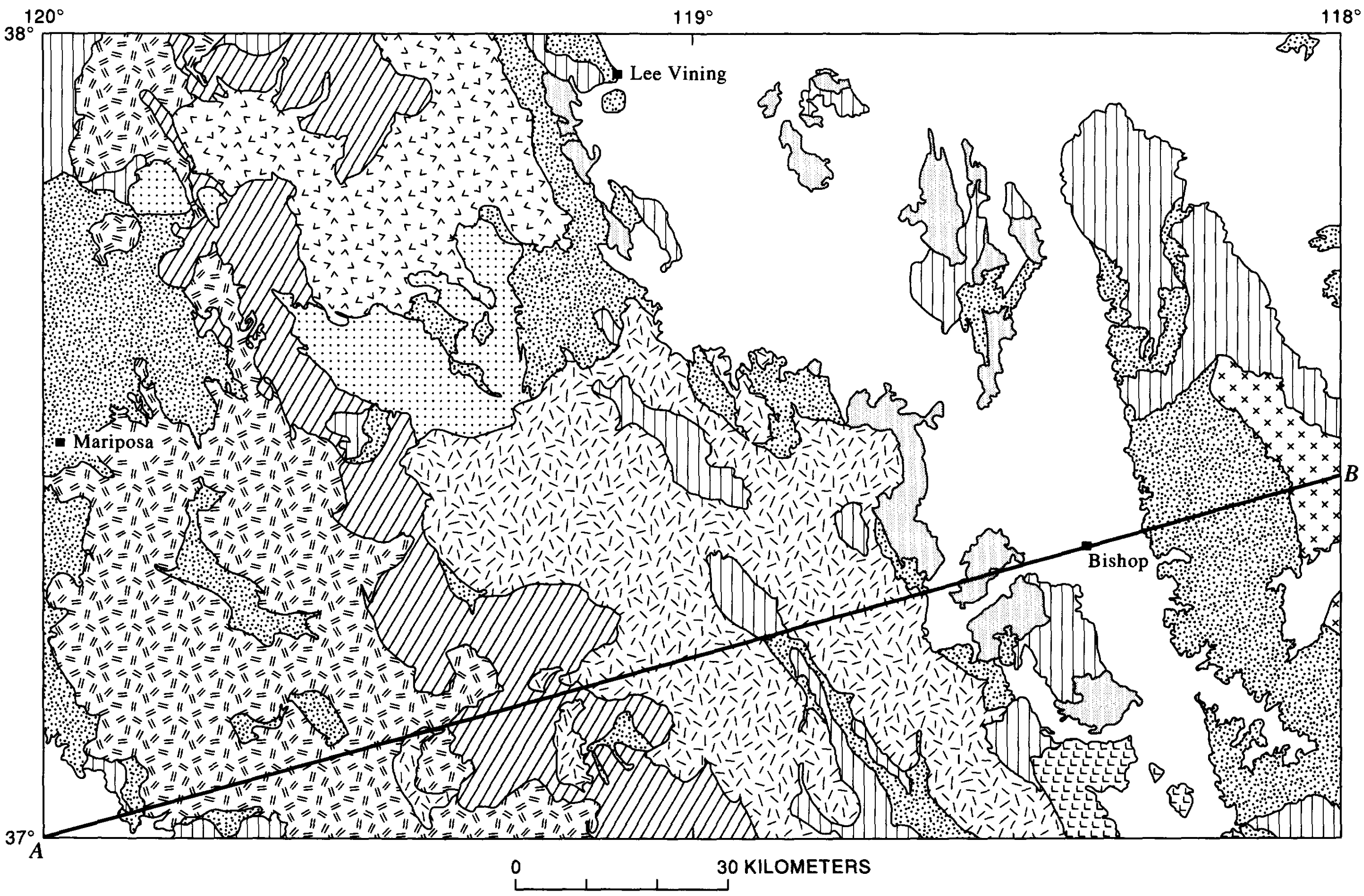


FIGURE 44.—Sidewall-accretion models for western lobe of the Tinemaha Granodiorite and granodiorite of McMurry Meadows. Arrows show direction of flow of depleted buoyant magma. Dot size and spacing indicate relative crystal size and density. See text for detailed discussion. Modified from Sawka (1985).

major intrusions. Samples from the northwestern part of the map area, which is distant from the line of projection, were not used. Because far fewer samples were analyzed chemically than modally and chemical data were projected from much greater distances, the plots of data points and moving-average curves are much less regular than for the modes. Curve irregularities are especially noticeable in the eastern part of the traverse, where samples are few and much of the major-oxide data has been projected long distances. Cumulative curves for both modes and major oxides are shown in figure 48.

The plot of K_2O shows the greatest differences along the traverse, and the plots of SiO_2 and Na_2O show the least differences (fig. 47). The moving-average curve for K_2O increases from about 2 percent in the west to about 4 percent in the central part of the traverse, remains constant as far as the east base of the Sierra Nevada, and then rises to



EXPLANATION

Cenozoic alluvial and volcanic rocks

Tuolumne Intrusive Suite

John Muir Intrusive Suite

Intrusive suites of Washburn Lake, Buena Vista Crest, and Merced Peak

Shaver Lake Intrusive Suite and intrusive suite of Yosemite Valley

Fine Gold Intrusive Suite

Palisade Crest Intrusive Suite

Soldier Pass Intrusive Suite

Scheelite Intrusive Suite

Unassigned intrusive rocks

Country rocks

FIGURE 45.—Simplified geology of area of Mariposa 1° by 2° quadrangle showing distribution of country rocks, intrusive suites, and unassigned intrusive rocks. Line A–B on map is line of projections shown in figures 46 through 48 and figure 57.

about 4.5 percent in the White Mountains (fig. 47). The moving-average curve for SiO_2 ranges nonsystematically from 61 to 72 percent; values are generally a little lower in the west (where samples were collected mostly from the Fine Gold Intrusive Suite) than in the middle and eastern parts of the traverse.

In figure 49, K_2O is plotted against SiO_2 for several representative suites. The plots clearly show that K_2O is progressively higher eastward and that this eastward increase occurs in steps between the different suites rather than continuously as implied by the curves projected along line A–B (fig. 47). The smallest amounts of K_2O (1.2 to 3 percent) are contained in the Fine Gold Intrusive Suite, and the largest amounts (3.6 to 5.4 percent) are contained in the Jurassic rocks of the White Mountains. The general range of SiO_2 for all suites represented in figure 49 is from 57 to 74 percent, but the range is slightly narrower in the Fine Gold Intrusive Suite

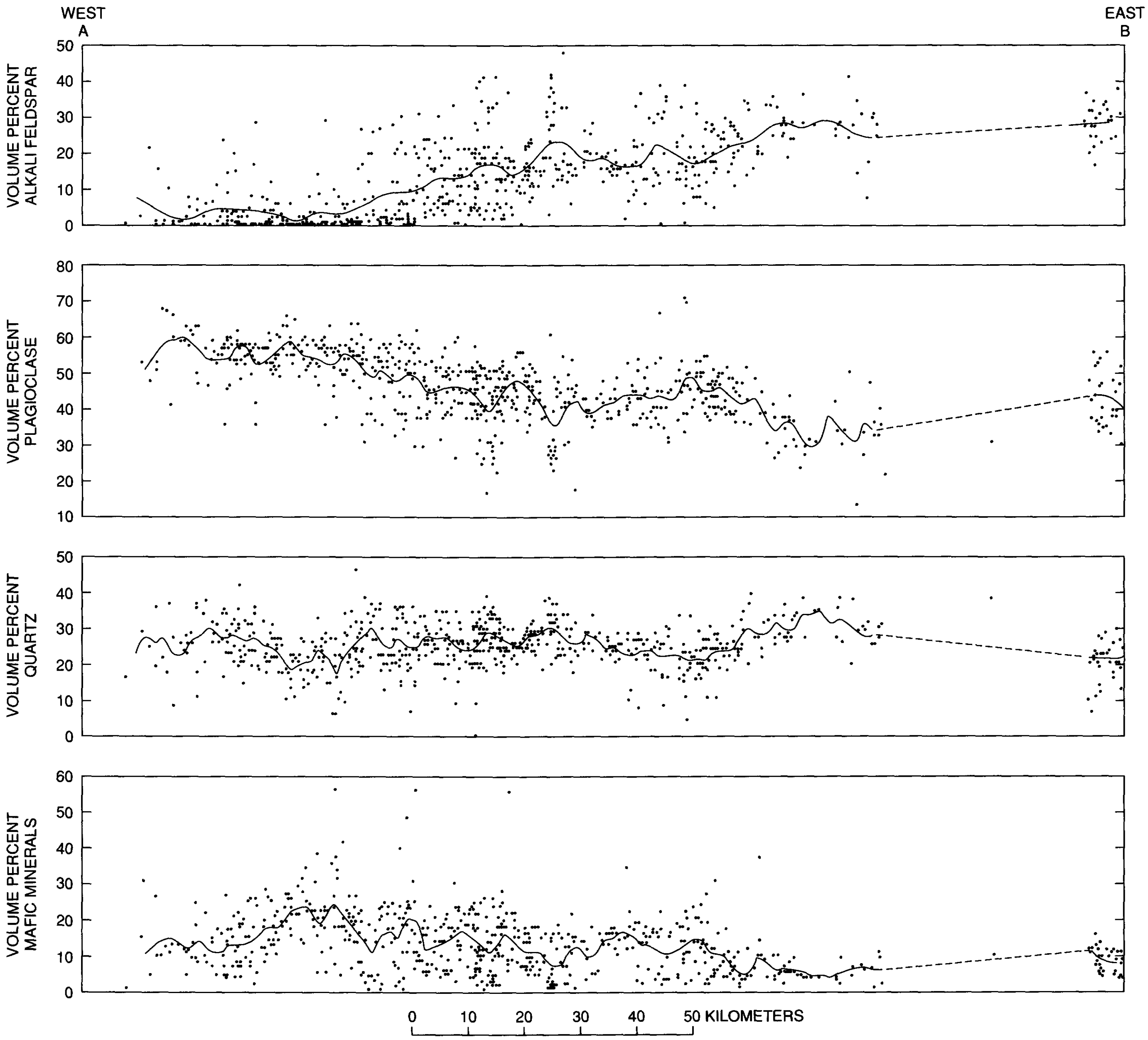


FIGURE 46.—Modal abundances of minerals in samples of intrusive rocks collected within 3 km of and projected onto line A–B, figure 45. Moving-average curves constructed by averaging zones that overlap 50 percent. Curves dashed where alluvium conceals bedrock and no samples were collected.

(58.5 to 72 percent). The smallest amount of SiO_2 is 55 percent, in a sample from the John Muir Intrusive Suite, and the largest amount is 77 percent, in a sample from the Shaver Intrusive Suite.

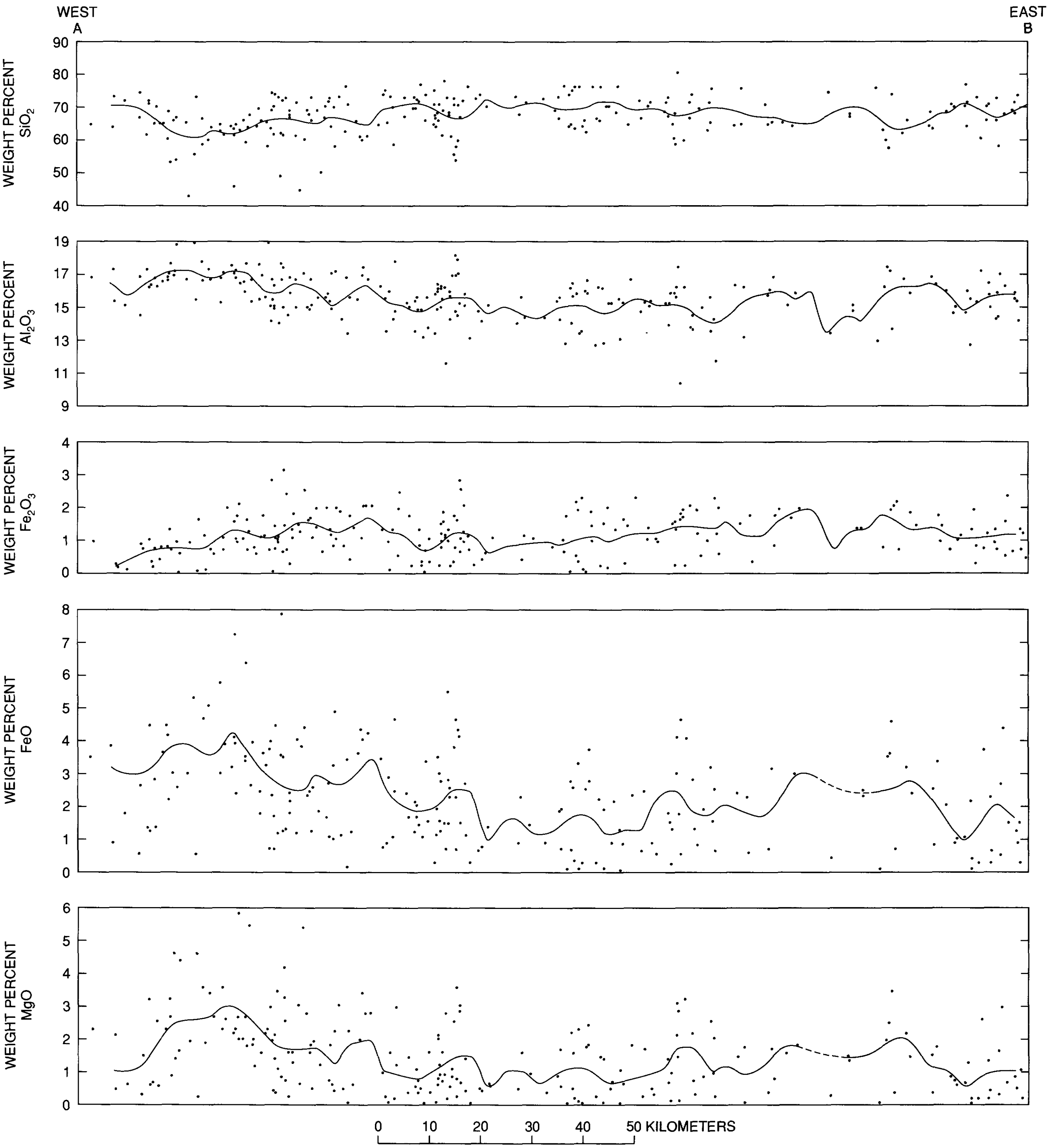


FIGURE 47.—Major-oxide contents of samples projected onto line A–B, figure 45. Includes all samples collected within Mariposa 1° by 2° quadrangle except those in northwest quarter of map area (pl. 1). Moving-average curves constructed by averaging zones that overlap 50 percent. Curves are dashed where few samples project onto line A–B. Figure continued on following page.

The moving-average curve for Na_2O (fig. 47) varies from 3.0 to 3.8 percent except along a short span in the western part of the traverse where it decreases to as low as 2.6 percent; it shows no systematic regional change. Eastward, the modal increase in alkali feldspar and decrease in plagioclase (fig. 46) suggest that Al_2O_3 and CaO also increase eastward, and although the plots of both oxides are irregular, they confirm this expectation. The moving-average curve for Al_2O_3 decreases from 17 percent in the west to about 15 percent in the central part of the curve, then remains in that range eastward, despite wide variations. The moving-average curve for CaO decreases from about 6 percent in the west to about 2 percent in the central part of the curve but varies widely in the eastern part of the traverse.

The eastward decrease in the content of mafic minerals (fig. 46) indicates that collectively Fe_2O_3, FeO, and MgO decrease eastward, and the plots substantiate this (fig. 47). However, the decrease is not large and the moving-average curve for Fe_2O_3 remains relatively constant, ranging from 0.7 to 1.7 percent. The moving-average curve for FeO shows a decrease from 4 percent in the west to 1.0 to 1.5 percent in the central part of the curve, then varies erratically from 1.0 to 3.0 percent in the eastern part of the traverse. The moving-average curve for MgO similarly shows a decrease from almost 3.0 percent in the west to less than 1.0 percent in the central part of the traverse and varies from 1.0 to 2.0 percent in the eastern part. Dodge (1972) showed that the oxidation ratio [mol $(2Fe_2O_3\times100)/(2Fe_2O_3+FeO)$] ranges from 7 to 65 and increases eastward across the batholith. He attributed this variation to an eastward increase of oxygen fugacity caused by an increase in the water

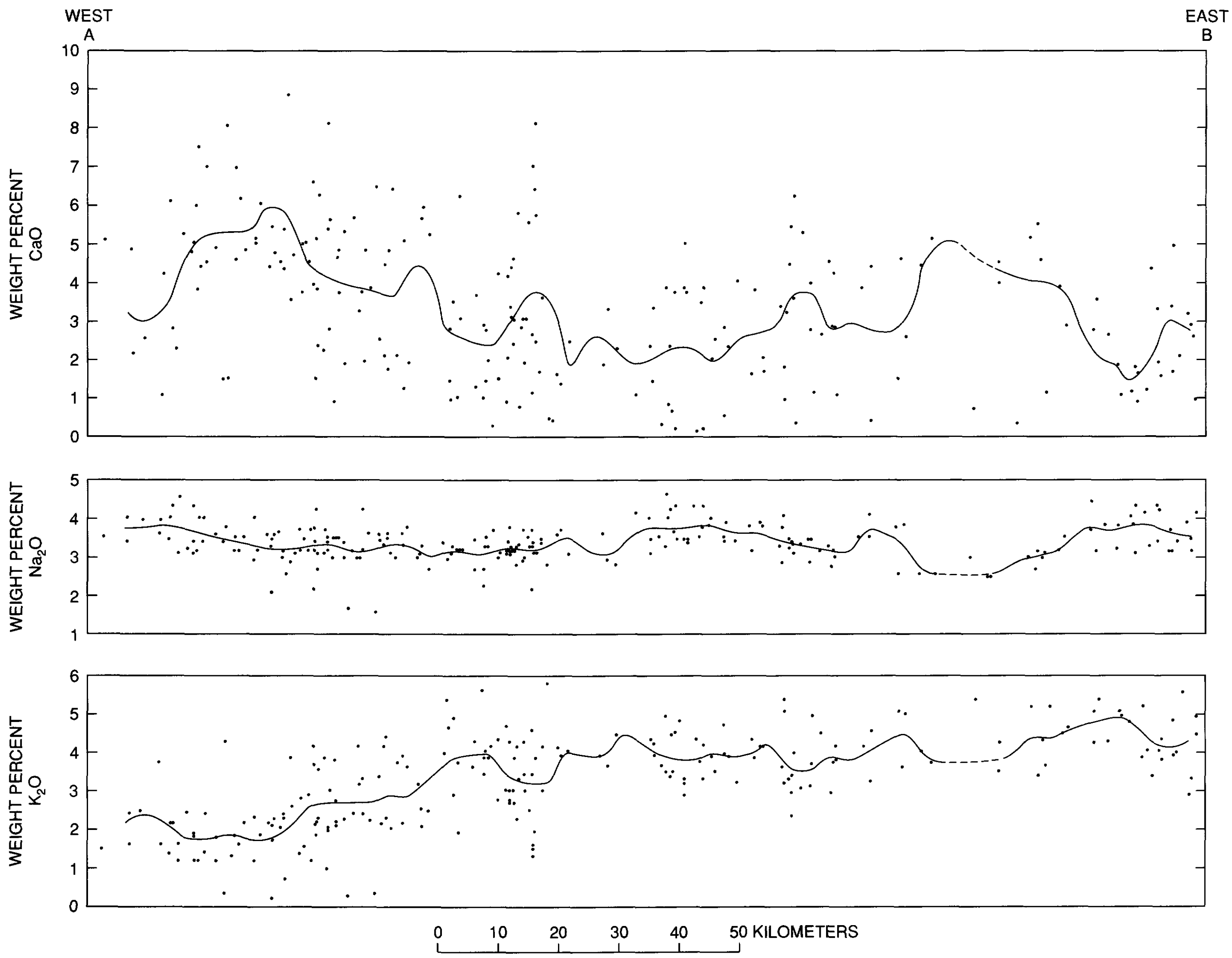


FIGURE 47.—Continued

content of the source materials. Isotopic data suggest that it reflects an eastward increase in the proportion of sedimentary crustal materials in the parent magma.

VARIATIONS IN THE ALKALI-LIME (PEACOCK) INDEX

Determination of the alkali-lime (Peacock) index—weight percent SiO_2 when CaO equals K_2O+Na_2O (Peacock, 1931)—is imprecise because of the scatter of data, and the results are thus unsatisfactory (fig. 50). Generally, the index decreases eastward across the batholith, ranging from about 63.5 in the Fine Gold Intrusive Suite to 55 in the Jurassic rocks of the White Mountains. However, the indices for the northern part of the Fine Gold Intrusive Suite, the intrusive suites of Yosemite Valley and Buena Vista Crest, and the Shaver and Scheelite Intrusive Suites are greater than 61; consequently, all these suites would be classed as calcic, even though most petrologists would consider them typical calc-alkalic suites.

VARIATIONS IN MINOR ELEMENTS

Uranium, thorium, rubidium, beryllium, tantalum, barium, and total rare-earth elements increase eastward across the Sierra Nevada (Wollenberg and Smith, 1968; Dodge and others, 1970; Dodge, 1972; Dodge and others, 1982). In support of heat-flow studies being carried out by A.H. Lachenbruch in the central Sierra Nevada, Wollenberg and Smith (1968) determined the uranium, thorium, and potassium contents of the granitic rocks of the central Sierra Nevada using gamma-ray spectrometry. More than 150 determinations were made, some in the laboratory but most in the field using a portable scintillation spectrometer. Both uranium and thorium increase eastward into the Mount Givens and Lamarck Granodiorites of the John Muir Intrusive Suite, then remain approximately constant eastward into the White Mountains. Uranium increases eastward from an average of 1.5 parts per million in the Fine Gold Intrusive Suite to 6.7 in the Mount Givens Granodiorite and 6.9 in the Lamarck Granodiorite before decreasing to 4.9 in the Tungsten

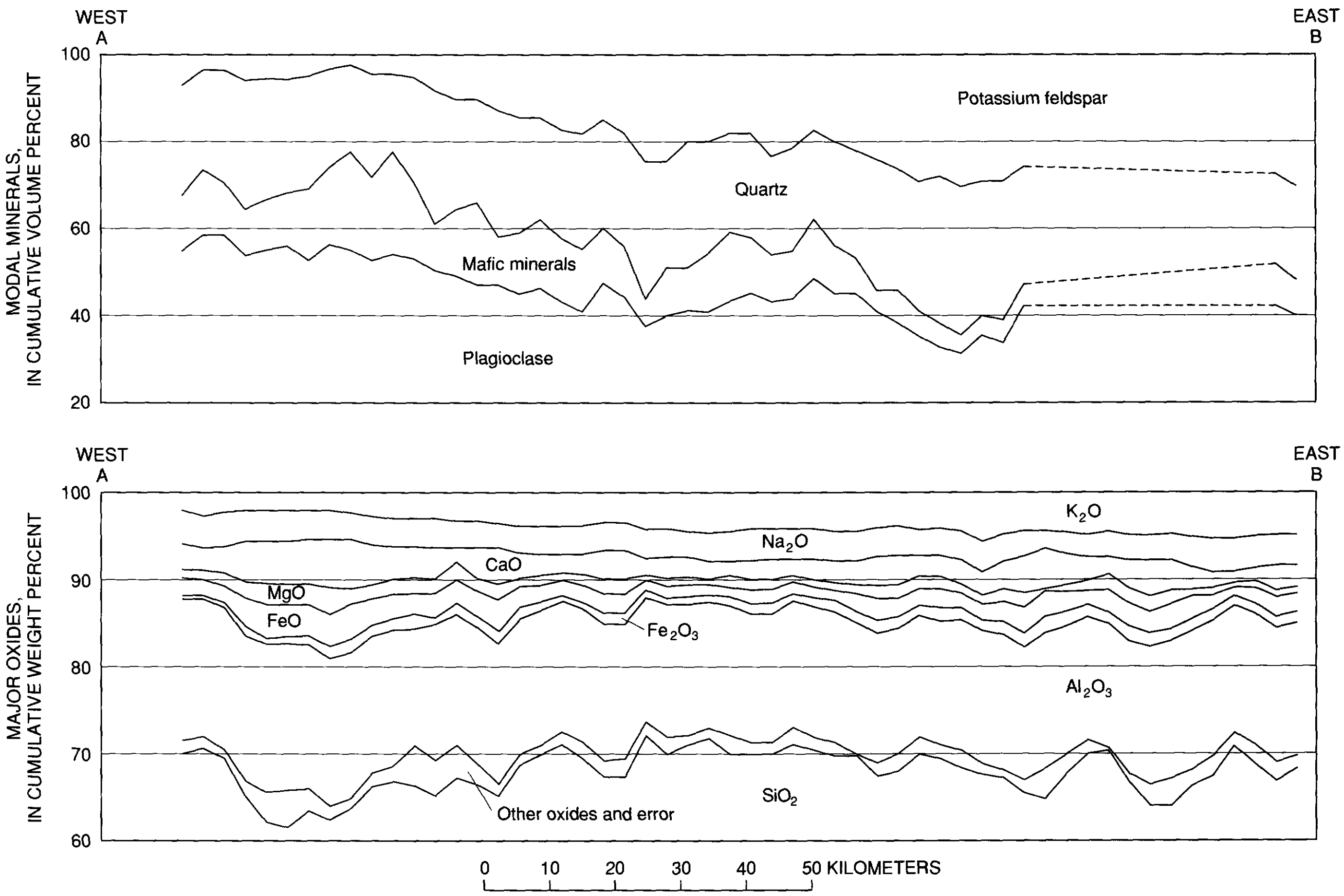


FIGURE 48.—Cumulative modal-mineral and major-oxide contents projected onto line A–B, figure 45. Curves for modes are dashed where alluvium conceals bedrock and no samples were collected.

Hills Granite of the Scheelite Intrusive Suite and the Tinemaha Granodiorite of the Palisade Crest Intrusive Suite. Thorium similarly increases from 4.3 parts per million in the Ward Mountain Trondhjemite of the Fine Gold Intrusive Suite to 21.5 in the Mount Givens and Tinemaha Granodiorites. The eastward-increasing amounts of uranium, thorium, and potassium affect radioactive-heat generation within the granitic rocks as shown in figure 57.

Dodge and others (1970), using semiquantitative spectrographic determinations, showed that rubidium generally increases eastward across the Sierra Nevada from less than 50 parts per million in the west to a wide spread of values that range from about 60 to about 230 parts per million in the east (fig. 51). Subsequently Dodge (1972) plotted semiquantitative data for other minor elements and found that beryllium increases eastward from an average of about 1 part per million in the west to about 3 parts per million in the east; although his plots of other elements are not definitive, they suggest that boron, barium, copper, lanthanum, niobium, lead, strontium, and zirconium increase eastward. Since these earlier determinations were made, neutron-activation analyses have confirmed the eastward increase of barium, tantalum, and rare-earth elements (fig. 51; Dodge and others, 1982).

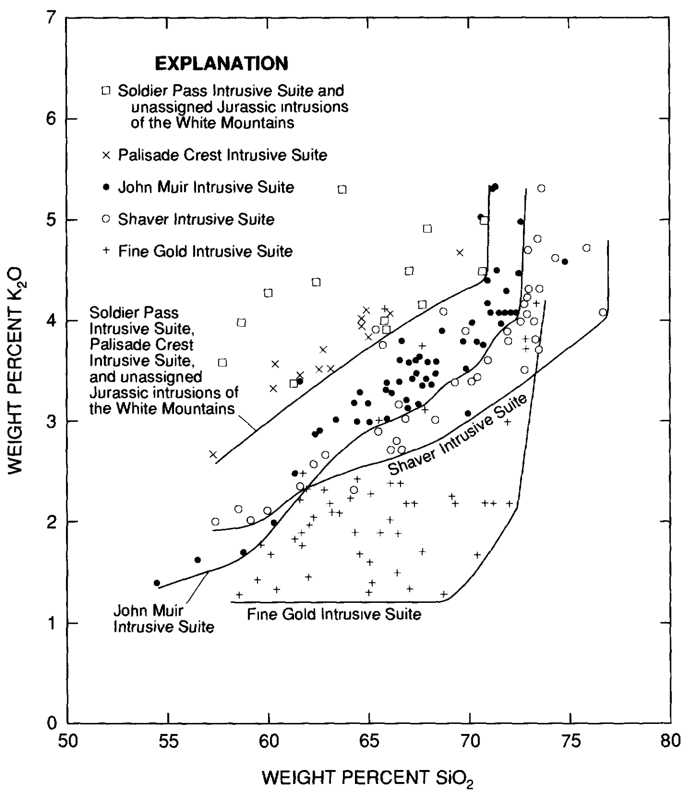


FIGURE 49.—Variation of K_2O with SiO_2 in representative intrusive suites. For each intrusive suite, solid curves are drawn at lower limit of weight percent K_2O (excluding aberrant data points). K_2O increases successively eastward in suites, whereas SiO_2 ranges from about 50 to 75 weight percent in all suites.

The neutron-activation analyses show that the rare-earth elements (REE), as expected, are enriched relative to chondritic abundances and that the light REE are much more enriched than the heavy REE (Frey and others, 1978; Dodge and others, 1982). Total REE increase gradually but irregularly across the batholith from an average of about 80 parts per million in the west to more than 160 in the east, but this increase is dominated by the light REE, especially lanthanum and cerium, which constitute from one-third to two-thirds of the total REE (fig. 52). A negative europium anomaly, seen only in leucogranites, presumably results from its prior removal from the feldspar of older cogenetic rocks. The Knowles Granodiorite, which is leucocratic but contains only small amounts of potassium feldspar, does not have a europium anomaly, presumably because crystallization of the older cogenetic Bass Lake Tonalite did not remove much potassium feldspar.

ISOTOPIC VARIATIONS

Considerable strontium isotopic data are available for rocks of the batholith; neodymium and oxygen isotopic ratios have also been studied but to a far lesser extent.

STRONTIUM

The study of strontium isotopes within the map area has been carried out almost exclusively by R.W. Kistler and his associates. Initial $^{87}Sr/^{86}Sr$ (the ratio at the time of crystallization) for samples of granitoids increases generally eastward from less than 0.704 in the western foothills to 0.707 in the High Sierra to more than 0.708 in the White Mountains (fig. 57). Kistler and Peterman (1973) attributed these regional variations to lateral changes across the Sierra Nevada in the composition of the source regions (mainly crustal) for the granitoid magmas—from eugeoclinal and upper-mantle material in the west to miogeoclinal deposits in the east. They suggested that the value of 0.706 is the lower limit for regions underlain by Precambrian and miogeoclinal Paleozoic crust, that 0.706 and 0.704 are the eastern and western limits, respectively, of eugeoclinal Paleozoic crust, and that 0.704 is the upper limit for regions underlain by oceanic crust. Initial $^{87}Sr/^{86}Sr$ probably also reflects an eastward-increasing proportion of crustal-to-mantle material in the parent granitoid magmas as the crust thickens and a lateral change in the composition of the crustal materials. Values of 0.7032 reported for a body of gabbro and for a body of trondhjemite that lie west of the map area indicate a predominantly mantle source for those magmas, whereas initial $^{87}Sr/^{86}Sr$ ranging from 0.71118 to 0.71210 for the peraluminous granite of Birch Creek and from 0.71031 to 0.71204 for the peraluminous megacrystic granite of Papoose Flat in

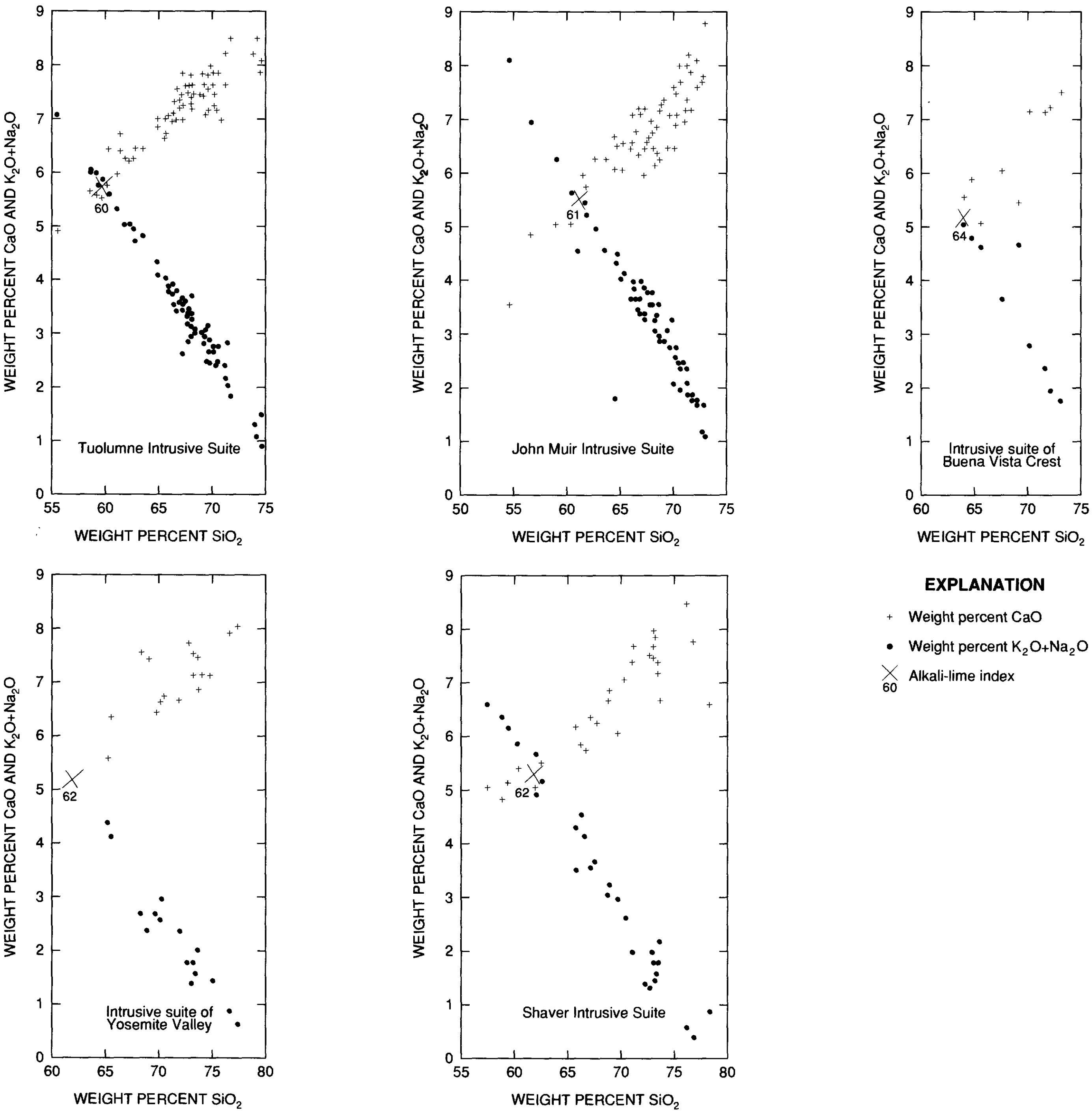


FIGURE 50.—Alkali-lime (Peacock) index values for intrusive suites. Index is weight percent SiO_2 where weight percent CaO equals weight percent K_2O+Na_2O (Peacock, 1931). Figure continued on following page.

the White Mountains (R.W. Kistler, written commun., 1988) clearly indicate a predominantly crustal source.

Data have been reported that indicate initial $^{87}Sr/^{86}Sr$ varies in some intrusions from one part to another. The clearest example of such variation is the inward increase of initial ratios within outer parts of the Tuolumne Intrusive Suite from 0.7057 to 0.7065 (Kistler and others, 1986). Additional data to show the vertical and lateral distribution of initial $^{87}Sr/^{86}Sr$ within other intrusions and intrusive suites are much needed because of their importance in interpreting the genesis of the granitoid magmas and in evaluating the reliability of ages determined from isochrons.

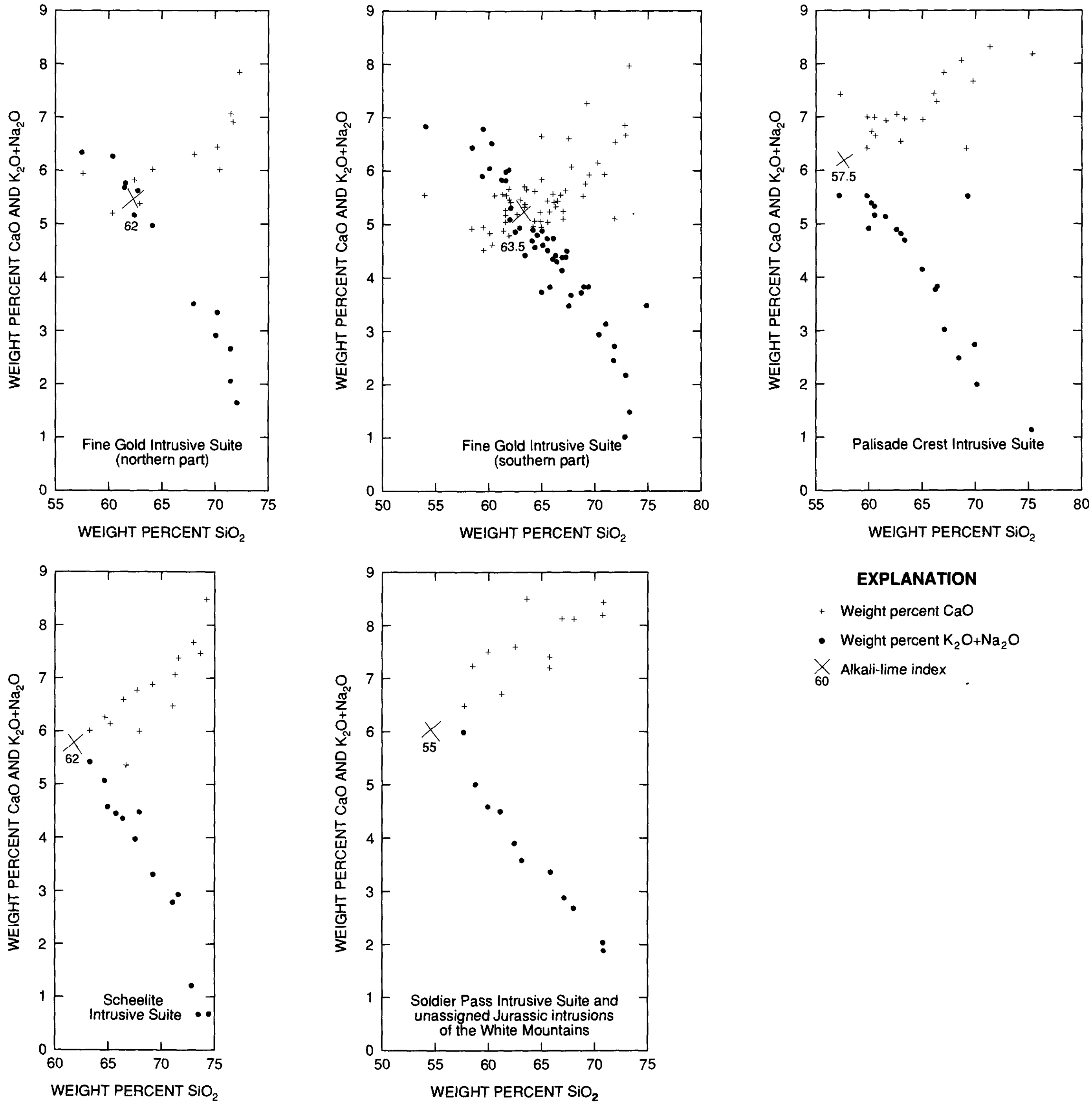


FIGURE 50.—Continued

Although changes in both the composition and isotopic characteristics of the crust and the proportions of crustal and mantle components affected the isotopic and compositional characteristics of the granitic rocks, it is nevertheless interesting to consider some crude computations in which the granitic rocks are assumed to result from the mixing of a mantle component of fixed isotopic composition with crust of variable isotopic composition. For these computations, initial $^{87}Sr/^{86}Sr$ is assumed to be 0.703 for the mantle component, slightly less than the lowest initial strontium ratio determined for plutonic rocks of the central Sierra Nevada, and to vary within and below the range in samples of exposed Precambrian and Paleozoic country rocks within and east of the Sierra Nevada for the crustal component. Measured crustal ratios of these rocks, adjusted to 100 Ma (when the most abundant granitic magmas were generated), range from 0.7120 to 0.7270 (table 2) and are similar to the initial ratios of garnet-bearing two-mica granites in the Ruby Mountains of eastern Nevada, which Kistler and others (1981) interpreted to be derived from crustal materials. If the exposed rocks are representative of the crust as a whole, values in the range of 0.71 to 0.73 are reasonable, but if the lower crust was underplated with basalt introduced from the mantle, the average strontium ratio of the crust at the time of magma generation could have been significantly lower.

Assuming initial ratios of 0.703 for the mantle component and 0.713 for the crustal component, initial $^{87}Sr/^{86}Sr$ of 0.704 for a granitic rock indicates 10 percent crustal component, 0.706 indicates 30 percent, and 0.708 indicates 50 percent. If the higher value of 0.723 (the average of measured values for Paleozoic and Precambrian rocks in and east of the Sierra Nevada) is assumed for the

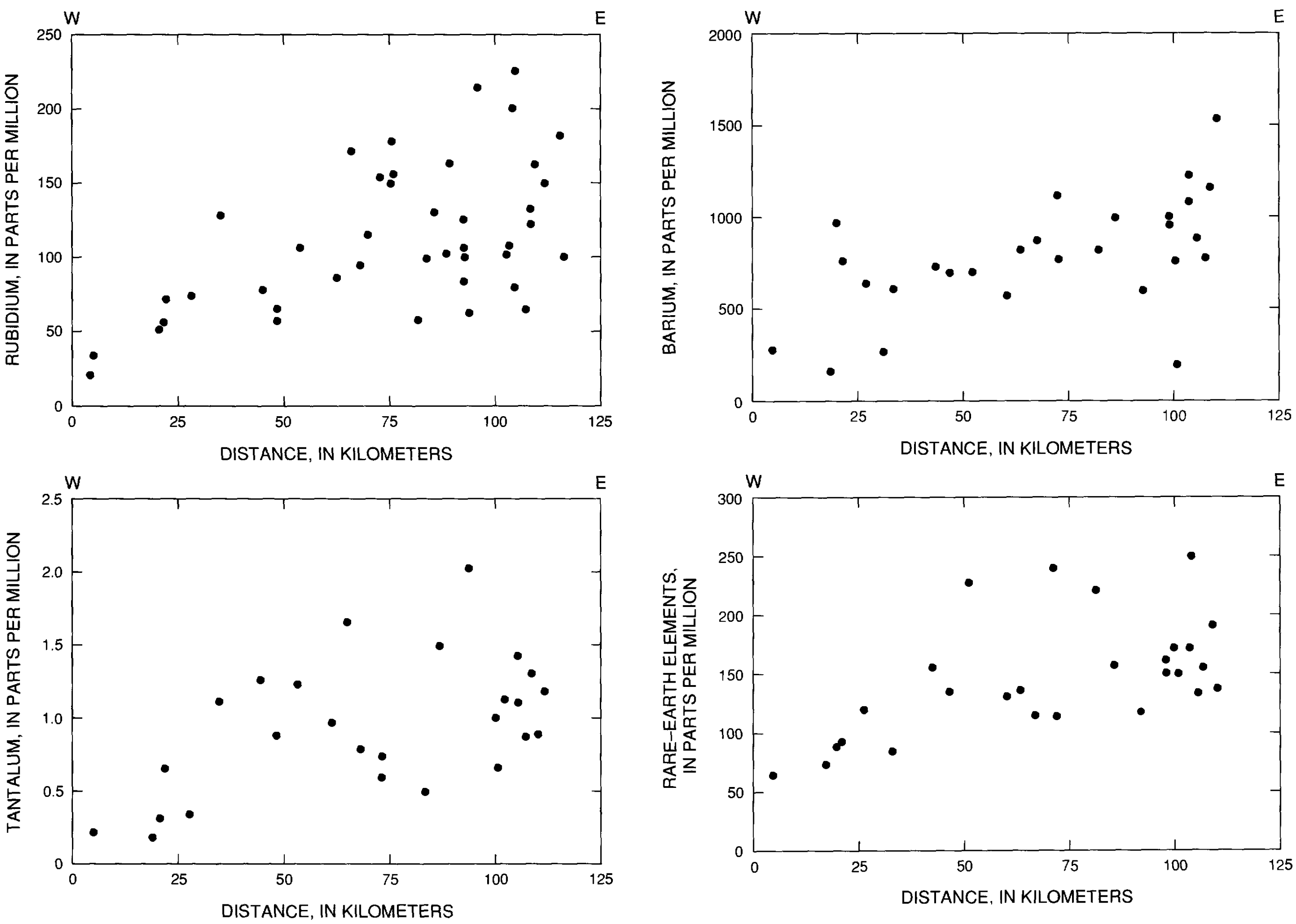


FIGURE 51.—Variation in selected minor elements east-west across Sierra Nevada. Rubidium plot modified from Dodge and others (1970) and barium, tantalum, and rare-earth element plots modified from Dodge and others (1982).

crustal component, the initial ratio of 0.704 indicates 5 percent crustal component, 0.706 indicates 15 percent, and 0.708 indicates 25 percent. However, if the crustal component is assumed to include less-radiogenic material from the lower crust and to have an average initial ratio of 0.708, 0.704 indicates 20 percent crustal component, 0.706 indicates 60 percent crustal component, and 0.708 indicates 100 percent crustal component.

NEODYMIUM

DePaolo (1980, 1981) determined initial $^{143}Nd/^{144}Nd$ values for many of the same samples as were used by Kistler and Peterman (1973) and converted both the neodymium values and the initial $^{87}Sr/^{86}Sr$ values to $\varepsilon Nd_{(T)}$ and $\varepsilon Sr_{(T)}$, which are 10^4 times the ratios of these values to the ratios in a model chondritic reservoir at the time of emplacement of the intrusion, minus 1. Thus the calculated e value is 0 when the measured ratio is equal to the chondritic ratio. A plot of $\varepsilon Nd_{(T)}$ against $\varepsilon Sr_{(T)}$ yields a hyperbolic curve. The $\varepsilon Nd_{(T)}$ decreases eastward across the batholith from +6.5 to −7.6, whereas $\varepsilon Sr_{(T)}$ increases from −16 to +54. Those $\varepsilon Nd_{(T)}$ values greater than 0 are close to the range of $\varepsilon Nd_{(T)}$ values for island arcs, and DePaolo (1981) interpreted rocks with these values to be derived largely from depleted mantle material. An $\varepsilon Nd_{(T)}$ value of 0 closely corresponds to the $^{87}Sr/^{86}Sr$ value 0.704, which Kistler and Peterman (1973) interpreted to be the upper limit for Paleozoic eugeoclinal crust. DePaolo (1980) postulated that rocks in which $eNd_{(T)}$ is less than 0 contain crustal material in amounts roughly proportional to the decrease in $\varepsilon Nd_{(T)}$ and increase in $\varepsilon Sr_{(T)}$.

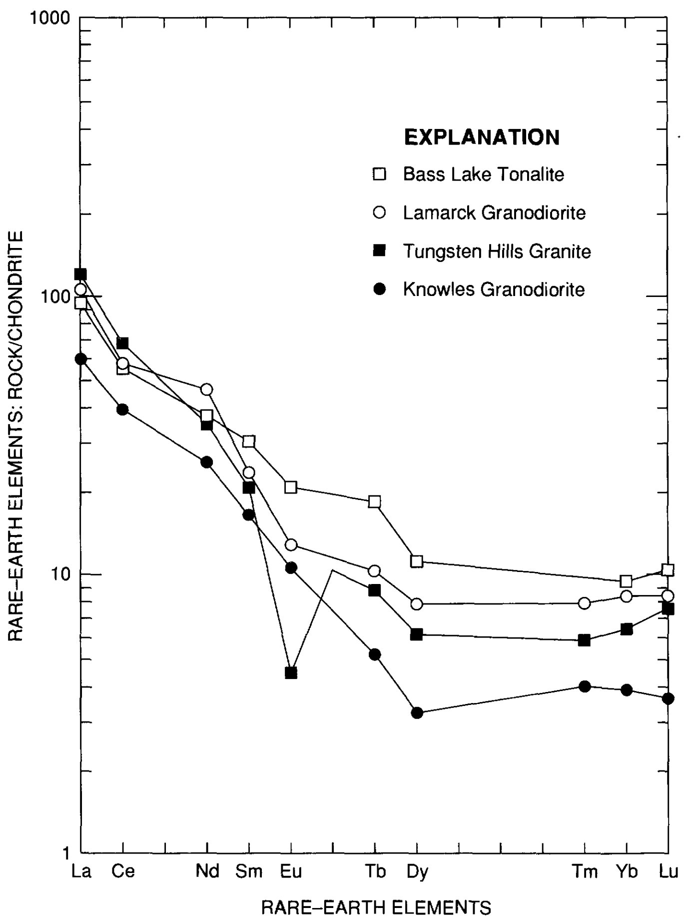


FIGURE 52.—Chondrite-normalized rare-earth element patterns for representative granitoids of central Sierra Nevada. Modified from Dodge and others (1982).

Determinations by Domenick and others (1983) of isotopic ratios for upper-mantle and lower-crustal xenoliths from trachyandesite that intrudes the Dinkey Creek Granodiorite and for a basanitoid lava flow in the eastern Sierra Nevada south of the map area suggest that, in addition to these west-to-east isotopic variations, systematic variations occur with depth. The $\varepsilon Sr_{(T)}$ and $\varepsilon Nd_{(T)}$ values for the xenolith samples cover the full range of values for granitoid samples collected across the batholith and lie on the same hyperbolic curve of $\varepsilon Sr_{(T)}$ plotted against $\varepsilon Nd_{(T)}$. The xenoliths from the trachyandesite have $\varepsilon Sr_{(T)}$ values that range from −10.2 to +382.7 and $\varepsilon Nd_{(T)}$ values that range from +4.8 to −17.4. The xenolith from the basanitoid lava flow extends the range of $\varepsilon Sr_{(T)}$ values down −24.2 and of $\varepsilon Nd_{(T)}$ values to +8.3. The values for the Dinkey Creek Granodiorite are within the ranges of values for the xenoliths: $\varepsilon Sr_{(T)}$=+44.5 and $\varepsilon Nd_{(T)}$=-7.7. Interpretation of the probable depth at which the samples attained their present mineral assemblages suggests that $\varepsilon Sr_{(T)}$ may decrease and $\varepsilon Nd_{(T)}$ increase downward as they do from east to west at the surface across the batholith.

OXYGEN

Masi and others (1981) reported that the range of oxygen isotope ratios (standardized values, $\delta^{18}O$) for granitoids of the Sierra Nevada is 5.5 to 12.3 per mil and that 80 percent of these values are within the range 7.0 to 10.0 per mil. The value $\delta^{18}O$ is defined as $10^3\times(^{18}O/^{16}O_{sample}-^{18}O/^{16}O_{SMOW})/^{18}O/^{16}O_{SMOW}$, where SMOW is standard mean ocean water. The distribution of $\delta^{18}O$ values is quite irregular, but within the map area (pl. 1) $\delta^{18}O$ appears to decrease slightly toward the east. Masi and others (1981) concluded that most values for Sierra Nevada rocks are primary and are related to the compositions of the source materials for the magmas.

RELATIONS OF VOLCANISM AND PLUTONISM

Although both volcanism and plutonism occurred during the Mesozoic, field observations and isotopic dating provide only limited support for the reasonable assumptions that they are coeval and that the plutons were reservoirs for volcanoes. The intrusive suites of Washburn Lake, Merced Peak, and Buena Vista Crest appear to be related to the Minarets volcanic sequence, the quartz monzonite of Mount Barcroft is clearly cogenetic with the volcanic strata it intrudes, and relatively small hypabyssal intrusions are associated with most masses of metavolcanic rocks in the Sierra Nevada. However, other correlations between intrusive suites and metavolcanic sequences are dubious at best. Furthermore, except for a few small felsic intrusions such as the Johnson Granite Porphyry and the granite of Hogan Mountain, none of the plutonic rocks show textural evidence of an eruptive phase. However, in regions less deeply eroded than the Sierra Nevada, such as the San Juan Mountains of Colorado (Lipman, 1975) or the southern Andes of central Chile (Hildreth and others, 1984), volcanism and plutonism have been shown to be closely related. In these areas, magmatism began with andesitic and basaltic eruptions from small centers and was followed by the outpouring of voluminous and widespread but easily erodible ash-flow sheets. The ash-flow sheets were erupted from calderas that were underlain by silicic-magma chambers comparable in size to large Sierran plutons. Wes Hildreth (written commun., 1986) suggested that the small masses of diorite and gabbro so widely distributed in the Sierra Nevada, especially in the margins of the batholith, were feeders for early mafic volcanism.

Isotopic age data indicate that magmatism in the central Sierra Nevada was more or less continuous from about 210 Ma (Late Triassic) to about 85 Ma (early Late Cretaceous), but the distribution of ages suggests that the peaks of volcanic activity alternated with peaks of plutonic activity (figs. 53 and 54). Alternating peak activity of plutonism and volcanism could reflect a changing state of strain—volcanism dominating during times of extension, when magmas could flow freely to the surface, and plutonism dominating during times of little or no extension, or even compression, when mantle-derived mafic magmas would pond and mix with crustal material before rising into the upper crust.

Most of the volcanic rocks in the map area are of Jurassic or Triassic age, whereas the most voluminous plutonic rocks are predominantly Cretaceous. This relationship probably resulted chiefly from uplift and deep erosion that accompanied and followed emplacement of the Cretaceous plutonic suites. The easily erodible ash-flow sheets of Cretaceous age were completely removed, whereas the steeply tilted Jurassic volcanic rocks were protected. The presence of volcanic detritus in the Late Jurassic and Cretaceous Great Valley sequence to the west and of voluminous ash falls in Jurassic and Cretaceous strata to the east (in the direction of the prevailing winds) is evidence of rampant volcanism during the timespan of plutonism.

Dickinson and Rich (1972) investigated the modal abundances of quartz and feldspar, representing plutonic sources, and lithic fragments, representing volcanic sources, in medium- to coarse-grained sandstones of the Great Valley sequence. Although their data do not provide a record of silicic ash-flow deposits, which would have been the most voluminous eruptive rocks, peak abundances of lithic fragments in the Late Jurassic (Tithonian) and early Late Cretaceous (Cenomanian) correspond to magmatic activity in the Sierra Nevada and Klamath Mountains during those times.

Cobban (1972) stated that nearly every unit of the Cretaceous section of the Rocky Mountains contains evidence of contemporaneous volcanic activity. Beds of bentonite, bentonitic shale, and bentonitic cement in sandstone are common, and some porcelaneous marine shale, especially the Mowry Shale, is enriched in silica as a result of ash fall. Bentonite beds have been dated isotopically at 110, 105, and 79 Ma, and the Mowry Shale at 112 to 110 Ma; these ages are in the general range of magmatic activity in the Sierra Nevada, although they do not correspond exactly to the magmatic episodes that have been identified within the map area. Although Cobban (1972) suggested that most of the volcanic material came from closer sources, the Sierra Nevada is the most likely source for large volumes of siliceous ash.

Uncertainty as to the reliability of the isotopic ages, especially for metavolcanic rocks, is a major difficulty in interpreting the relative ages of the volcanic and plutonic rocks. Disagreement between U-Pb and Rb-Sr ages for the same volcanic units can be quite large; only for the Minarets sequence do the ages by both methods agree (fig. 53). For the oldest sequences, differences generally increase with the increasing ages of the strata. Thus, U-Pb ages for the older succession of the Koip sequence range from 214

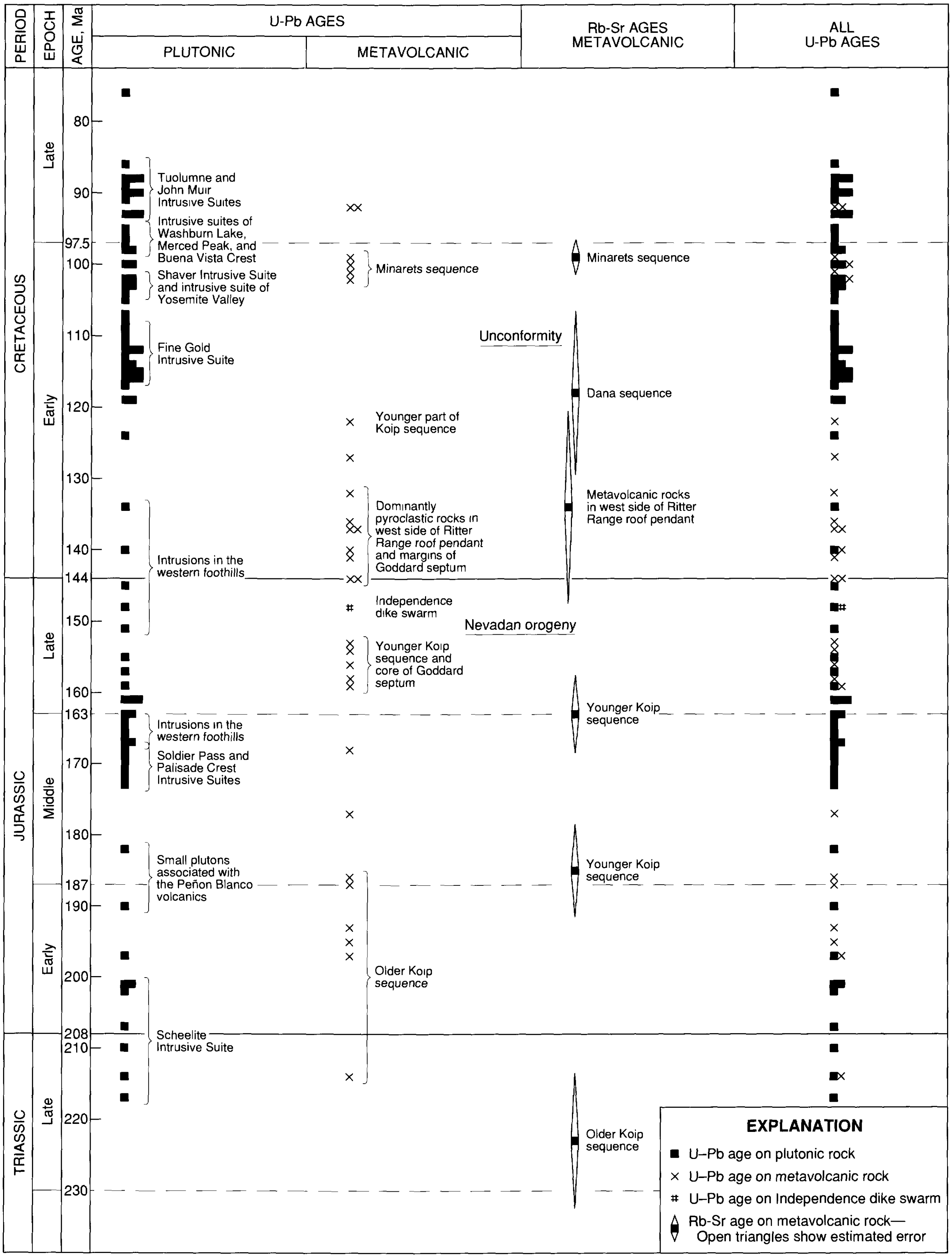


FIGURE 53.—U-Pb and Rb-Sr isotopic ages of volcanic and plutonic rocks within Mariposa 1° by 2° quadrangle (pl. 1) plotted in accordance with Decade of North American Geology time scale (Palmer, 1983). Includes data from Sylvester, Miller, and Nelson (1978), Gillespie (1979), Stern and others (1981), Kistler and Swanson (1981), Chen and Moore (1982), Dodge and Calk (1986), Hanson and others (1987) and unpublished data from R.S. Fiske, R.W. Kistler, T.W. Stern, and O.T. Tobisch (oral and written communs., 1981–86).

to 186 Ma, whereas the Rb-Sr whole-rock age for the same rocks is 237±11 Ma, using only samples collected from within the map area, or 224±14 Ma if additional samples from north of the map area are included (Kistler and Swanson, 1981). Another problem is that the U-Pb ages for volcanic rocks do not always agree with the order indicated by the positions of the sampled units within apparently unbroken stratigraphic successions. Unless these successions are repeated along unrecognized faults, the deviation of measured U-Pb ages from true ages is significantly greater for volcanic rocks than for plutonic rocks.

The most convincing pairings of volcanic sequences with plutonic suites are the ~100-Ma Minarets sequence with the ~98-Ma intrusive suite of Merced Peak and the Jurassic quartz monzonite of Mount Barcroft with the volcanic rocks it intrudes, some of which have Cretaceous isotopic ages. Confinement of the fragmented western part of the Minarets sequence to the area of the granodiorite of Jackass Lakes strongly suggests that the granodiorite is a later manifestation of the same magmatic event that produced the Minarets caldera. Close spatial and temporal association of the Minarets sequence with the intrusive suites of Washburn Lake and Buena Vista Crest indicates that they belong to the same magmatic episode.

Another likely pairing of a plutonic suite with a metavolcanic sequence is the Scheelite Intrusive Suite and the contiguous, older part of the Koip sequence—the oldest rocks of each type in the central Sierra Nevada. Although most of the isotopic ages for the Scheelite Intrusive Suite are older than those for the older Koip sequence, Kistler (1966a) interpreted a contact between the ~210-Ma granite of Lee Vining Canyon and the Koip sequence to be intrusive, and Bateman (1965) also interpreted a contact between the ~200-Ma Tungsten Hills Granite and metavolcanic rocks in the southern part of the Pine Creek septum as intrusive. The relations of these two groups of rocks to each other merits further examination.

SHIFTS IN THE LOCUS OF MAGMATISM

In figure 54, the U-Pb isotopic ages are arranged in columns to show the distribution of rocks and their U-Pb ages from west to east across the map area (pl. 1). Only one locus of magmatism is indicated at any one time during the Cretaceous, and it shifted eastward several times during the interval 114 to 85 Ma. The locus of magmatism also shifted from time to time during the Jurassic, but it is less certain that only one locus existed at any one time during that period. Nevertheless, in figure 54, the arrows indicating shifts in the locus of magmatism during the Jurassic are drawn on the assumption that only one locus existed at any time and ignore the distinct possibility that magmatism in the western metamorphic belt was continuous. The most likely causes of the shifts in the locus of magmatism are changes in the nature of activity along the convergent western boundary of the North American plate.

The principal but not the only cause of uncertainty as to whether one or two magmatic loci existed simultaneously during much of the Jurassic is a sparsity of information as to the ages of the volcanic rocks in the western foothills. Whether volcanism in the western foothills was more or less continuous or confined to relatively brief timespans in the Early Jurassic and in the late Middle or Late Jurassic is not known. Clark (1964) reported sparse late Middle or Late Jurassic (Callovian to Oxfordian or early Kimmeridgian) ammonites from the western part of the greenstone of Bullion Mountain and from the Logtown Ridge Formation, northwest of the map area, and Bateman and others (1985) reported an Early Jurassic Rb-Sr age for samples from the eastern part of the greenstone of Bullion Mountain. Early Jurassic U-Pb ages for two small plutons west of the map area, which intrude the Peñon Blanco Volcanics, also indicate the presence of Early Jurassic or older volcanics in the western foothills. If volcanism was more or less continuous in the western foothills during the Jurassic, two magmatic loci could have existed simultaneously, one in the east and one in the west, as proposed by Schweickert and Cowan (1975), but if volcanism was confined to certain intervals in the Early and Late Jurassic, only one locus that shifted position from time to time may have existed. More than one locus is most likely to have existed when the ~163-Ma tonalite of Granite Creek (Stern and others, 1981) in the western foothills and the approximately coeval quartz monzonite of Mount Barcroft in the White Mountains were emplaced, when the younger part of the Koip sequence and the Late Jurassic volcanic rocks of the western foothills were erupted, when the dominantly pyroclastic strata of Early Cretaceous age were erupted in the High Sierra, and when plutons were emplaced in the western foothills. Aberrant isotopic ages for the Palisade Crest, Scheelite, and Soldier Pass Intrusive Suites obscure age relations and erroneously suggest overlapping magmatism in the Late and Middle Jurassic, during the interval from 177 to 155 Ma (fig. 54).

Despite a dearth of evidence of compressive regional deformation during the emplacement and solidification of the intrusive rocks, the country rocks appear to have been deformed repeatedly during the general period of intrusion. One possible explanation is that the intruding magmas deformed the country rocks but

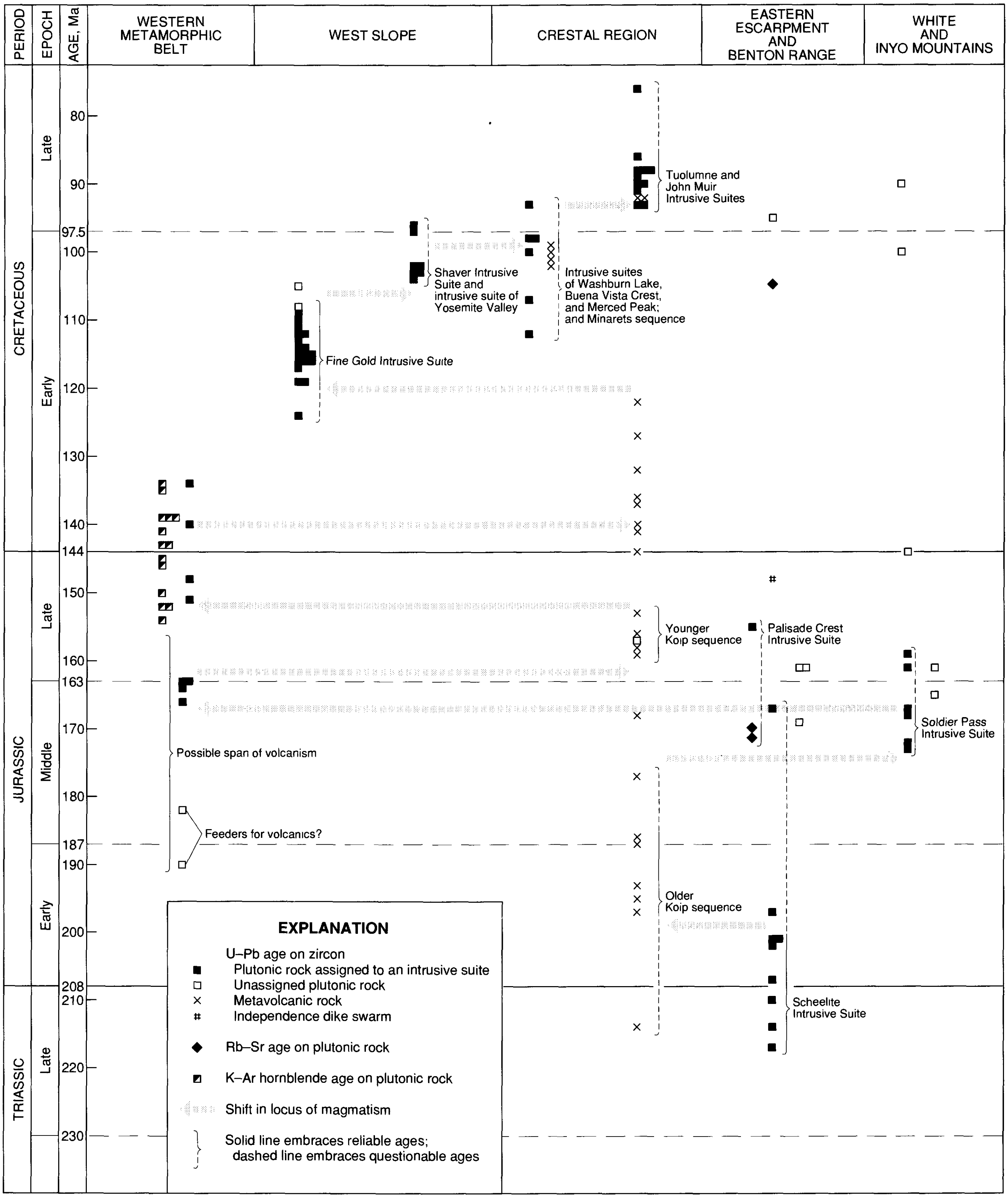


FIGURE 54.—Isotopic ages plotted by zones across Mariposa 1° by 2° quadrangle (pl. 1) to show loci of magmatism through time. Data from Evernden and Kistler (1970), Sylvester, Miller, and Nelson (1978), Gillespie (1979), Stern and others (1981), Chen and Moore (1982), Dodge and Calk (1986), and Hanson and others (1987). Time scale from Palmer (1983).

came to rest before crystallization had advanced far enough for the intrusions to deform themselves. Alternatively, events violent enough to have caused the locus of magmatism to shift could also have caused the deformation in the country rocks. If so, the deformations probably immediately preceded the intrusions resulting from the shift; older, solidified intrusions presumably would have acted as buttresses.

SUBSURFACE STRUCTURE

Interest in the deep structure beneath the Sierra Nevada began in the 1930's when A.C. Lawson published "The Sierra Nevada in the Light of Isostasy" (Lawson, 1936). He used average crustal and upper-mantle densities in conjunction with the principles of isostasy to estimate a crustal thickness of about 68 km in the vicinity of Mount Whitney. In a comment on Lawson's report, Byerly (1938) confirmed the presence of the Sierran root on the basis of delay in the arrival times of earthquake waves at seismograph stations east of the Sierra Nevada from origins west and northwest of the range. Seismic and gravity studies have since shown that the base of the crust is depressed beneath the Sierra Nevada and forms an asymmetric north-trending trough that parallels the axis of the batholith.

SEISMIC DATA

Seismic studies provide the principal evidence for the interpretation of the deep structure of the Sierra Nevada, including the shape of the Sierran root. Seismic-refraction measurements show that the crust beneath the Sierra Nevada is layered and that the Moho has the form of an asymmetric trough that has its axis beneath the Late Cretaceous intrusions of the eastern Sierra Nevada at a depth of about 50 km. From this axis, the base of the crust rises westward to less than 20 km beneath the east side of the Great Valley of California and much more steeply eastward to less than 35 km beneath the Basin and Range province.

The locations of the sections of a fence diagram showing variations in seismic velocities and of several other features that bear on the deep structure of the Sierra Nevada are shown in figure 55. In detail, the fence diagram (fig. 56) may be incorrect, but it is the best available approximation of the gross seismic structure of the Sierra Nevada. The principal control is section *A–A'*, which closely follows the Sierran crest and parallels the gross structure (Eaton, 1966). This section is based on reversed measurements along the profile of refracted seismic waves that were created by exploding chemical charges in Shasta Lake (northern California), in Mono Lake, and near China Peak (south of the section). Transverse sections *B–B'* and *C–C'* were constructed with reference to this section from other seismic measurements, including two seismic profiles, one of which coincides with section *B–B'*.

The fence diagram shows that the Moho is depressed beneath the eastern High Sierra to a depth of a little less than 50 km at the latitude of Lake Tahoe and to a little more than 50 km at the latitude of Bishop. Lines connecting depths of 45 km to the Moho on sections *B–B'* and *C–C'* (fig. 55) show that in the northern half of the Sierra Nevada the root strikes about N. 30° W. and that the axis closely follows the eastern escarpment of the Sierra Nevada. In this construction, the west half of the Sierran root underlies Owens Valley and other down-faulted blocks east of the Sierra Nevada.

Carder and others (1970) and Carder (1973) disputed the preceding configuration and concluded that unless the material of the root has an anomalously high seismic velocity, the thickness of the crust under the Sierra Nevada is about 30 km, thickening eastward to 35 to 40 km under Owens Valley and the White Mountains. Pakiser and Brune (1980) reevaluated the data of Carder and others (1970) and Carder (1973) and, drawing on seismic-reflection and additional seismic-refraction data, concluded that Eaton's interpretations (Eaton, 1963, 1966; Bateman and Eaton, 1967) are generally correct. However, Pakiser and Brune (1980) concluded that the root is more strongly asymmetric than Eaton's data indicated and that the east flank rises steeply just east of the Sierra Nevada. This modification of the Moho is shown as an alternative in profiles *B–B'* and *C–C'* of figure 56.

In the central Sierra Nevada, above the axis of the root, *P*-wave velocities increase downward from 6.0 km/s at <12-km depth to 6.4 km/s at about 24-km depth, 6.9 km/s at the base of the crust, and 7.9 km/s in the upper mantle (Bateman and Eaton, 1967). A general increase in *P*-wave velocity with depth (but with numerous reversals) is interpreted to indicate an overall increase of dense mineral phases with depth. Bateman and Eaton (1967) suggested that such rocks as gabbro, amphibolite, and epidote amphibolite, having densities of 3.0 to 3.2 g/cm^3, produce *P*-wave velocities of 7.0 km/s in the upper part of the lower crust and that pyroxenite and eclogite, having densities of 3.2 to 3.4, produce *P*-wave velocities of 7.5 to 7.9 km/s at the base of the crust.

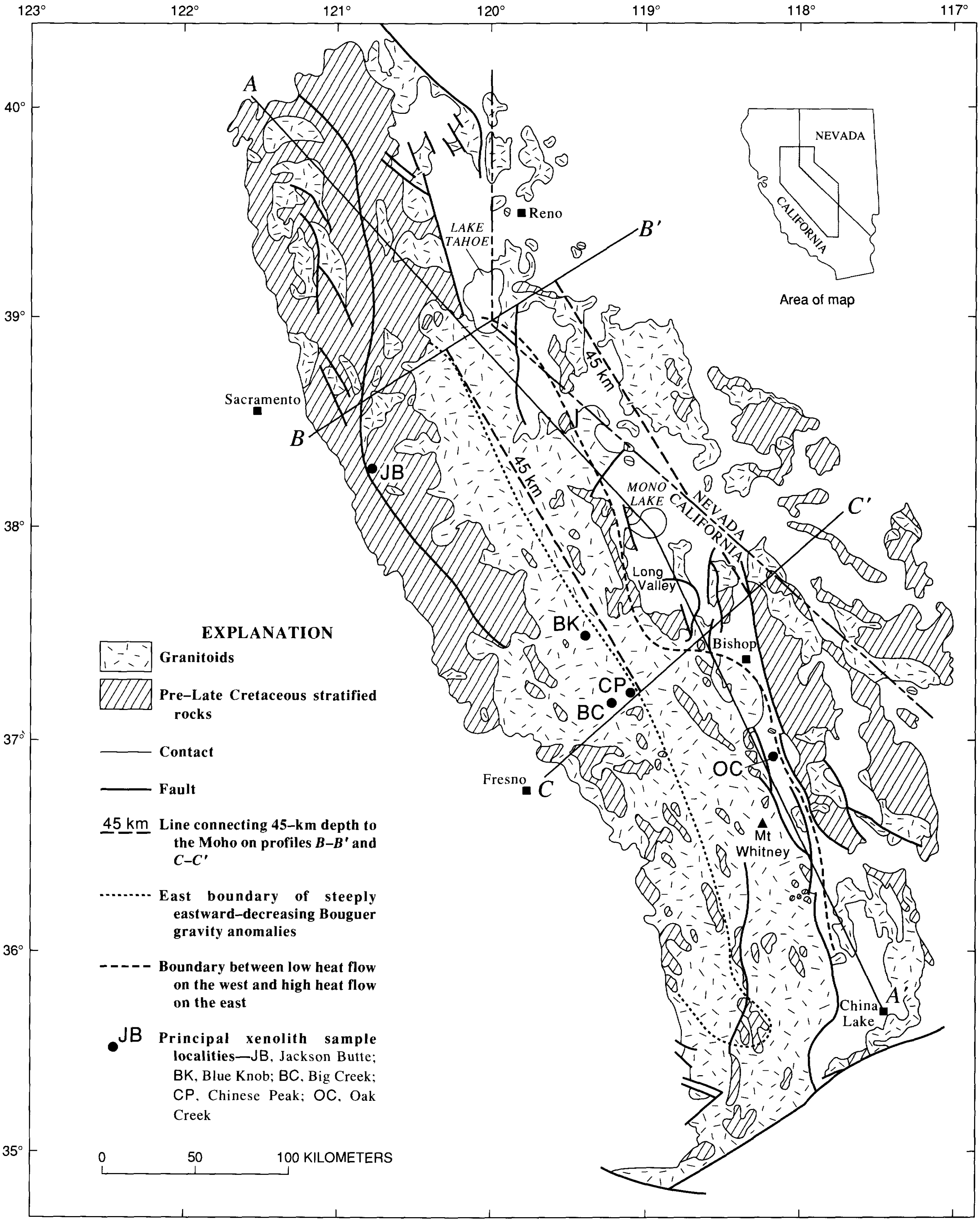


FIGURE 55.—Generalized geology of Sierra Nevada with locations of seismic-refraction sections A–A', B–B', and C–C' used in construction of fence diagram (fig. 56), sample localities of xenoliths collected from Cenozoic volcanic rocks, and gravity-anomaly and heat-flow boundaries.

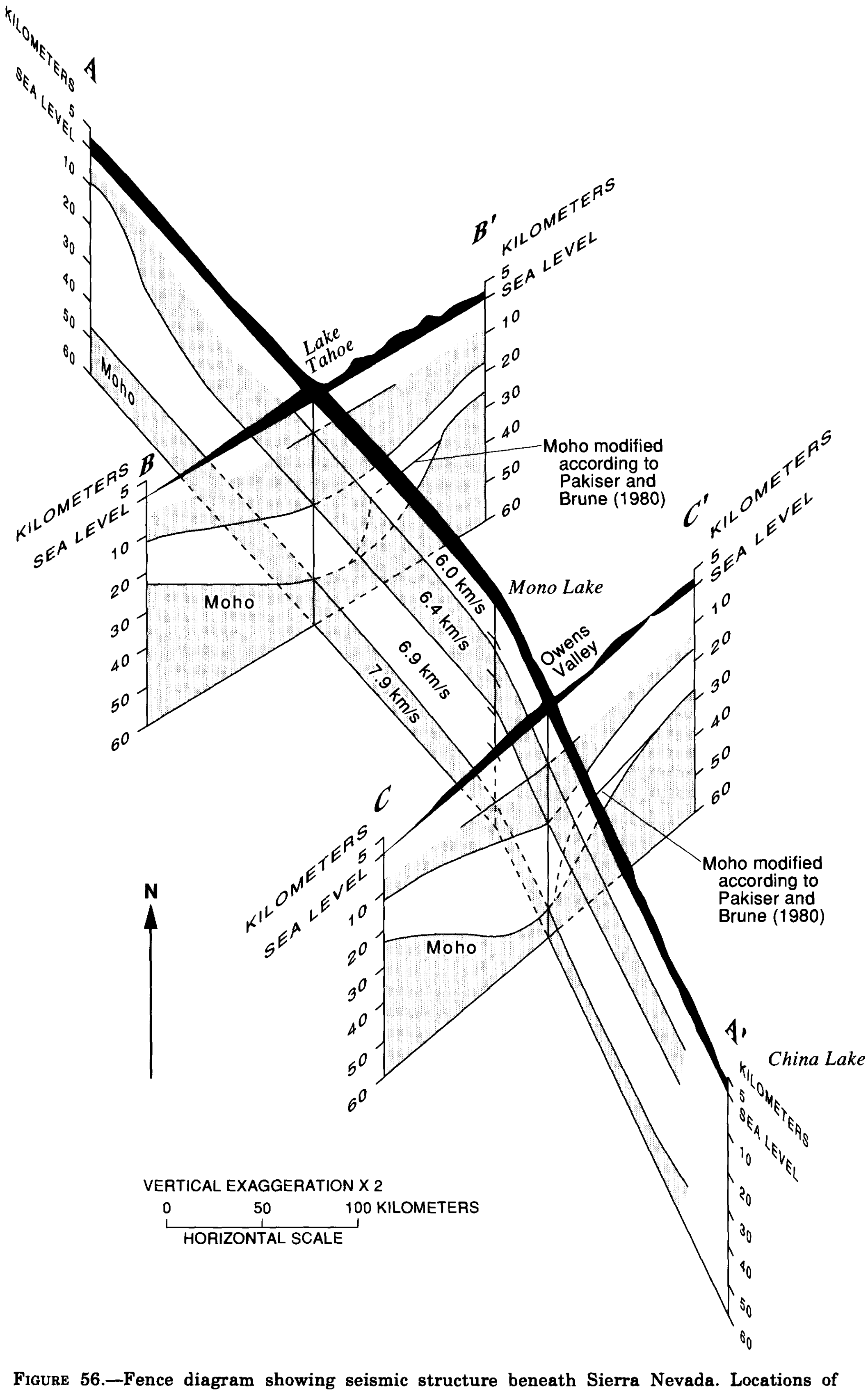


FIGURE 56.—Fence diagram showing seismic structure beneath Sierra Nevada. Locations of seismic-refraction sections shown in figure 55. Seismic-refraction profile data from Eaton (1963, 1966) and Bateman and Eaton (1967).

GRAVITY DATA

Gravity data also show that the base of the crust beneath the Sierra Nevada has the form of an asymmetric trough. However, on the basis of gravity data, Oliver (1977) suggested that the axis of the trough lies 30 km west of that shown by Bateman and Eaton (1967), who had positioned theirs on the basis of seismic data. The east edge of a zone of steeply eastward-decreasing contours on Bouguer gravity anomalies (Oliver, 1977) runs the length of the Sierra Nevada and between the seismic profiles *A–A′* and *B–B′* (fig. 55) coincides with the west boundary of the seismic root at 45 km. The break in slope of the gravity contours continues southward about half way between the east and west sides of the Sierra Nevada and presumably defines the west side of the root in the southern Sierra Nevada.

Both Bouguer gravity and isostatic-gravity anomalies (fig. 57) for an assumed constant density of 2.67 g/cm^3 decrease generally eastward to Owens Valley, then rise abruptly in the White Mountains (Oliver and Robbins, 1973; Oliver, 1982, unpub. data, 1984). Strong negative Bouguer anomalies coincide with thick accumulations of sediment—beneath Owens Valley, Long Valley, the Mono Lake basin, and the Great Valley of California.

From west to east along line A–B (fig. 45), Bouguer gravity values increase from −85 mGal at the west end to −80 mGal 15 km to the east, decrease to about −230 mGal in the Owens Valley region, then rise to about −200 mGal in the White Mountains (fig. 57). Isostatic anomalies show the same general pattern in subdued form; they increase from about −40 mGal at the west end of the profile to −20 mGal 15 km to the east and to −10 mGal 40 km farther to the east, decrease gradually to −40 mGal in Owens Valley, then increase abruptly to −10 mGal in the White Mountains.

The high values at the west end of the Bouguer gravity profile coincide with the first exposures of metamorphic and granitic rocks to the east, whereas the highest values in the western part of the isostatic profile coincide with the Mountain View Peak roof pendant farther east. The gradual decrease of the Bouguer anomaly eastward across the batholith reflects eastward thickening of the low-density batholithic and lower-crustal rocks and also their eastward-decreasing surface densities, which were not taken into account in calculating the gravity values. Whether the negative anomalies associated with deep Cenozoic sedimentation in the Owens Valley and the Great Valley of California are compensated at depth is not known.

MAGNETIC DATA

Magnetic patterns in the rocks of the batholith correlate closely with variations in the abundance of magnetite in the metamorphic and granitic rocks and are quite irregular. The original residual magnetic-intensity profile trends N. 22° E. and crosses line A–B (fig. 45) just east of the summit of Tamarack Mountain (C in fig. 57). The west end of the original profile extends south of the map area. The two features of greatest interest are the twin highs at the west end of the profile, which reflect outcrops of olivine-bearing hornblende gabbro south of the map area, and the low just east of the twin highs. The granitoids that correspond to this low are chiefly tonalite, much of which in the western Sierra Nevada is notably deficient in magnetite. The broad dip in the eastern part of the profile reflects the structural and topographic depression of Owens Valley, which affects the magnetic data because measurements were made at a constant altitude.

HEAT GENERATION AND HEAT FLOW

Heat flow in the Sierra Nevada is extremely low, but both heat flow and heat generated in the granitic rocks from the radioactive disintegration of isotopes of potassium, uranium, and thorium increase eastward to the Sierran crest (fig. 57). Heat flow measured in six boreholes across the map area show a steady increase eastward, rising from a little more than 0.4 heat-flow units (HFU) in the west to about 1.8 HFU in the White Mountains (Lachenbruch, 1968; Lachenbruch and others, 1976). An abrupt increase in heat flow occurs at the eastern escarpment of the Sierra Nevada (fig. 57), and heat-flow values are high and variable (2 to 7 HFU and locally greater) farther east across the Basin and Range province (Lachenbruch, 1968; Roy and others, 1972; Henyey and Lee, 1976).

All the measurements of heat flow within the Sierra Nevada are lower than the average for continental areas, and 0.4 HFU is among the lowest values that have been measured in any granitic rocks and is lower than those obtained in the ocean basins. A plot of heat-flow measurements (in HFU) against heat-generation measurements (in HGU) made at the drill sites yields an approximately straight line that intersects 0 HGU at 0.4 HFU. Lachenbruch (1968) and Roy and others (1972) interpreted this amount of heat as the contribution of the mantle beneath the Sierra Nevada and the excess heat flow (>0.4 HFU) as coming from radioactivity in the crust. This inferred mantle

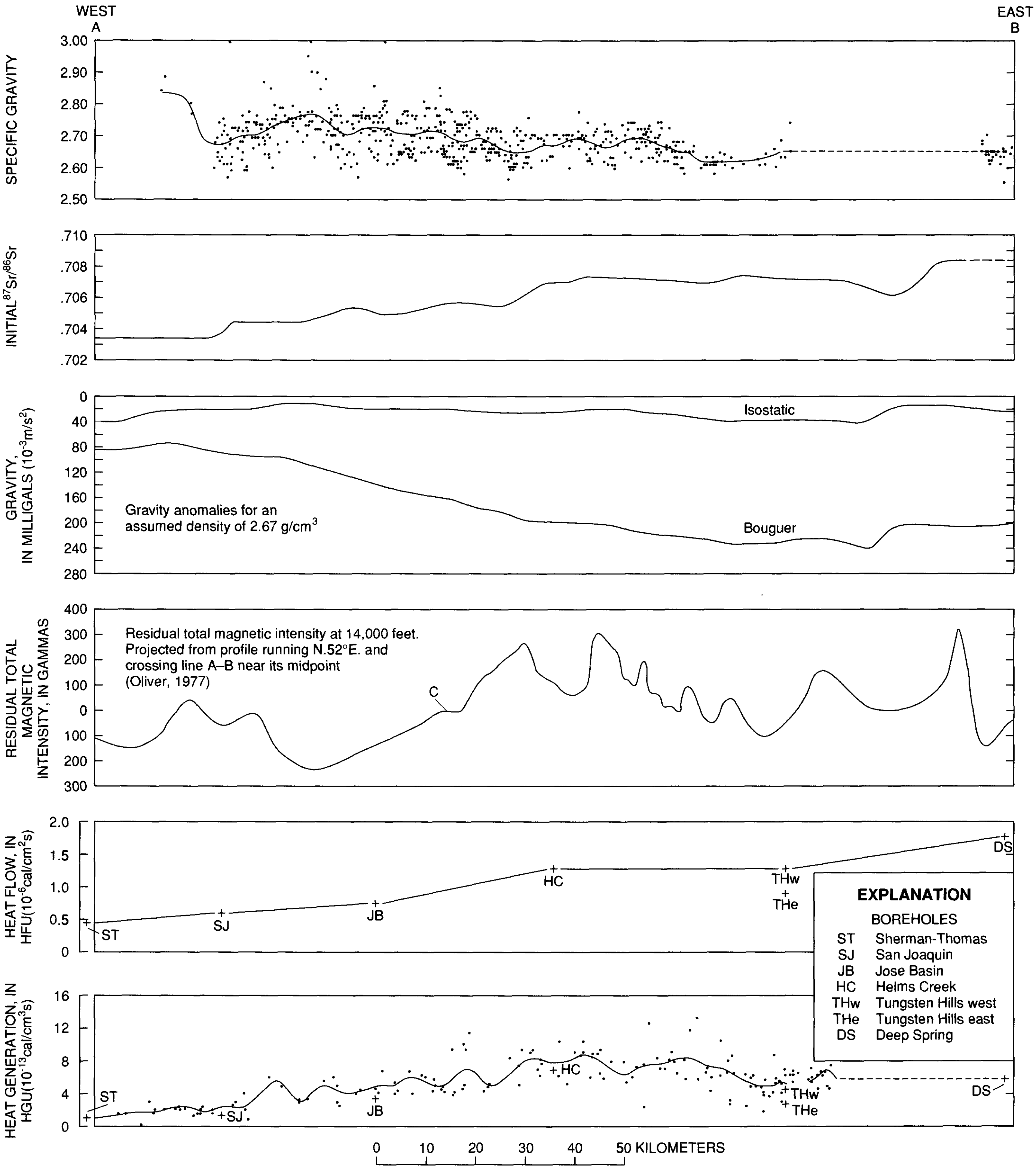


FIGURE 57.—Specific-gravity, initial $^{87}Sr/^{86}Sr$, and geophysical data across Sierra Nevada along line A–B (fig. 45). Initial $^{87}Sr/^{86}Sr$ data from Kistler and Peterman (1973). Residual total magnetic-intensity curve (at 4,267 m) is projected from profile that strikes N. 52° E. (Oliver, 1977) and crosses line A–B just west of summit of Tamarack Mountain (C on curve). Heat-flow data from Lachenbruch (1968) and Lachenbruch and others (1976); heat-generation data from Wollenberg and Smith (1968). Dashed lines indicate few data points (projection uncertain).

contribution is low compared to that in other regions and was interpreted by Henyey and Lee (1976) to indicate the presence (now or in the geologically recent past), at depth, of a cold subducted slab, shielding the batholith from mantle-derived heat. High and variable heat flow within the unprotected Basin and Range province is associated with uplift and tectonic extension. Upwelling warmer material replaces material moving laterally in the extending layer and thus preserves isostatic equilibrium.

XENOLITHS IN TERTIARY VOLCANIC ROCKS

Xenoliths brought to the surface in late Cenozoic flows and feeders at several localities (fig. 55) within and adjacent to the map area (pl. 1) provide direct evidence of the composition of the lower crust beneath the batholith. The Jackson Butte and nearby Golden Gate Hill dacite domes in the western foothills northwest of the map area and the Oak Creek and associated Waucoba basalt flow at the base of the eastern escarpment of the Sierra Nevada just south of the map area are along the west and east margins of the batholith. Three other studied xenolith sources, the Blue Knob alkali basalt plug, the Chinese Peak trachyandesite flow, and the Big Creek andesite neck, are in the interior of the batholith and within the map area (pl. 1).

The Jackson Butte and nearby Golden Gate Hill dacite domes (Rose, 1959) on the west side of the batholith and the Oak Creek and associated Waucoba basalt flows on the east side (Wilshire and others, 1972) contain magnesian peridotites (generally hornblende- or spinel-bearing lherzolites) in addition to abundant fragments of near-surface country rocks. The crust is relatively thin beneath these localities along the margins of the batholith, and the source of the peridotite presumably was the upper mantle.

Each of the localities in the core of the batholith has yielded a unique assemblage of xenoliths. The Blue Knob alkali basalt plug contains peridotite, pyroxenite, and sparse feldspathic granulite (Wilshire and others, 1988); the Big Creek andesite neck contains abundant eclogite and garnet granulite, less abundant peridotite and feldspathic granulite, and sparse peridotite (Dodge and others, 1988); and the Chinese Peak trachyandesite flow contains abundant pyroxenite and feldspathic granulite and sparse peridotite (Dodge and others, 1986).

Wilshire and Pike (1975) and Wilshire and others (1988) suggested that plagioclase-bearing peridotite xenoliths such as those at Blue Knob originated as a result of the diapiric rise of chrome diopside peridotite through the upper mantle and into the crust, where it mingled with crustal melts of varying composition and was concentrated in thin layers, producing the broad compositional range of feldspathic and nonfeldspathic peridotites, pyroxenites, and gabbroids. Wilshire and others (1988) suggested that the mingling at Blue Knob occurred at the time the enclosing alkali basalt was erupted, about 3.5 Ma, but it seems equally possible that the mingling occurred during the Mesozoic.

The andesite pipe near Big Creek, which intrudes the megacrystic facies of the ~104-Ma Dinkey Creek Granodiorite, contains peridotites, both olivine-bearing and olivine-free pyroxenites, eclogites, several varieties of quartz-free metamorphic rocks, and partially melted gabbroids and granitoids (Domenick and others, 1983; Dodge and Bateman, 1988). Inclusions having sedimentary protoliths probably represent the refractory residue of crustal material that remained in the lower crust after the parent magma of the Dinkey Creek Granodiorite was generated and rose into the upper crust. The abundant eclogite and sparse grospydite (carbonate-bearing grossularite-pyroxene rock) are especially significant because their presence supports the suggestion of Henyey and Lee (1976) that ocean floor has been subducted beneath the root of the batholith. The Big Creek eclogites are chemically similar to eclogites from the glaucophane schist terranes of the Franciscan Complex of the California Coast Ranges, and their protoliths probably were ocean-floor basalt. Presumably the grospydite was derived from limestone.

Values of $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ range widely in the different xenoliths from the Big Creek locality and approximately duplicate the spread of values for the granitic rocks across the batholith. Domenick and others (1983) postulated that the samples came from different depths and that their isotopic ratios represent the ratios at those depths at the time the batholith was emplaced. They further suggested that the granitic magmas were generated within zones at different depths and that their isotopic diversity reflects the magma compositions of these zones rather than simple mixing of different proportions of crustal and mantle material, as had been suggested by DePaolo (1980).

Olivine-free orthopyroxenites, websterites, and clinopyroxenites are the dominant rock types contained in the toe of a trachyandesite flow remnant at Chinese Peak, but orthogranulites (metagabbro) and paragranulites are abundant, and iron-rich harzburgites and lherzolites are also present. Dodge and others (1986) interpreted this assemblage to indicate the presence of a mafic-ultramafic complex at depth. U-Pb and Rb-Sr isotopic data are compatible with this complex having been emplaced in the Mesozoic, probably contemporaneous with the nearby ~104-Ma Dinkey Creek Granodiorite, although they do not precisely indicate that age. Rare, garnet-bearing ultramafic xenoliths yield pressure estimates of 13 kbar, corresponding to a depth of about 43 km, but an absence of garnet or garnet pseudomorphs in most of the ultramafic xenoliths suggests that lower equilibration pressures—5 to 9 kbar, equivalent to depths of 17 to 30 km—are more likely. Because initial $^{87}Sr/^{86}Sr$ values for the granitoids are intermediate between those for the paragranulites and those for mafic and ultramafic xenoliths, Dodge and others (1986) interpreted the granitoids to be the product of mixing of basaltic magma (represented by the mafic and ultramafic xenoliths) and crustal materials (represented by the paragranulites) followed by crystal fractionation.

COMPOSITION OF WATERS IN SPRINGS AND WELLS

Barnes and others (1981) determined the chemical composition of waters in soda springs in the Sierra Nevada including five within the map area, and Mack and Ferrell (1979) determined the composition of waters from wells and springs in the western foothills. Both groups of investigators concluded that the waters are mixtures of connate waters from metamorphic rocks and meteoric waters. The water from the soda springs is saturated with amorphous silica, has high concentrations of magnesium, calcium (120 to 620 mg/L), chlorine (as much as 2500 mg/L), boron (as much as 33 mg/L), and detectable amounts of bromine and iodine. Barnes and others (1981) considered the combination of amorphous silica and magnesium to indicate the presence of serpentinite at relatively shallow depth, calcium to indicate the presence of calcite, and chloride, bromine, and iodine to indicate the presence of marine clastic rocks. Three of the five soda springs sampled within the map area are in granitoids; the other two issue from alluvium peripheral to Mono Lake.

Mack and Ferrell (1979) determined that 31 wells and springs in the western foothills have abnormally high concentrations of sodium chloride (averaging 1,300 mg/L) and contrast markedly with the good quality bicarbonate-rich water in several thousand other wells drilled in granitic rocks. Of these 31 springs and wells, 18 are within the map area and include 15 springs and wells in the vicinity of Oakhurst, wells at the Rancheria Fire Station and nearby Arnold Spring, and a well at Shaver Lake Point. (The remaining 13 NaCl-rich wells and springs are more than 30 km south of the map area.) All the wells within the map area except one are in the Bass Lake Tonalite. Mack and Ferrell (1979) interpreted the water in the wells to be connate water from metamorphic rocks strongly diluted with meteoric water and postulated that the wells and springs lie along a concealed lineament that strikes N. 30° W. and lies just northeast of the Mountain View Peak and Coarsegold roof pendants. They suggested that the lineament marks the position of the foothills suture, which they postulated separates continental basement from oceanic basement.

AGE, ORIGIN, AND COMPOSITION OF THE SIERRAN ROOT

Two alternative interpretations of the origin of the Sierran root have received substantial support: (1) It formed in the late Cenozoic when the Sierra Nevada was tilted westward and uplifted to its present height as a result of tectonic and (or) magmatic processes; or (2) it formed during the Mesozoic when the batholith was emplaced, chiefly during the 30-m.y. timespan from 115 to 85 Ma, when the most voluminous granitoids were emplaced, and is essentially the underpinning of the batholith.

Christensen (1966) concluded that because the region is in isostatic equilibrium, the root must have originated concurrently with Cenozoic uplift. He argued that uplift could not have resulted from delayed isostatic adjustment to a Mesozoic root that had lain dormant for at least 50 m.y. while the land was being eroded. He suggested that the generation of magma at depth caused crustal thickening and uplift. Hamilton and Myers (1967) concurred with the opinion that the root is of Cenozoic age but postulated that the cause of uplift was radioactive decay in the crust, which retarded cooling and conduction of mantle heat and caused phase changes in the upper mantle and lower crust with consequent enlargement of the Sierran root.

In contrast to these interpretations that the root is of Cenozoic age, Bateman and Wahrhaftig (1966), Bateman and Eaton (1967), Schweickert and Cowan (1975), Crough and Thompson (1977), and Chase and Wallace (1986) assumed that the Sierran root originated concurrently with the batholith, during the Mesozoic. Dodge and others (1986) reported isotopic data that indicate xenoliths brought to the surface at

Chinese Peak in a 10-Ma trachyandesite flow are of the same general age as the nearby Dinkey Creek Granodiorite. They suggested that the development of mafic-ultramafic complexes in the lower crust during generation of the batholith may account for the present crustal thickness. Hay (1976) suggested that the root originated in the Mesozoic and that movements along the San Andreas fault during the late Cenozoic produced northwest-oriented tensional stresses that allowed the Sierra Nevada to rise and still maintain isostatic equilibrium. A proposal by Chase and Wallace (1986) that the root originated in the Mesozoic and that isostatic equilibrium was attained only when crustal extension in the Basin and Range province during the Tertiary permitted the Sierra Nevada to rise is questionable because it seems to imply that a significant gravity anomaly coincident with the present Sierra Nevada existed for many tens of millions of years. Crough and Thompson (1977) assumed that the root originated in the Mesozoic and that the Cenozoic uplift was caused by the conversion of lithospheric mantle to lower density asthenosphere as suggested by abnormally low *P*-wave velocities (υ_p=7.9 km/s) in the upper mantle. They compared the Sierra Nevada with the Appalachian Range, which is only about half as high and has a similar crustal thickness and similar *P*-wave velocities but has normal upper-mantle velocities (~8.2 km/s). They suggested that a slab of cold lithosphere subducted beneath the Sierra Nevada, as postulated by Henyey and Lee (1976), shielded the Sierra Nevada from high regional asthenospheric heat flux and caused the presently low heat flow.

Seismic measurements and the study of xenoliths brought to the surface in Cenozoic volcanic rocks show that the Sierra Nevada batholith and its underlying crustal rocks are layered. Granitoid and associated greenschist- to amphibolite-grade metamorphic rocks exposed in the central Sierra Nevada correspond to a surface seismic-density layer (velocity υ_p=6.0 km/s; density ρ=2.67 g/cm^3). Studies concerning the southernmost Sierra Nevada, where deeper-crustal materials are reportedly exposed (Ross, 1983, 1985), indicate that these rocks overlie a largely metaigneous assemblage of hornblende-rich gneissic amphibolite- to granulite-grade rocks. These metaigneous rocks form a roughly 10-km-thick lens (υ_p=6.4 km/s; ρ=2.83 g/cm^3) in the lower part of the upper crust, which thins towards the margins of the batholith. The lower crust (υ_p=6.9 km/s; ρ=3.03 g/cm^3) constitutes about half the entire crust. Xenoliths from Tertiary volcanic feeders indicate that in the interior of the batholith the upper part of the lower crust, to a depth of about 40 km, is made up of a series of deformed mafic-ultramafic intrusions. The mafic-ultramafic intrusions cut granulite-grade feldspathic metamorphic rocks. Less-feldspathic granulite occurs below this level in the lowest parts of the crust. Beneath the deeper part of the Sierran root, a seismic discontinuity at a depth of 55 km may mark downward transition to an eastward-dipping, downdragged slab of ocean-floor basalt that was transformed to eclogite (υ_p=7.9 km/s; ρ=3.23 g/cm^3) (Henyey and Lee, 1976). In theory, at shallower depths the eclogite grades westward into basalt, but a transition has not been recognized. Beneath the deepest part of the Sierran root, olivine-rich ultramafic rocks occur at greater mantle depths, but in the margins of the batholith they immediately underlie the Moho. They also occur as diapirs or tectonic interlayers in the lower crust.

The incidence if basaltic volcanism during the Cenozoic shows that mantle-derived magma was introduced into the crust during that era, but very much larger volumes must have been introduced during the Mesozoic to provide heat for the widespread generation of granitic magmas. The likelihood of a Mesozoic mafic-ultramafic complex at depth beneath the Chinese Peak trachyandesite flow (Dodge and others, 1986) and a spread of isotopic ratios determined for different xenoliths from the trachyandesite at Big Creek equal to that across the batholith (Domenick and others, 1983) support this view. The lower crust probably is largely composed of both mantle-derived intrusions and of refractory material left behind during the rise of the granitic magmas.

WESTWARD TILTING AND UPLIFT OF THE SIERRA NEVADA

Ongoing uplift and westward tilting of the Sierra Nevada are related to the late Cenozoic extension and deformation affecting the Basin and Range province and began many tens of millions of years after the granitoids were emplaced and after the landscape had been reduced to one of low relief during earlier uplift and erosion that accompanied emplacement of the batholith. Granitic cobbles in conglomerate at the base of the Minarets sequence show that granitic plutons were exposed as early as about 100 Ma. By Eocene time, erosion had produced the low topographic relief visible today in interfluvial upland surfaces (the so-called Eocene peneplain). During the late Eocene, about 40 Ma, the San Joaquin River drained a large area to the east and flowed across the Sierra Nevada (Huber, 1981). The uplift and tilting that account for the present configuration and height of the range began about 25 Ma, but two-thirds of the uplift has taken place during the last 10 m.y. (Huber, 1981). The initial rate probably did not exceed 0.03 mm/yr at the present drainage divide, and the rate has increased dramatically to an estimated 0.3 mm/yr at present.

In isostatically compensated terranes, crustal thickness generally correlates poorly with terrane elevation and in many places requires that the mantle was involved. For example, the Moho apparently is not depressed relative to regional depths beneath either the Cascade Range or the Rocky Mountains, and crustal thicknesses in the structurally high Basin and Range province are in the general range of 25 to 30 km, compared to thicknesses of 40 to 45 km under the Colorado Plateau and even greater thicknesses under extensive areas of the relatively low Eastern United States. In general, terrane elevations appear to correlate better with variations in *P*-wave velocities (and deduced densities) in the upper mantle than with crustal thicknesses. Thus the *P*-wave velocities of the upper mantle average 7.9 km/s under the Sierra Nevada, about 7.8 km/s under the structurally high Basin and Range province, about 8.0 to 8.1 km/s under the structurally lower Colorado Plateau, and more than 8.1 km/s in the relatively low Eastern United States.

These considerations favor the interpretations of Crough and Thompson (1977) and Mavko and Thompson (1983) that the late Cenozoic uplift and tilting is related to high heat flow and conversion of upper-mantle lithosphere to lower density asthenosphere. The subducted slab of eclogite postulated by Henyey and Lee (1976) to underlie the root shields the Sierra Nevada from the high asthenospheric heat flow that is affecting the Basin and Range province but allows the upper-mantle lithosphere beneath the Sierra Nevada to be converted to lower density asthenosphere, causing uplift in the Sierra Nevada.

ORE DEPOSITS

The discovery of gold in the western foothills of the Sierra Nevada in 1848, which led to the California gold rush of 1849, has led many people to think of the Sierra Nevada as a vast storehouse of valuable mineral commodities. It is true that the western metamorphic belt has yielded significant amounts of gold, copper, chrome, and limestone, and that important deposits of tungsten, molybdenum, copper, and gold have been mined from skarn deposits in the eastern Sierra Nevada. However, virtually all these deposits are located within the country rocks. In ore deposition, the batholith served primarily as a "heat engine," mobilizing fluids that carried and concentrated metals which had been dispersed in the country rocks. The formation of tungsten skarn deposits is an exception, however; they were formed by the reaction of tungsten-bearing magmatic fluids with carbonate in the country rocks.

GOLD

Most of the production of gold in California has been from placer deposits, and probably all the streams that cross the western metamorphic belt within the map area have been worked in placer operations, especially the Merced, Fresno, and Chowchilla Rivers and the Coarsegold, Fine Gold, and Mariposa Creeks. Placer-mining activity in this area was intense within a few months of the discovery of gold at Sutter's mill in 1848 but virtually ceased by 1870. Since then a few small operations have been undertaken, apparently without notable success, although panning for gold is still a popular recreational activity.

Gold has also been extracted from lode deposits, chiefly along the Mother Lode—a 1- to 6-km-wide system of en echelon quartz veins and mineralized rock that follows the Melones fault zone. The Mother Lode extends only into the southwest corner of the map area, near Mariposa, but lode deposits farther east in the western metamorphic belt have also yielded significant amounts of gold. Past production has been mainly from quartz veins, but current interest is in mineralized country rock. The veins dip steeply northeast and pinch and swell abruptly; rarely can they be followed for more than 1 km. The veins consist of milky quartz that contains sulfide minerals (chiefly pyrite) and free gold, and many are ribboned with layers of slate and schist (Dodge and Lloyd, 1984). Many of the finest crystalline gold specimens have been recovered from pockets in the veins, and the Colorado Mine, located about 8 km north of Mariposa, reportedly was still yielding excellent specimens in 1982 (Kampf and Keller, 1982).

The gold in country rock is finely disseminated in altered greenstone and schist. Mineralized greenstone is known as "gray ore" and consists of ferromagnesian carbonates and lesser amounts of sericite, albite, quartz, pyrite, and arsenopyrite; it is interlaced with veinlets of quartz, carbonate, and albite. Mineralized schist consists chiefly of ferromagnesian carbonate and subordinate sericite, quartz, albite, and pyrite.

The age of the Mother Lode veins has long been debated. Traditionally, the veins have been thought to be Late Jurassic "Nevadan." Most of the veins postdate significant movement along the Melones fault zone and parallel the regional schistosity, but in a few places there are older veins that were deformed during the development of schistosity (Clark, 1964; F.C.W. Dodge, oral commun., 1985). Isotopic dating of three samples of mariposite-bearing rock northwest of Mariposa and west of the map area by the K-Ar method indicates ages that range from 127 to 108 Ma (Kistler and others, 1983). One sample yielded a K-Ar age of 116±3 Ma and a Rb-Sr age of 114.6±3 Ma, similar to K-Ar ages on mariposite from the Alleghany district (Böhlke and McKee, 1984) and to K-Ar ages for samples from other Mother Lode veins (Böhlke and Kistler, 1986).

These ages coincide with the approximate age of the Fine Gold Intrusive Suite, which lies along the west side of the batholith within the map area and extends to the south and possibly also to the north. The correspondence of isotopic ages and an almost complete absence of quartz veins within the intrusive rocks suggest that the role of the Fine Gold Intrusive Suite was to supply the heat needed to mobilize fluids, which concentrated gold and sulfides disseminated in the country rocks in fracture zones and altered the adjacent rocks. Older ages for veins may be correlative with the emplacement of Late Jurassic and Early Cretaceous plutons, which, however, are of much more limited volume than the granitoids of the Fine Gold Intrusive Suite.

Within the map area the only large mine that lies within the Mother Lode proper is the Mariposa Mine, but other mines with significant past production lie farther east. The Mariposa Mine, on the southern outskirts of Mariposa, is reputed to have been discovered by Kit Carson and is one of several on the Las Mariposas land grant acquired by General John C. Fremont in 1847. Lode mining at the Mariposa Mine was started in 1849, and the first stamp mill in California was set up there. The value of the gold extracted from this mine, estimated at $2,395,000 (equivalent to about 116,000 troy oz of gold at $20.67 per troy oz, the price of gold during most of the period from 1835 to 1934), is nonetheless much less than that of mines farther north. Most of the Mariposa gold was mined before 1915, but the mine was reopened briefly in the 1930's and again in 1955. The mine workings consist of an inclined shaft that extends to a vertical depth of 380 m and more than 2,700 m of drifts on eight levels. The most prominent quartz vein dips southwest, opposite to the dip of most Mother Lode veins, and contains no gold; the gold was in a branching vein that dipped northeast. Most of the gold was free gold, but some was contained in sulfides (chiefly pyrite and arsenopyrite) that also yielded small amounts of silver and copper.

The area east of the Mother Lode proper in the southern part of the western metamorphic belt is riddled with old workings, and deposits at Clearing House, on the north side of the Merced River, and at Hite Cove on the South Fork of the Merced River have greater recorded production than the Mariposa Mine. According to Bowen and Gray (1957), who described many of the lode deposits in this area, ore mined at the Clearing House Mine yielded about $3,350,000 and ore mined at the Hite Cove Mine yielded $2,750,000 to $3,000,000; at $20.67 an troy oz, these amounts are equivalent to 162,000 and 133,000 to 145,000 troy oz of gold, respectively. Both of these deposits are in quartz veins within the western margin of a zone of carbonaceous metapelite, which may coincide with the Calaveras-Shoo Fly thrust fault.

South of the western metamorphic belt, gold has been mined in the Coarsegold and Mountain View Peak roof pendants and in Hildreth Mountain. Several small deposits at Grub Gulch, in the western part of the Coarsegold roof pendant, are said to have been worked as early as 1849. Exploration there reportedly reached a depth of 250 m and production totaled about $960,000, equivalent to about 46,400 troy oz of gold (Clark, 1970). Farther south in the Mountain View Peak roof pendant and in Hildreth Mountain, gold-bearing quartz veins range from a few centimeters to 1 m thick and dip at angles of less than 45°. Several veins have been explored to depths of 150 to 180 m. Only a few production figures for these mines are available. Clark (1970) reported that six mines in Hildreth Mountain yielded about $450,000, equivalent to about 21,800 troy oz of gold.

In the eastern Sierra Nevada lode gold deposits have been mined in the Bishop Creek septum, the Mount Morrison and Ritter Range roof pendants, and in the White Mountains. Gold has also been extracted from ore mined for tungsten at the Pine Creek Mine and other mines in the Pine Creek septum. The Cardinal Mine, in a spur of the Bishop Creek septum that extends to the north across the North Fork of Bishop Creek, reportedly yielded $1,570,000, equivalent to about 76,000 troy oz of gold, from 1910 to 1938 (Tucker and Sampson, 1938). The ore body was 1 to 3 m thick and was localized in a shear zone in quartzite. According to Knopf (1918), the ore consisted of quartzite that contained disseminated sulfides, predominantly pyrrhotite but including arsenopyrite, sphalerite, chalcopyrite, pyrite, and molybdenite.

Several mines in the north end of the western volcanogenic lobe of the Ritter Range roof pendant, near Mammoth Lakes, together produced less than $1,000,000 in gold and silver. Tucker (1927) reported that the ore in the Mammoth Mine, the most productive in the district, was "14 to 40 feet" (4–12 m) thick and consisted of quartz and decomposed metalatite that contained magnetite, auriferous pyrite, chalcopyrite, sphalerite, and free gold.

Deposits east and south of Bishop have produced relatively small amounts of gold. The Poleta Mine, east of Bishop in the lower slopes of the White Mountains, has produced more than 2,000 troy oz of gold and 800 troy oz of silver since 1900, and the total production may be twice this amount. The deposit is in a quartz-sulfide vein that lies within a bedding-plane gouge zone in the Early Cambrian Harkless Formation. In the deeper part of the mine the vein consists chiefly of quartz, calcite, auriferous pyrite, and a little chalcopyrite, but in the upper levels the sulfides are altered to limonite that carries free gold. Sulfide-bearing quartz veins at Fish Spring Hill south of Big Pine are in the Jurassic Tinemaha Granodiorite. Most of the veins strike somewhat north of west and dip 20° to 50° N. The many veins in the area have been exploited in the Cleveland and Commetti Mines. The total recorded production from 1893 to 1949 from these mines was about 42,400 troy oz of gold, about 34,200 troy oz of silver, and about 900 kg of copper (Bateman, 1965b).

COPPER

The foothills copper belt, which lies west of the Mother Lode in the western foothills of the Sierra Nevada, extends southeastward into the map area, and the Pine Creek tungsten mine in the eastern Sierra Nevada has also yielded significant amounts of copper as a byproduct. Five small deposits that are part of the foothills copper belt lie within the southwest corner of the map area, in strata west of the Melones fault zone. All the mines exploit sulfide-bearing quartz veins that occupy zones of intense shearing. The following remarks are based on reports by Laisure (1928), Cox and Wyant (1948), Bowen and Gray (1957), and other reports contained in several Annual Reports of the California State Mineralogist and its successor, California Journal of Mines and Geology.

The principal sulfides in the Jesse Belle, Green Mountain, and Lone Tree Mines are chalcopyrite, pyrite, and pyrrhotite; pyrrhotite is the most abundant sulfide in the Buchanan and Daulton Mines. At all five mines, the gangue minerals are quartz, biotite, muscovite, feldspar, graphite, and epidote; the country rock is andalusite schist at the Dalton and Jesse Belle Mines, quartz-biotite schist at the Buchanan Mine, and greenstone at the Green Mountain and Lone Tree Mines. Much of the ore recovered at the Green Mountain and Lone Tree Mines, and probably at the other mines as well, was enriched near-surface material. The Buchanan Mine was opened during the Civil War, in 1863, and operated continuously until 1909; the other mines probably were opened about the same time. The Daulton Mine was last worked from 1942 to 1945 (during World War II), and the bulk of the production at the Jesse Belle Mine was also during that period. Incomplete records suggest that the total production from the Daulton, Jesse Belle, and Buchanan Mines was 1.4 to 1.8 kg of copper and some gold. The Green Mountain Mine reportedly had 1220 m of underground workings and produced large quantities of high-grade oxide and carbonate ore, but no production records are available.

TUNGSTEN SKARN DEPOSITS

Tungsten is by far the most important metal that has been mined within the map area. Mines in the eastern Sierra Nevada near Bishop have yielded a large part of the total tungsten production in the United States. Much of this production has been from the Pine Creek Mine, and substantial additional amounts have come from nearby mines. With a few minor exceptions, the tungsten deposits of the Sierra Nevada are in skarns and the tungsten is contained in the mineral scheelite ($CaWO_4$). Skarns (tactite) are common along contacts between granitoids and carbonates. Two factors may account for the localization of minable tungsten deposits in the eastern Sierra Nevada: (1) The eastern Sierra Nevada marks the west limit of abundant Paleozoic limestones, and (2) the lower-crustal component of the source materials for the granitoids and for the skarns may have been enriched in tungsten. The second factor is suggested by the progressive change in the compositions of the granitoids across the Sierra Nevada and by a general association of skarn-type tungsten deposits in western North America with granite and granodiorite rather than with quartz diorite, tonalite, and cogenetic felsic rocks. However, analyses indicate that few Sierran granitic rocks contain more than 0.5 parts per million tungsten and fail to reveal any evidence of an eastward increase in the amount. Two exceptions to the low tungsten content of the granitoids are the granite of Dinkey Dome and the Evolution Basin Alaskite, both of which contain more than 400 parts per million tungsten. The tungsten in these highly evolved rocks probably was carried in a vapor phase.

GENERAL FEATURES OF THE SKARNS

The mineral assemblages in most of the tungsten-bearing skarn deposits were formed during two distinct but probably continuous or overlapping stages of metasomatism, which followed nonadditive thermal metamorphism that resulted from the intrusion of the granitoids. Thermal metamorphism took place with rising temperatures, and most of the metasomatism occurred with falling temperatures. During thermal metamorphism, limestone was metamorphosed to marble, and impure limestone and noncalcareous country rocks were converted to calc-silicate hornfels and noncalcareous hornfels, respectively. Although this metamorphism was nonadditive, constituents originating within the country rocks moved across bedding planes and along fractures. The garnet and pyroxene that formed in marble and calc-silicate hornfels during this stage are generally light colored, indicating a low iron content.

The skarns were formed by the reaction of ore-forming fluids emanating from the cooling and crystallizing magma with marble and calcite-bearing hornfels. Iron-rich reddish-brown garnet, gray-green pyroxene, and tiny disseminated grains of scheelite ($CaWO_4$) are the characteristic minerals of an early, anhydrous stage of metasomatism that closely followed the nonmetasomatic thermal metamorphism and thus occurred at relatively high but probably decreasing temperatures. The localization of skarn along irregularities in contacts between granite and marble, which obviously guided the movement and entrapment of ore fluids, indicates that most skarns were formed after the granitic rocks had begun to solidify. The anhydrous skarns were formed at relatively high temperatures estimated to be 550 to 650 °C (Morgan, 1975; Newberry and Einaudi, 1981) and at pressures in the range of 1.5 to 2.0 kbar, equivalent to depths of 5 to 7 km (Newberry, 1982).

As temperatures decreased, the anhydrous garnet-pyroxene skarns were replaced, partially or completely, by calcite and hydrous minerals such as epidote and amphiboles. The anhydrous skarns and the adjacent granitic rocks were locally silicified. During the silicification, quartz veins formed; scheelite was remobilized and redeposited in sporadically distributed, small but rich, coarse-grained masses associated either with quartz and epidote or with feldspars, quartz, calcite, and amphibole; and sulfides were deposited—chiefly molybdenite, pyrite, pyrrhotite, chalcopyrite, and bornite, but including sphalerite, galena, and bismuthinite. Significant amounts of byproduct copper and molybdenum have been recovered at the Pine Creek Mine.

The richest skarns were formed in relatively pure marble, which reacted readily with the mineralizing fluids. Iron-poor, light-colored calc-silicate minerals (such as wollastonite, plagioclase, grossularite, idocrase, and diopside) that were formed earlier, during the nonadditive thermal metamorphism, reacted with the ore-forming fluids much less readily than did calcite and formed skarn that generally contains little or no tungsten. Rocks in which pure calcite marble is interlayered with pelitic and arenaceous material form layered skarns. Darker colored, scheelite-bearing layers, formed from clean carbonate beds, alternate with barren or tungsten-poor layers composed of light-colored silicate assemblages.

STRUCTURAL AND STRATIGRAPHIC SETTING

Scheelite-bearing skarn ore bodies range widely in size, from a few centimeters to 10 m or more in thickness and from a few meters to as much as 100 m in outcrop length. Commonly skarn bodies are sporadically distributed along contacts between marble and granite, and, between skarn ore bodies, only a thin layer of wollastonite may occur along the marble-granite contact. Many ore bodies are localized at irregularities in otherwise concordant contacts between granite and marble, where stresses may have caused fracturing and increased permeability. Irregularities are most common in places where the carbonate beds diverge from the contact, either in strike or in dip, and, conversely, few occur along contacts where bedding is concordant with the contact. Irregularities that have vertical or steeply plunging axes appear to have guided the generally upward flow of ore-forming fluids, whereas those that have nearly horizontal axes appear to have inhibited flow and to have caused stagnation, reaction with calcite, and the formation of skarn.

The Pine Creek Mine lies along a contact between marble of probable late Paleozoic age in the west side of the Pine Creek septum and the granite that composes the Morgan Creek pluton of the Cretaceous Lake Edison Granodiorite. At the surface the contact extends for more than 5 km, but thin layers of Jurassic diorite and quartz diorite separate the granite and marble along much of the contact. The ore bodies are confined to a span of about 1 km at the north end of the contact, where diorite and quartz diorite are scarce. Several ore bodies are present along this span; some have strike lengths of more than 100 m and thicknesses of as much as 15 m (fig. 58). Between the ore bodies, the skarn is thin and contains little scheelite. The ore bodies crop out at an average altitude of about 3,500 m and have been exploited to depths of almost 1,000 m.

MINERALOGICAL AND CHEMICAL RELATIONS

In skarn formation from limestone, Si, Al, Fe, Mn, Mg, and W are added, and Ca and CO_2 are subtracted. Much is yet to be learned about the formation of skarn, but several interesting studies of Sierran skarns have been made. Significant studies of the mineralogic and chemical relations in tungsten skarns of the eastern Sierra Nevada are summarized in the following sections.

PINE CREEK MINE

In their study of the Donut ore body in the northern part of the Pine Creek Mine (fig. 58; Gray and others, 1968), the geologists at the mine were the first to show that a skarn ore body is compositionally zoned (fig. 59). This ore body is in a mineralized skarn that occupies a 1- to 7-m-wide zone in the margins of a blocky inclusion of marble, which in plan view averages about 25 m across. Most of the skarn is composed chiefly of garnet and pyroxene, but irregularly shaped masses of hornblende skarn penetrate inward from the outer margins into the garnet-pyroxene skarn. Along the margins, epidote skarn is present locally, and the granite adjacent to the epidote skarn has been silicified. This compositional pattern indicates an increase in iron and a decrease in calcium inward in the inclusion. Because of the small size of the inclusion and its complete enclosure within granite, Gray and others (1968) considered a significant contemporaneous temperature gradient to be an unlikely explanation for the compositional zoning; apparently they did not consider the effect of decreasing temperature over time in the formation of hornblende and epidote.

Newberry (1982) made the most recent and most comprehensive mineralogical and geochemical study of the skarns in the Pine Creek Mine. Most of his observations and conclusions, based largely on studies of deeper parts of the mine, agree with observations made 30 years earlier in the upper part of the mine (Bateman, 1956, 1965).

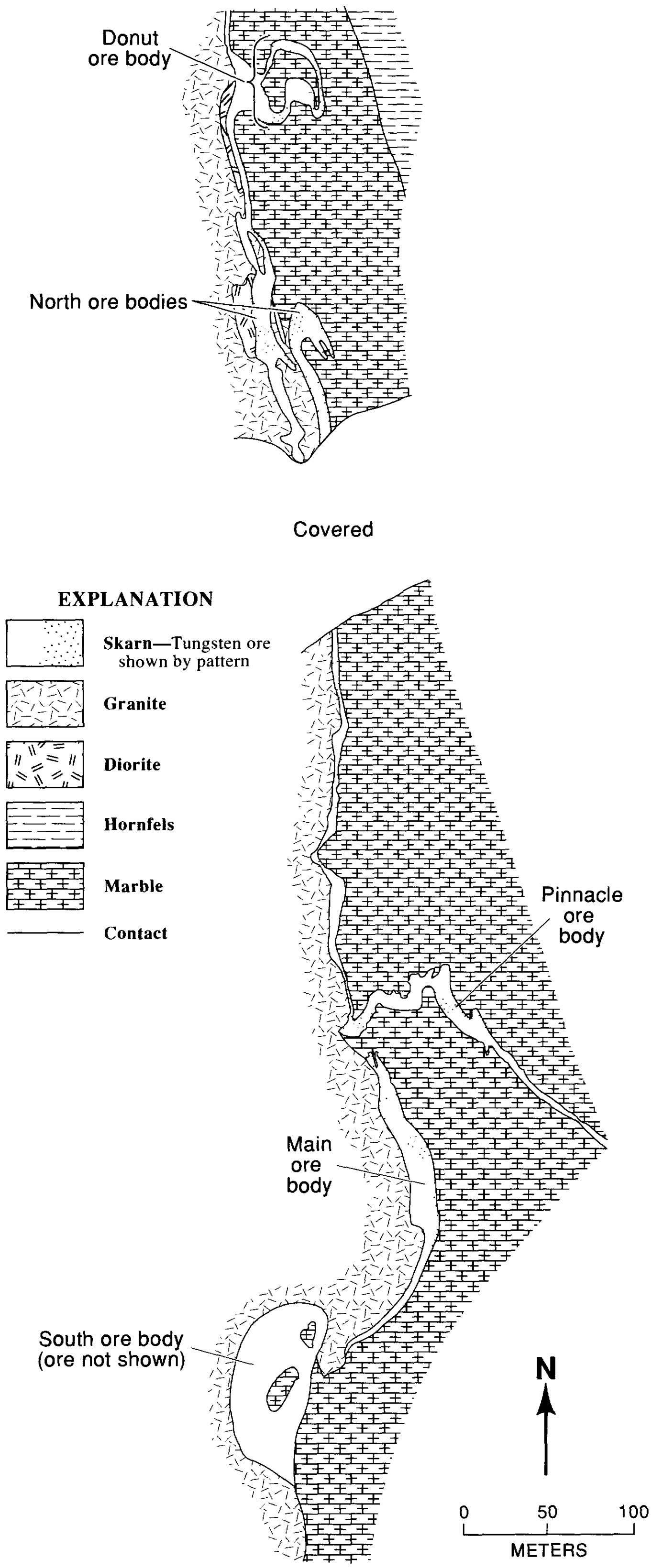


FIGURE 58.—Generalized geology of mineralized contact zone at Pine Creek tungsten mine, 17 km west of Bishop on north side of Pine Creek canyon. Modified from Lemmon (1941).

Newberry (1982) recognized the nonadditive metamorphism of the country rocks, followed by an early high-temperature, anhydrous stage of metasomatism, and a later, lower temperature stage of hydrosilicate alteration. He divided the skarns formed during the early, anhydrous stage of metasomatism into three thin zones: an inner, 1-cm- to 2-m-thick zone adjacent to granite and composed of garnet+quartz ±pyroxene; a main zone, generally 1 cm to 10 m thick, composed chiefly of garnet+pyroxene+scheelite; and an outer, 1- to 50-cm-thick zone adjacent to marble and composed of idocrase+wollastonite+

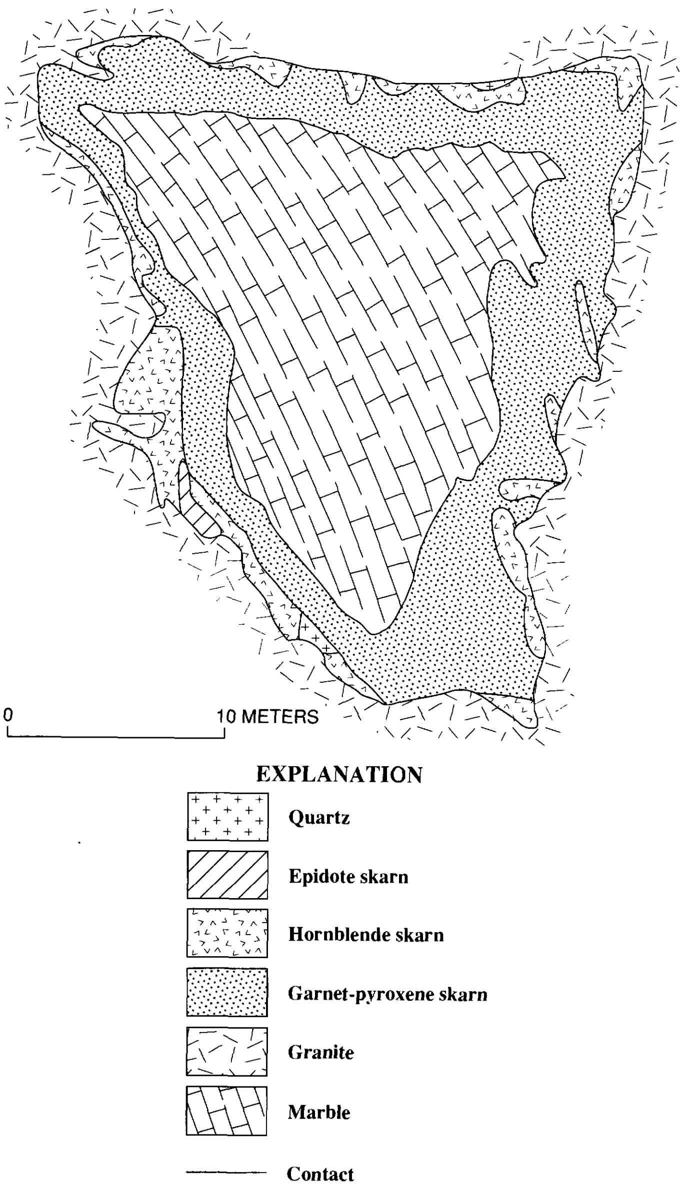


FIGURE 59.—Subsurface plan of Donut ore body, Pine Creek tungsten mine. See figure 58 for location. From Gray and others (1968).

scheelite±pyroxene. The boundary between the inner zone and the main zone is gradational and is marked by the penetration of barren veins of iron-rich garnet from the inner zone into the main zone, whereas the boundary between the main zone and the outer zone is abrupt and marked by the replacement of garnet by idocrase.

The garnet and pyroxene in all these skarn zones are much richer in iron than garnet and pyroxene formed earlier by the nonadditive thermal metamorphism of calcareous country rock. However, the mineral compositions are different in the different zones, both garnet and pyroxene becoming richer in iron toward the granite contact. Garnet in the main zone contains 25 to 55 percent andradite ($Ca_3Fe_2Si_3O_{12}$), 35 to 70 percent grossularite ($Ca_3Al_2Si_3O_{12}$), and 3 to 10 percent combined almandine ($Fe_3Al_2Si_3O_{12}$) and spessartine ($Mn_3Al_2Si_3O_{12}$), whereas garnet in the inner zone contains 20 to 80 percent almandine +spessartine. Pyroxene in the main zone contains 15 to 40 percent diopside ($CaMgSi_2O_6$), 50 to 75 percent hedenbergite ($CaFeSi_2O_6$), and about 10 percent johannsenite ($CaMnSi_2O_6$), and pyroxene in the inner zone contains about 30 to 35 percent diopside, 50 percent hedenbergite, and 15 to 20 percent johannsenite.

During the early, anhydrous stage of metasomatism, both calc-silicate and pelitic hornfelses within and adjacent to the skarn were altered to pyroxene-dominant skarn that contains little scheelite. Thin, discontinuous pyroxene-rich layers within the skarn zones probably reflect impure layers within the marble. Between the inner skarn and the granite is a 1-cm- to 1-m-thick discontinuous zone composed of massive quartz with traces of subcalcic garnet and alkali feldspar on the skarn side and hedenbergitic pyroxene and plagioclase with minor amounts of epidote on the granite side. Locally, especially in the upper parts of the mine, zones of massive quartz with minor amounts of garnet, plagioclase, and hedenbergitic pyroxene contain substantial amounts of molybdenite and lesser amounts of chalcopyrite. Gradational contacts of the massive quartz with the skarn and with the granite show that the massive quartz replaced both. Across distances of as much as 30 m, the granite adjacent to the quartz zone has been subjected to calcic alteration. The altered rock consists chiefly of plagioclase and pyroxene close to the skarn and, with distance from the contact, grades to hornblende, plagioclase, potassium feldspar, and quartz, then to unaltered granite. Undoubtedly the source of the calcium was the marble that was replaced by silicate minerals.

Subsequent to the formation of anhydrous skarn and accompanying relatively high-temperature metasomatic effects, the Pine Creek deposits underwent extensive retrograde hydrosilicate alteration at lower temperatures. Iron, potassium, and sulfur were introduced, calcium was leached, and metals were mobilized and redeposited. Masses of coarse-grained scheelite, medium-grained chalcopyrite, and sparse pyrite and apatite are present in the altered rocks and constitute high-grade copper ore and low- to high-grade tungsten ore. The mineral assemblages resulting from this alteration vary with the composition of the host rock, but in general they are characterized by biotite, quartz, amphibole, and sulfides, as well as variable amounts of magnetite, plagioclase, calcite, chlorite, epidote, and fluorite.

Retrograde alteration of the inner-zone skarn is characterized by extreme iron and potassium enrichment and calcium depletion. The alteration assemblage consists of subequal amounts of iron-rich biotite and magnetite, with minor amounts of plagioclase (An_{20-35}) and quartz. Coarse-grained scheelite is common, and some partly altered subcalcic garnet has survived.

Where the alteration was intense, an assemblage consisting of hastingsitic amphibole with lesser quartz, chalcopyrite, scheelite, and minor amounts of calcite, apatite, and titanite was produced near the contact between the inner and main skarn zones. Epidote also was formed locally in small amounts. In the main skarn zone, amphibole-rich skarn is abundant and forms veins that feather out toward the marble contact. Typical rock consists of knobs of partly altered garnet in a matrix of amphibole+quartz+scheelite+sulfides±calcite. In the outer-zone skarn, wollastonite is replaced by fine-grained quartz and calcite, and marble adjacent to the inner skarn is replaced by masses of fine- to medium-grained epidote, chlorite, and fluorite.

In addition to retrograde alteration of the anhydrous skarn zones and the redistribution of metals, the granite adjacent to zones of intense skarn alteration was altered to biotite+quartz+chalcopyrite. Quartz veins with biotite+chalcopyrite±pyrite envelopes are common, as are zones of coarse biotite+quartz with accessory chalcopyrite. Where the alteration of skarn was less intense, the adjacent granite was overprinted by an epidote-dominant retrograde assemblage consisting of an epidote+calcite replacement of plagioclase and garnet and an amphibole+quartz±calcite replacement of pyroxene. Apatite, pyrite, sphene, and locally scheelite are present in small amounts.

The lowest temperature alteration assemblage consists of zeolite zones composed of laumontite, quartz, calcite, chlorite, kaolinite, and native copper. The resulting alteration zones cut across all older rock and form pipelike masses, one of which extends upward from an altitude of 3,300 m to the surface at 4,000 m.

Metals were distributed somewhat differently during the later hydrosilicate alteration than during the earlier higher temperature anhydrous metasomatism. During the early metasomatism, scheelite was concentrated most abundantly in the outer skarn zone, adjacent to marble, where it occurs as small (0.2–0.4-mm) subhedral grains interstitial to wollastonite and idocrase. The probable explanation for this distribution of scheelite is that the ore solutions were undersaturated with respect to $CaWO_4$ and deposited scheelite where the activity of calcium ions in solution was raised by interaction with calcium-rich rocks. This resulted in continuous dissolution of scheelite close to the metasomatic source at the granite contact, where the activity of calcium ions was low, and precipitation near the marble front, where the activity of calcium ions was high.

The mechanism of sulfide deposition during the stage of anhydrous skarn formation is unclear. The sulfides are mainly in pyroxene-dominant skarn and may have formed by reduction of the metasomatic fluids upon coming into contact with rocks containing significant amounts of ferrous iron or simply as a result of the impounding and cooling of the metasomatic fluids.

The metals deposited during the high-temperature anhydrous stage were remobilized and redeposited together with newly introduced metals during the retrograde hydrosilicate alteration. High copper grades and low to high tungsten grades typically are associated with the hydrosilicate alteration. Embayment of scheelite by biotite and amphibole show that scheelite was dissolved during retrograde alteration. Coarse-grained scheelite, medium-grained chalcopyrite, and sparse pyrite were deposited in zones of biotite and amphibole alteration. These scheelite grains are low in molybdenum relative to the scheelite in anhydrous skarn. Although tungsten generally is enriched, the retrograde alteration caused tungsten depletion in some places, such as biotite-dominant zones. Although the late-stage precipitation of scheelite could be related to lower solubility of $CaWO_4$ at lower temperatures, it is more likely that it was caused by release of calcium ions into the ore solutions during the replacement of calc-silicate minerals by calcium-poor hydrosilicate assemblages.

STRAWBERRY MINE

Nokleberg (1981a) described in detail vertical compositional changes in a layer of skarn that partially replaced a steeply dipping bed of marble interstratified with plagioclase hornfels at ore body No. 7 of the Strawberry tungsten mine, which is northwest of Bishop on the west slope of the Sierra Nevada (fig. 60). From the lower contact with granodiorite upward toward the marble, four different kinds of skarns are present: hornblende skarn, pyroxene skarn, garnet skarn, and wollastonite skarn. Contacts between the different skarns are sharp, and veining of one skarn by another shows that each lower skarn was formed by replacing part of the skarn above. The composition of each skarn is fairly constant and shows no systematic internal compositional variations. Modal and chemical compositions of the different skarns in relation to one another and compared to marble are shown in figure 61. At each downward step—from marble to hornblende skarn—the most calcium-rich mineral disappears: first calcite, followed successively by wollastonite, garnet, and pyroxene. Nevertheless, until they disappear, wollastonite, garnet, and pyroxene are increasingly abundant and garnet and pyroxene are more iron rich in each replacing skarn.

Chemical compositions also differ by successive stages (fig. 61). CO_2 virtually disappears in the wollastonite skarn, and CaO diminishes in each successively underlying skarn. Al_2O_3 and SiO_2 vary by only small amounts in all the skarns except in the hornblende skarn, where SiO_2 increases. MgO increases regularly downward. Both FeO and Fe_2O_3 increase to a maximum in the pyroxene skarn and then decrease downward, but Fe_2O_3/FeO decreases downward. MnO is at its maximum in the garnet skarn and decreases downward, indicating a reducing trend. WO_3 varies inversely with Fe_2O_3 in the garnet, pyroxene, and hornblende skarns.

Nokleberg (1981a) attributed the compositional pattern of these skarns to infiltration diffusion. Ore fluids moved upward and reacted with marble to form wollastonite skarn. The deposition of Si, Al, Mg, Mn, and Ti and removal of large amounts of CO_2 and CaO caused significant changes in the compositions of the ore fluids. After local equilibrium had been attained and maintained for a period of time, another wave of mineralizing fluids worked its way upward, and then these fluids reached equilibrium with garnet skarn. The pyroxene and hornblende skarns were formed in a similar manner by later waves of mineralizing fluids.

As Gray and others (1968) had done for the Donut ore body at the Pine Creek Mine, Nokleberg (1981a) considered and dismissed the possibility of a temperature gradient that decreased outward from the granodiorite contact, and he also seems not to have considered the effect of decreasing temperature over time in his analysis of successive skarn formation. Another possibility that seems not to have been considered is that early introduction of meteoric water and later depletion of the

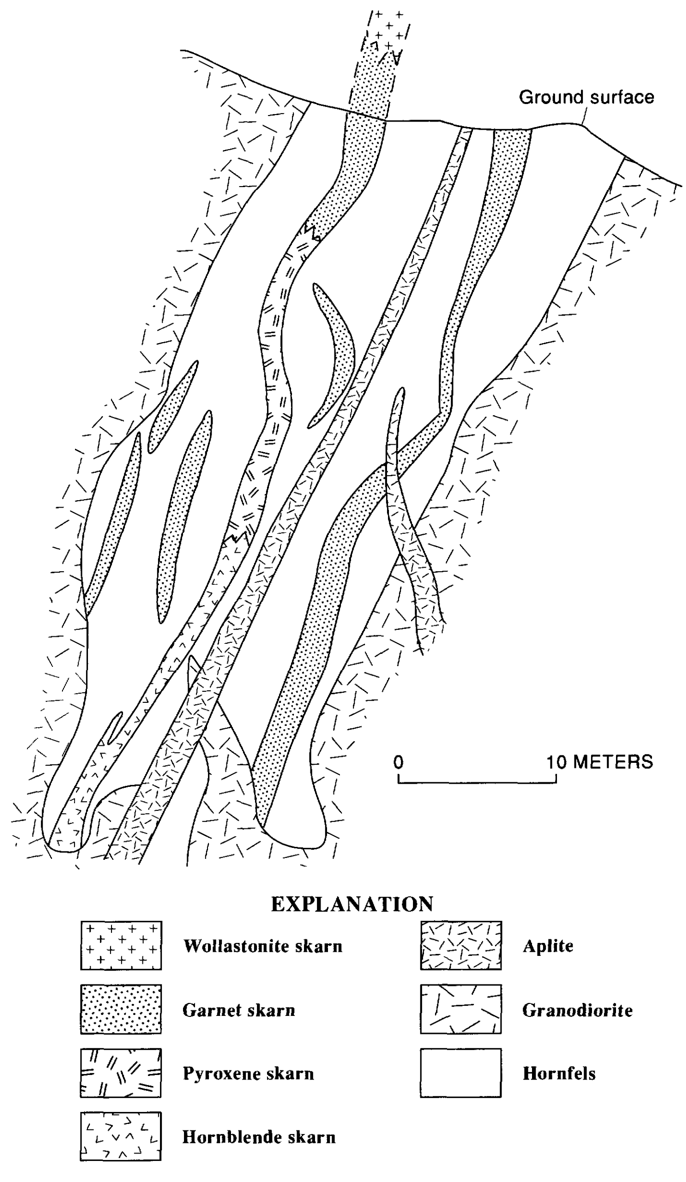


FIGURE 60.—Section across ore body No. 7 of Strawberry tungsten mine, about 90 km northeast of Fresno, Calif., near Clover Meadow in eastern Madera County. Part drawn above ground surface represents projection (along strike) of mined-out but historically well documented skarn. Modified from Nokleberg (1981b).

supply could account for the pattern of decreasing Fe_2O_3/FeO in the garnet, pyroxene, and hornblende skarns.

SKARNS IN THE MOUNT MORRISON ROOF PENDANT

Morgan (1975) studied four skarns in the Mount Morrison roof pendant, which is north of Bishop and adjacent to a pluton of the Round Valley Peak Granodiorite. He concluded that their mineralogic and chemical characteristics are the result of the mixing of different proportions of late-stage magmatic fluids with fluids produced by the decarbonation of carbonate rocks and that the mixing ratios were strongly influenced by the local attitude of the contact. He plotted $Fe^{2+}/(Fe^{2+}+Mg^{2+})$ weight percent in pyroxene against $Fe^{3+}/(Fe^{3+}+Al^{3+})$ weight percent in garnet and found that two of the deposits lie on a reducing trend and two lie on an oxidizing trend. The deposits on the reducing trend are composed of iron-rich pyroxene and iron-poor garnet, and those on the oxidizing trend are composed of iron-rich garnet and iron-poor pyroxene. The characteristic retrograde minerals are amphibole, calcite, and quartz in deposits on the reducing trend and epidote in deposits on the oxidizing trend.

On the basis of field observations, Morgan (1975) drew a schematic section across the pluton to show the relative positions of the skarns (fig. 62). The skarns with reducing trends are in the roof of the pluton above a horizontal contact, whereas those with oxidizing trends are along the sides, one along a near-vertical contact and the other beneath an overhang. Morgan (1975) postulated that the deposits with reducing trends were formed largely from hydrous magmatic fluids that were trapped at the roof of the pluton and that those with oxidizing trends were formed from mixtures of magmatic fluids with CO_2 that was derived by decarbonation of marble. A possible alternative interpretation, which requires isotopic data to be properly evaluated, is that intrusion of the pluton caused convective circulation of fluids in the wallrocks, which would have carried more meteoric water to the sides of the pluton than to the top (where magmatic fluids predominated).

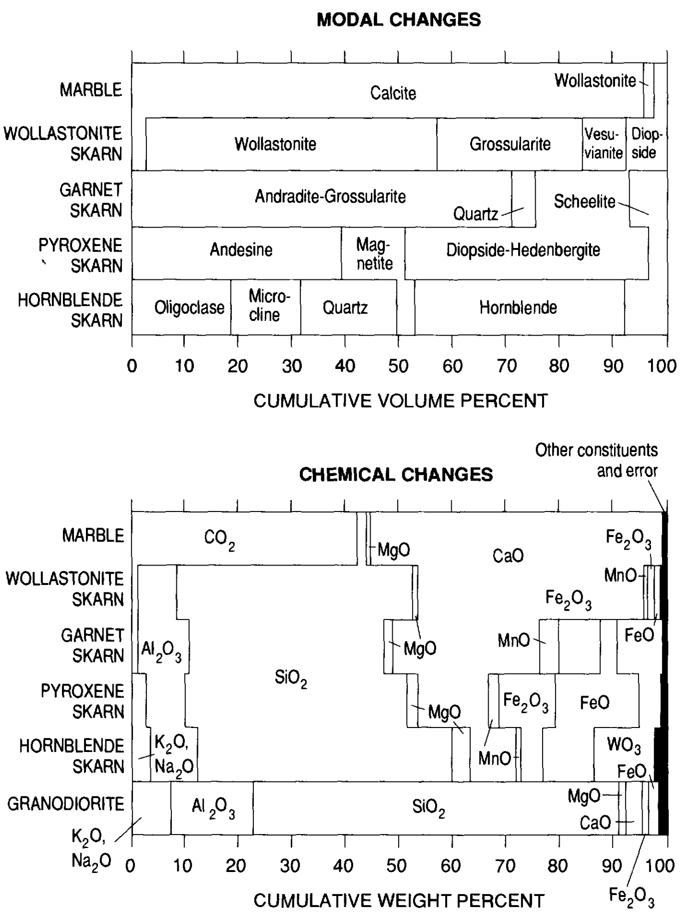


FIGURE 61.—Vertical modal and chemical changes in skarns of ore body No. 7 of Strawberry tungsten mine. Data from Nokleberg (1981b).

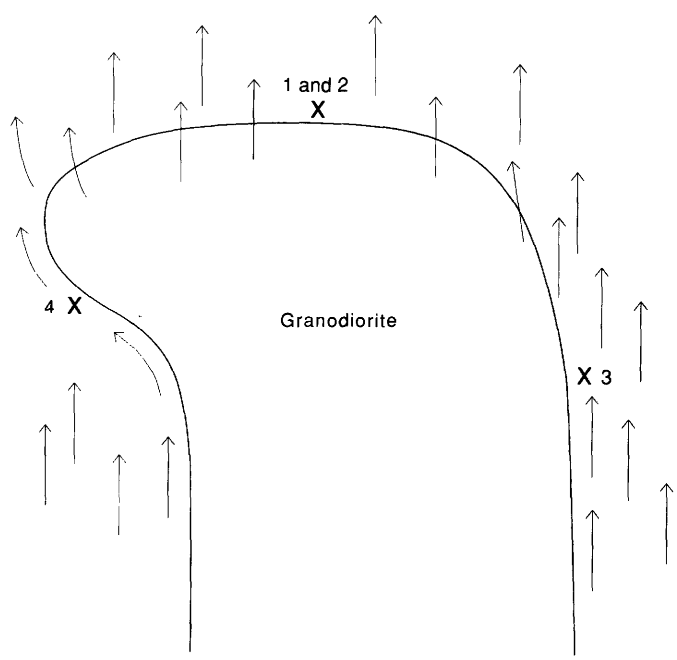


FIGURE 62.—Schematic section across pluton of the Round Valley Peak Granodiorite showing inferred positions of skarn with reducing (1 and 2) and oxidizing (3 and 4) trends. Arrows indicate flow of volatiles. Modified from Morgan (1975).

WORKING MODEL FOR THE ORIGIN AND EMPLACEMENT OF THE BATHOLITH

The larger questions concerning the Sierra Nevada batholith that need to be answered include the following: What was the tectonic setting during the time when the granitic magmas were generated? What were the source materials for the magmas, and how were the magmas generated? What processes are responsible for the diversity of rock compositions within the intrusive suites? How was space made for the intrusions? Why are the Cretaceous intrusive suites successively younger eastward? Why do the potassium content, initial $^{87}Sr/^{86}Sr$, and radiogenic-heat production of the plutonic rocks increase eastward? What is the origin of the structures in the country rocks? What were the patterns of stress and strain during emplacement of the batholith?

The following discussion attempts to provide plausible answers to these and other questions even though drawing conclusions from a collection of seemingly unrelated facts and clues requires bridging gaps with little more than imagination. The goal is to develop a working model that is consistent with the available data and that can be tested—if not now, at least in the foreseeable future. Nevertheless, any attempt to explain the many puzzling questions about the genesis of the Sierra Nevada batholith in the face of present uncertainties, rapidly advancing technology, and changing concepts is bound to be incorrect in some aspects.

PREVIOUS PROPOSALS

An early interpretation (Bateman and others, 1963) was that granitic magmas generated in the lower crust as a result of radioactive decay rose and intruded a complex, faulted synclinorium. At that time, the thinking on orogenic problems was dominated by the tectogene concept of Griggs (1939) and Vening Meinesz (1948). Strata that are progressively younger toward the west in the Mount Morrison and Ritter Range roof pendants (Rinehart and Ross, 1964) and predominantly east-facing bedding tops reported for the western metamorphic belt (Clark, 1964) were cited as primary evidence of a synclinorium. The root of the Sierra Nevada was postulated to have formed concurrently with the synclinorium during the Mesozoic in response to convection currents. The synclinal model gradually fell into disfavor because no single unit had been identified in both limbs, because it became apparent that radioactive decay alone could not supply all the heat required to generate granitic magma, especially in the western part of the batholith where the crust is thin and heat generation low, and because the reported predominance of east-facing bedding tops in the strata east of the Melones fault zone in the western metamorphic belt became suspect, but mostly because plate tectonics displaced the geosynclinal model in the interpretation of orogenic belts.

Among the first plate tectonic models were those of Hamilton (1969) and Moores (1970). Both proposed that ophiolitic and island-arc assemblages were swept against and subducted beneath the North American plate. In these models, heat produced during subduction was assumed to melt both mantle and lower-crustal materials to form granitic magma.

Later, Bateman (1974) revised the geosynclinal model to incorporate an eastward-dipping subduction zone and other plate tectonic features, and Schweickert and Cowan (1975) published a rather complex model involving two partly coeval subduction zones that dipped outward from a core area of oceanic lithosphere. One zone dipped west beneath oceanic crust and produced volcanism in the area of the present western foothills, and the other dipped east beneath continental crust and produced volcanism in the present High Sierra. Both subduction zones originated in the Early Jurassic or Late Triassic, consumed intervening oceanic crust, and eventually collided in the Late Jurassic, causing the Nevadan orogeny. Magma generation then ceased in both arcs, but farther west, eastward subduction was renewed in Late Jurassic time to produce complex structures in the California Coast Ranges and granitic magmas in the Sierra Nevada.

Kistler and Peterman (1973) proposed that the granitic magmas formed above linear zones of melting similar to oceanic-rise systems in a depth range that intersected both the lower crust and the upper mantle and that all the source reservoirs for the magmas had been formed 1,700 Ma. Their studies showed that initial $^{87}Sr/^{86}Sr$ values increase eastward from about 0.704 in the west to about 0.708 in the east. They interpreted the 0.706 isopleth to mark the west limit of Precambrian continental crust and 0.704 to be the east limit of oceanic crust. Later Kistler and Peterman (1978) interpreted the 0.706 isopleth to also mark the east boundary of two stages of rifting that occurred 1,250 to 800 Ma and 600 to 350 Ma, and the 0.704 isopleth to mark the west boundary of mantle material introduced into the lower crust during these stages of rifting. They attributed variations in Rb, Sr, and initial $^{87}Sr/^{86}Sr$ in the granitic rocks to variations in these lower-crustal and upper-mantle source rocks.

Kistler and Peterman (1978) also postulated fracture control for Jurassic magmatism. They suggested that a zone of major disruption (Silver and Anderson, 1974), striking N. 50° W. from near Hermosillo in the Sierra Madre Occidental (Sonora, Mexico) to the southern Inyo Mountains, continues northwestward through the central Sierra Nevada. This zone of disruption follows the axis of Jurassic magmatism south of the map area, and they suggested that it was also the locus of Jurassic magmatism farther north (within and north of the map area). Silver and Anderson (1974) estimated 700 to 800 km of left-lateral offset occurred along this zone between mid-Triassic and mid-Jurassic time.

In contrast to the model of Kistler and Peterman's (1973, 1978), DePaolo (1981) interpreted the isotopic variations within the batholith to reflect different proportions of mantle and crustal components in the granitic magmas. He assumed the mantle component to be high-alumina basalt that was depleted in incompatible elements and the crustal component to be Paleozoic and Proterozoic sedimentary strata. He dismissed the idea that material carried down a subduction zone was a possible source for the magmas, partly because the largest contribution of subducted material would be required for magmas most distant from the trench and partly because neodymium isotopic ratios do not support such a source. Specifically, hydrothermally altered oceanic crust has high positive values of $\varepsilon Nd_{(T)}$, whereas the granitic rocks have negative or low positive values.

DePaolo (1981) calculated the proportions of crustal and mantle material in the batholith as a whole using oxygen and neodymium isotopes. Assuming $\delta^{18}O$ values of +15 for the crustal component, +6 for the mantle component, and +9 for the batholith as a whole, he calculated that the batholith is composed of two-thirds mantle component and one-third crustal component. For Nd, assuming the crustal component ($\varepsilon Nd_{(T)} \approx -12$) to be 2 to 3 times greater than the mantle component, he estimated the crustal contribution to be about 20 to 30 percent, in general agreement with his estimate from oxygen isotopes (DePaolo, 1981). In an earlier report, DePaolo (1980) had estimated that 50 to 70 percent of the Nd in some rocks, especially from the eastern part of the batholith, may have been derived from older crust whereas some samples from the western part contain virtually no Nd from the crust and represent material wholly extracted from the mantle and added to the crust.

Saleeby (1978) proposed large-scale wrench movements and oblique subduction, which caused more westerly blocks to move northward relative to more easterly blocks. He proposed that early Mesozoic strata were deposited across a major suture, called the foothills suture, which truncated sialic crust (Paleozoic and older strata) to the east and brought it into contact with late Paleozoic to early Paleozoic oceanic crust, including ophiolite, to the west. The ophiolites have isotopic ages of 300, 200, and 160 Ma (Saleeby, 1982). Dextral wrench faulting along the continental margin dominated during the Triassic and incorporated an oblique convergent component during the Late Triassic and Early Jurassic. Saleeby and others (1978) inferred from the presence of a thick section of clean radiolarian chert that the ophiolites originated and remained for some extended time beyond the depositional area of continental sediments. Nevertheless, Saleeby (1982) later postulated that the 200- and 160-Ma ophiolites originated close to the continental margin in basins formed by early Mesozoic accretion of the late Paleozoic sea floor.

Influenced by interpretations of suspect terranes of Alaska (Jones and others, 1981), Nokleberg (1983) proposed a model for the country rocks of the Sierra Nevada that involves accretion of six northwest-trending, fault-bounded tectonostratigraphic terranes, each composed of distinct lithologic assemblages having separate structural histories. He postulated that these terranes were amalgamated and accreted sequentially to the North American plate during the Triassic and Jurassic, prior to the Late Jurassic Nevadan orogeny.

More recently, Tobisch and others (1986) postulated extension and listric faulting rather than convergence, subduction, or wrench faulting to explain several repeated fault-bounded, steeply dipping and west-facing, essentially homoclinal sequences in the Sierra Nevada's eastern belt of metavolcanic rocks. (These sequences are part of the west-facing strata that form the east limb of the synclinorium central to the earlier geosynclinal model (Bateman and others, 1963).) Some of these rocks have Early Cretaceous isotopic ages, so the faulting must have occurred after the compressional events that are attributed to the Nevadan orogeny. They postulated it to have resulted from regional extension during the timespan when the Cretaceous granitoids were emplaced. They reasoned that the upward flow and lateral swelling of the granitoid magmas caused cleavage to develop in the country rocks.

PRESENT RECONSTRUCTION

Magmatism began in the central Sierra Nevada in the Late Triassic (~210 Ma), approximately coincident with the breakup of Pangaea, and continued episodically into the Late Cretaceous (~85 Ma) as the North American continent moved northward. The worldwide association of batholiths having the wide compositional range of the Sierra Nevada batholith with zones of convergence at continental margins leaves little doubt that they are related in a fundamental way. Ophiolites present in the western foothills of the Sierra Nevada (Saleeby, 1982) and the Franciscan Complex of the California Coast Ranges, which is widely interpreted to be an accretionary wedge, indicate the presence of one or more subduction zones along the west coast of North America during the Mesozoic. The Sierra Nevada batholith probably was emplaced within the margin of the North American plate above an east-dipping subduction zone that generated linear belts of mantle-derived high heat flow and magmatism, and corresponding uplift, extension, and faulting; this deformation, in turn, exerted control over the rise of magma in the upper crust. The crosscutting regional trend of Cretaceous magmatism relative to the trend of Jurassic magmatism may reflect counterclockwise rotation of the North American plate relative to the underlying linear heat source near the end of the Jurassic, possibly coincident with the Nevadan orogeny. Although geologic, paleontologic, and paleomagnetic data show that some terranes along the west coast of North America have moved relatively northward great distances, no compelling evidence has been recognized that requires the presence of far-traveled terranes within the area of the Mariposa 1° by 2° quadrangle.

Isotopic dating indicates that within the area of the Mariposa 1° by 2° quadrangle only one locus of magmatism existed at any one time during the Cretaceous and that the locus shifted episodically (fig. 54). The locus of magmatism also shifted episodically during the Jurassic, but it is less certain that only one locus existed at all times. The shifts in the locus of magmatism and accompanying renewed magmatic activity probably were caused by events at the plate boundary such as the arrival of exotic terranes, advance or retreat (rollback) of the subducting slab, or changes in the rate of subduction or in the dip angle of the subducting plate (see section "Shifts in the Locus of Magmatism").

Events violent enough to cause the locus of magmatism to shift could also have caused deformation of the country rocks, which isotopic dating shows occurred at several different times during the Mesozoic and was not confined to the Late Jurassic Nevadan orogeny as has been widely assumed. Apparently most of the plutonic rocks lack regional structures related to their emplacement; therefore if the deformations of the country rocks were regional, they were largely confined to times when no intrusions were being emplaced—before, between, or after intrusive episodes. The plutonic rocks may have acted as buttresses during the deformations because of their greater competence and been so little deformed that no obvious structures were imposed on them. However, the Jurassic tonalite of Granite Creek apparently was deformed during the Nevadan orogeny, long after it was emplaced. In these interpretations, tectonic and magmatic responses to changes in activity at the convergent-plate margin were almost immediate rather than delayed for extended periods as in a model proposed by Toksöz and Bird (1977) in which tens of millions of years may be required for the completion of a cycle.

Alternatively, the rise of magma in a new location could have caused deformation of the adjacent country rocks. Clearly, magma bodies that ballooned as they were emplaced—such as the Tuolumne Intrusive Suite, the north ends of the Mount Givens and Lake Edison Granodiorites, the Ward Mountain Trondhjemite, and the megacrystic granite of Birch Creek, and bulges and tongues in the margins of the Round Valley Peak and Dinkey Creek Granodiorites, the Bass Lake Tonalite, and the megacrystic granite of Papoose Flat—disturbed their wallrocks during intrusion. However, elongate intrusions that parallel the northwest-trending regional structure show no obvious evidence of having displaced the adjacent country rocks or of having produced new cleavages and lineations within them. Nevertheless, detailed study could reveal the presence of such structures.

No studies have been carried out to determine whether the ballooning intrusions produced new cleavages and lineations in the bordering stratified country rocks or merely reoriented preexisting structures, but structures in older plutonic rocks bordering these intrusions suggest that new cleavages could have formed. Ballooning of the Tuolumne Intrusive Suite caused ductile foliation and shearing in adjacent parts of the older intrusive suite of Yosemite Valley as it was bowed outward, whereas the tongue of Dinkey Creek Granodiorite that penetrates the Bass Lake Tonalite bowed foliations in the Bass Lake Tonalite without producing ductile structures. As the Bass Lake Tonalite is about 10 m.y. older than the Dinkey Creek Granodiorite, it undoubtedly was completely crystallized when the Dinkey Creek magma intruded the Bass Lake Tonalite. The most likely explanation for this relationship is that the Dinkey Creek magma heated the Bass Lake Tonalite to above its solidus.

The generation of intrusive suites involved melting and mixing of source materials, chiefly in the lower crust, and differentiation processes in the upper crust. Magma generation began with the rise of heat and magma (chiefly basalt) from the mantle into the lower crust. Some mantle-derived magma continued to rise with little or no contamination by crustal material and solidified in the upper crust as small intrusions or erupted at the surface. Initial $^{87}Sr/^{86}Sr$ values of 0.70371 for the eastern part of the greenstone of Bullion Mountain (Bateman and others, 1985) and 0.7032 for two intrusions just west of the map area (pl. 1), the Guadalupe igneous complex and a body of trondhjemite (Kistler and Peterman, 1973), indicate a preponderance of source material that was either derived directly from the mantle or that had been in residence in the crust for only a short time. However, none of the many widely distributed small bodies of diorite and gabbro, which are the rocks most likely to have crystallized from little-contaminated mantle or lower-crustal magma, have been studied isotopically.

Most of the mantle-derived magma probably mixed and mingled complexly with lower-crustal material—melting less-refractory material, reacting with more-refractory material (Bowen, 1928), and mixing with anatectic magmas generated within the crust. Differences in temperature and viscosity undoubtedly made these processes extremely complex (Sparks and others, 1984). Because of decreased density and increased volume, the variously hybridized magmas rose buoyantly into the upper crust. The inward increase of initial $^{87}Sr/^{86}Sr$ in the margins of the Tuolumne Intrusive Suite shows that the magma that formed the outer part of the suite continued to assimilate crustal material as it was emplaced.

The continued rise of heat from the mantle by conduction or by underplating with basaltic magma caused sufficient melting (a minimum of 30–35 percent, according to Van der Molen and Paterson, 1979) of less-refractory lower-crustal materials for them to become mobile and to rise into the upper crust as anatectic magmas. The interior of the Tuolumne Intrusive Suite may be an example of rock that solidified from such anatectic magma. If so, initial $^{87}Sr/^{86}Sr$ for the lower crust in the Tuolumne Meadows region about 90 Ma was about 0.7064. Other, perhaps more creditable examples of intrusions that crystallized from anatectic magmas are the Cretaceous granites of the northern White Mountains, where the crust is thicker, initial $^{87}Sr/^{86}Sr$ values are significantly higher (>0.710 in some intrusions), and cogenetic tonalites and granodiorites, which would suggest differentiation from a more femic parent, are absent.

As the magmas rose into the upper crust, temperatures decreased and crystallization became the dominant process. Crystals were sorted by the affiliated processes of crystal fractionation and crystal-liquid fractionation. As used herein, crystal fractionation is the separation of crystals from the coexisting melt phase by such processes as sidewall accretion, settling, and filter pressing, whereas crystal-liquid fractionation is the precipitation of minerals that contain different proportions of elements than the coexisting melt phase in cooler, generally outer parts of a magma chamber, and replenishment of the elements depleted in the coexisting melt from hotter interior parts of a magma by convection of the magma and diffusion in the melt phase. Minerals such as pyroxene, hornblende, and biotite contain more iron and magnesium than coexisting melt, and plagioclase contains more calcium. The result is concentration of the elements that are relatively enriched in early-forming crystals in the cooler parts of the magma and of all the other elements in the hotter parts. Crystal-liquid fractionation must operate wherever a thermal gradient exists within a cooling and crystallizing magma.

Accretion of crystals at the cooler margins of a magma chamber apparently is responsible for the inward compositional changes in the margins of the Tinemaha Granodiorite and the granodiorite of McMurry Meadows, where initial $^{87}Sr/^{86}Sr$ remains constant. The efficiency with which early-precipitated crystals were concentrated in the margins of these intrusions, especially in the granodiorite of McMurry Meadows, suggests that these processes also were operative and perhaps dominant in the margins of the Tuolumne Intrusive Suite, where isotopic data indicate mixing. Sawka's (1985) proposal that sidewall accretion of crystals in the Tinemaha Granodiorite led to the upward streaming of the crystal-depleted, less-dense adjacent magma, as suggested by laboratory experiments using saline solutions (McBirney, 1980; Turner, 1980), provides a reasonable explanation for the downward increase in the density and mafic-mineral content of the upper part of the Tinemaha Granodiorite and may also explain the vertical gradients inferred to exist in the Dinkey Creek and Mount Givens Granodiorites (see sections "Shaver Intrusive Suite" and "Mount Givens Granodiorite," under "Patterns Within Intrusive Units," and figs. 37 and 38).

The presence of aplites and leucocratic intrusions of the same isotopic age and assigned to the same intrusive suite as enclosing intrusions of tonalite and granodiorite but that show no common mineralogical or textural characteristics and no transitions one to another requires that residual pockets of felsic magma were concentrated at depth within the tonalitic and granodioritic magmas and later remobilized. Examples of felsic intrusions having these relations are the Ward Mountain Trondhjemite and the Knowles Granodiorite of the Fine Gold Intrusive Suite. It is inconceivable that any readily assimilable crustal rocks survived generation of the enclosing Bass Lake Tonalite magma. Residual core zones of felsic magma in the deep interior of the tonalite and granodiorite magmas as a result of crystal-liquid fractionation during inward solidification seems the most reasonable explanation for later emplacement of these felsic magmas.

Space for the batholith was provided in a variety of ways, chiefly by the incorporation of crustal materials in the magmas (as indicated by isotopic data), uplift and erosion of the batholith to depths of 3.5 to 7 km (as indicated by geobarometers), expulsion of volcanic materials (chiefly in ash flows and Plinian-type eruptions), and possibly moderate extension normal to the axis of the batholith. However, the different chemical compositions (especially K_2O contents) of the Jurassic granitoids on the two sides of the batholith make it unlikely that these granitoids were emplaced adjacent to each other and spread apart the full distance that now separates them during intrusion of the Cretaceous granitoids. Passage of the Independence mafic-dike swarm through the crust, apparently without either producing significant amounts of anatectic magma or mixing with crustal materials to produce parent magmas for granitoids, suggests that extension does not necessarily favor the generation of large reservoirs of granitic magma. Many intrusions made space for themselves by crowding aside and upward the rocks that they intruded, but this mechanism provided additional space for the batholith as a whole only where it caused uplift and erosion of the overlying rocks or spreading of the external walls of the batholith. Some intrusions appear to have made space by stoping, but this mechanism merely redistributes material and does not result in increased space for the batholith as a whole.

A possible modern example of an incipient magmatic arc that may eventually produce an intrusive suite like those that were formed during the Mesozoic is the tectonic zone that extends along the east side of the Sierra Nevada northward through Owens Valley, Long Valley, the Mono Lake basin, and beyond. Owens Valley, Long Valley, and the Mono Basin are downdropped blocks between the westward-tilted Sierra Nevada and the White and Inyo Mountains and other eastward-tilted desert ranges. Both antithetic (mountain-side down) and synthetic (valley-side down) faults are present, and the structural environment is one of uplift and collapse, although lateral extension undoubtedly is involved. Basaltic volcanic rocks, present sporadically along the full length of this zone, may correspond to Mayo's (1941) "basic forerunners" of the Sierra Nevada batholith. A pluton comparable in size to larger Sierran plutons is known to underlie the Long Valley caldera (Bailey and others, 1976; Hill and others, 1985), and another pluton may underlie the Coso volcanic field (Bacon, 1985). However, this setting differs significantly from the setting during the Mesozoic in the absence of a subduction zone, and this may be a critical difference.

Eastward increase in initial $^{87}Sr/^{86}Sr$, decrease in initial $^{143}Nd/^{144}Nd$, and systematic changes in the abundances of various oxides and trace elements in the granitic rocks, especially the eastward increase of K_2O, could reflect (1) an eastward increase in the thickness of the prebatholithic crust, which would cause an eastward increase in the proportion of crustal-to-mantle material in the parent magmas, (2) eastward compositional gradients within the prebatholithic crust, or (3) an eastward transition from depleted upper mantle to less-depleted or undepleted upper mantle.

Wollenberg and Smith (1968, 1970) showed that the potassium, uranium, and thorium contents of the exposed country rocks increase eastward across the Sierra Nevada, though by smaller amounts than in the granitoids, and it is reasonable to suppose that the prebatholithic crust thickened with distance from the continental margin. Moore (1962) has correlated increase of K/Na values in Cenozoic volcanic rocks eastward across the Sierra Nevada with increase in the present crustal thickness, and Condie and Potts (1969) postulated a general linear correlation between the potassium content of Cenozoic volcanic rocks and crustal thickness. A K_2O content of more than 4 percent in the Cretaceous granites of the northern Inyo and White Mountains provides strong support for this interpretation. Initial $^{87}Sr/^{86}Sr$ generally greater than 0.71 (table 2) and an absence of mafic precursors to indicate that these granites are the products of crystal-liquid fractionation of more mafic magmas imply that the source materials for these granites were largely crustal and were similar to the nearby Precambrian rocks. Probably only select crustal materials that contain relatively large amounts of K_2O, Na_2O, and SiO_2 were incorporated in the magmas. Such relatively anhydrous, low-potassium rocks as chert, quartzite, carbonates, and mafic and ultramafic igneous rocks are not likely to have been incorporated in the magmas in significant amounts. Refractory xenoliths derived from sedimentary protoliths in the andesite pipe near Big Creek (Domenick and others, 1983) and in the trachyandesite flow at Chinese Peak (Dodge and others, 1986) lend support to this interpretation.

The Jurassic granitoids of the White and northern Inyo Mountains contrast strongly with the Cretaceous granites in the same general area. Except for the Cottonwood Granite and the granodiorite of Beer Creek, the youngest members of the Soldier Pass Intrusive Suite, their compositions are too mafic for them to have been derived from sources like the nearby Precambrian rock. Nevertheless they have K_2O contents that are as high or higher than the Cretaceous granites. Sylvester, Miller, and Nelson (1978) argued that the Jurassic monzonites in the White and Inyo Mountains are distinct from Sierran granitoids and must have had a mantle source because they contain far too much K_2O and Sr to be on the Sierran trend of eastward-increasing K_2O. Initial $^{87}Sr/^{86}Sr$ values have not been determined for the Jurassic granitoids but probably are much lower than the ratios for the Cretaceous granitoids. The high potassium content of the Jurassic monzonites and quartz monzonites relative to rocks in the western foothills of the Sierra Nevada having approximately the same color index supports the suggestion that the upper mantle is less depleted in the eastern part of the batholith than in the western part.

Whether the initial strontium ratios of the surface rocks are representative of ratios at depth is uncertain. Domenick and others (1983) reported that the Sr and Nd ratios of xenoliths collected from the trachyandesite near Big Creek span the full range of isotopic ratios across the batholith. $^{143}Nd/^{144}Nd$ ranges from 0.51169 to 0.51306 and $^{87}Sr/^{86}Sr$ from 0.7031 to 0.7333. The compositions of several of the xenoliths indicate sedimentary protoliths and the presence of refractory crustal material in the lower crust. The depths at which the different xenoliths resided prior to their incorporation in the trachyandesite is not known, but they can be positioned so that the femic constituents and initial Sr isotopic ratios increase and Nd values decrease with depth. Thus, the isotopic changes with increasing depth may be as large as and in the same range as the isotopic variations across the batholith at the surface. Higher $^{87}Sr/^{86}Sr$ and lower $^{144}Nd/^{143}Nd$ values in the Dinkey Creek Granodiorite than in any of the xenoliths (except for a sillimanite gneiss) suggest that the inferred positions at depth represent the existing arrangement within the crust and mantle rather than the order before the batholith was emplaced.

DESCRIPTIONS OF THE PLUTONIC ROCKS

Summaries of the characteristics of the intrusive suites, lithodemes, and the larger and more important unassigned plutons follow. For most of these, the referenced publications provide expanded descriptions. The intrusive suites are discussed in terms of three separate areas (pl. 1): (1) the west slope of the Sierra Nevada, (2) the eastern Sierra Nevada and the Benton Range, and (3) the White and northern Inyo Mountains. For each area, intrusive suites are described first, in the order of their decreasing ages; unassigned plutons and lithodemes are described after the suites. Ultramafic, gabbroic, and dioritic rocks are described last even though some may be relatively young. In general, the most-complete petrographic descriptions available are of the rocks in the eastern Sierra Nevada where the first studies (1945–1952) were conducted. However, modal and chemical data are both more reliable and more abundant for the western part of the batholith, which was studied more recently. Modal data are presented on pertinent parts of Q-A-P diagrams (fig. 6). Detailed descriptions of some rocks, especially in the Yosemite region, have not yet been published; only abbreviated descriptions of them are presented herein. Symbols in parentheses in section headings are the map unit symbols used in plate 1 and do not necessarily correspond to those used in any map figure.

WEST SLOPE OF THE SIERRA NEVADA

FINE GOLD INTRUSIVE SUITE

The Fine Gold Intrusive Suite, previously called the Fine Gold granitoid sequence by Stern and others (1981), is herein formally named for exposures along Fine Gold Creek. The type area is considered to be the Ward Mountain-Bass Lake area. The Fine Gold Intrusive Suite is the oldest formally named intrusive suite in the western Sierra Nevada. The suite consists of the tonalite of Ross Creek, the extensive Bass Lake Tonalite, the granodiorite of Arch

Rock and other rocks considered to be correlative with the granodiorite of Arch Rock, the Ward Mountain Trondhjemite, and the Knowles Granodiorite. The order of emplacement is from mafic to felsic, but transitions between units and compositional and textural similarities are lacking. The distribution of lithodemes and plutons that compose the suite is shown in figure 63.

The suite is characterized by low contents of potassium and of alkali feldspar. Apparently alkali feldspar began to crystallize very late; it is interstitial to the other minerals and does not occur as megacrysts. The low potassium content and an initial $^{87}Sr/^{86}Sr$ of about 0.7045 indicate source materials that contained a preponderance of mantle material or mantle-derived lower-crustal material that had been in residence in the crust only a short time.

TONALITE OF ROSS CREEK (Kro)

The tonalite of Ross Creek forms a small pluton in the southeastern part of the Fine Gold Intrusive Suite. The rock is fine- to medium-grained, even-textured hornblende-biotite tonalite, locally grading into augite-biotite granodiorite (fig. 64). It has a dark color that is caused by abundant mafic minerals, which make up 16 to 30 percent of the rock (fig. 65), and by unusually dark plagioclase. The rock is intruded by the Bass Lake Tonalite (also of the Fine Gold Intrusive Suite) and by the granodiorites of Whisky Ridge and of the Pick and Shovel Mine area, which are part of the Shaver Intrusive Suite. Stern and others (1981) reported a slightly discordant U-Pb zircon age of 113 Ma for the tonalite of Ross Creek.

BASS LAKE TONALITE (Kbl)

Rock at the west end of Yosemite Valley that was called granodiorite of The Gateway by Calkins (1930) and the Gateway Granodiorite by Evernden and others (1957) has been shown to be in the north end of a tongue of tonalite that is continuous southward with an extensive area of tonalite in the southwestern part of the map area (pl. 1) where the rock was called tonalite of Blue Canyon (Bateman and Busacca, 1982; Bateman and others, 1982; Krauskopf, 1985). Because the name Blue Canyon is preempted in formal lithostratigraphic nomenclature and The Gateway does not appear on current topographic maps and is poorly situated for the type locality of this lithodeme, the rock is herein designated the Bass Lake Tonalite for exposures along the shores of Bass Lake, which are designated its type locality. Most of the rock forms a single large pluton that encloses large masses of metamorphic and plutonic rock, but several smaller plutons assigned to the Bass Lake Tonalite lie farther west within the western metamorphic belt and to the north. The granodiorites of Poopenaut Valley and of Hazel Green Ranch were mapped as separate lithodemes by Dodge and Calk (1986, 1987). The distribution and names of the larger plutons that are separated from the main mass are shown in figure 63.

Typical Bass Lake Tonalite is medium-gray, medium-grained, equigranular tonalite with a conspicuous foliation that is shown both by the preferred orientation of minerals, chiefly hornblende and biotite, and by crudely lens-shaped mafic inclusions. On a Q-A-P diagram, most modes plot in the tonalite field, but many plot in the granodiorite field, and a few plot in the granite and quartz diorite fields (fig. 64). The color index ranges from 4 to 40 but for most samples is 10 to 30 (fig. 65).

Rocks with a high color index generally contain blocky hornblende prisms, whereas rocks with a lower color index contain anhedral hornblende grains that are intergrown with biotite and accessory minerals. Notwithstanding the apparent abundance of hornblende in rocks that contain blocky hornblende crystals, biotite is more abundant than hornblende in most samples. The proportion of hornblende increases with the color index, and in rocks with a color index greater than 30 hornblende is almost as abundant as biotite. Black opaque minerals generally are sparse or absent, but sulfides are present locally. The paucity of magnetite in the tonalite is an unusual feature for a Sierran granitoid and has been interpreted to indicate low oxygen fugacity in the magma (Dodge, 1972).

Stern and others (1981) interpreted a lobe of the Bass Lake Tonalite that is enclosed on three sides by the Oakhurst roof pendant to be a separate and younger intrusion, which they called the Oakhurst pluton, because zircon from this lobe yielded a concordant U-Pb age of 105 Ma and a discordant U-Pb age of 108 Ma. However, the rock in this lobe is identical to rock in other parts of the Bass Lake Tonalite, and isopleths showing variations in specific gravity and in the abundance of minerals cross the zone where a high-angle contact would be expected to occur. A careful search for a contact between the rock in the lobe and tonalite to the east was unsuccessful. For these reasons the rocks called Oakhurst pluton by Stern and others (1981) herein are included in the Bass Lake Tonalite despite the U-Pb ages.

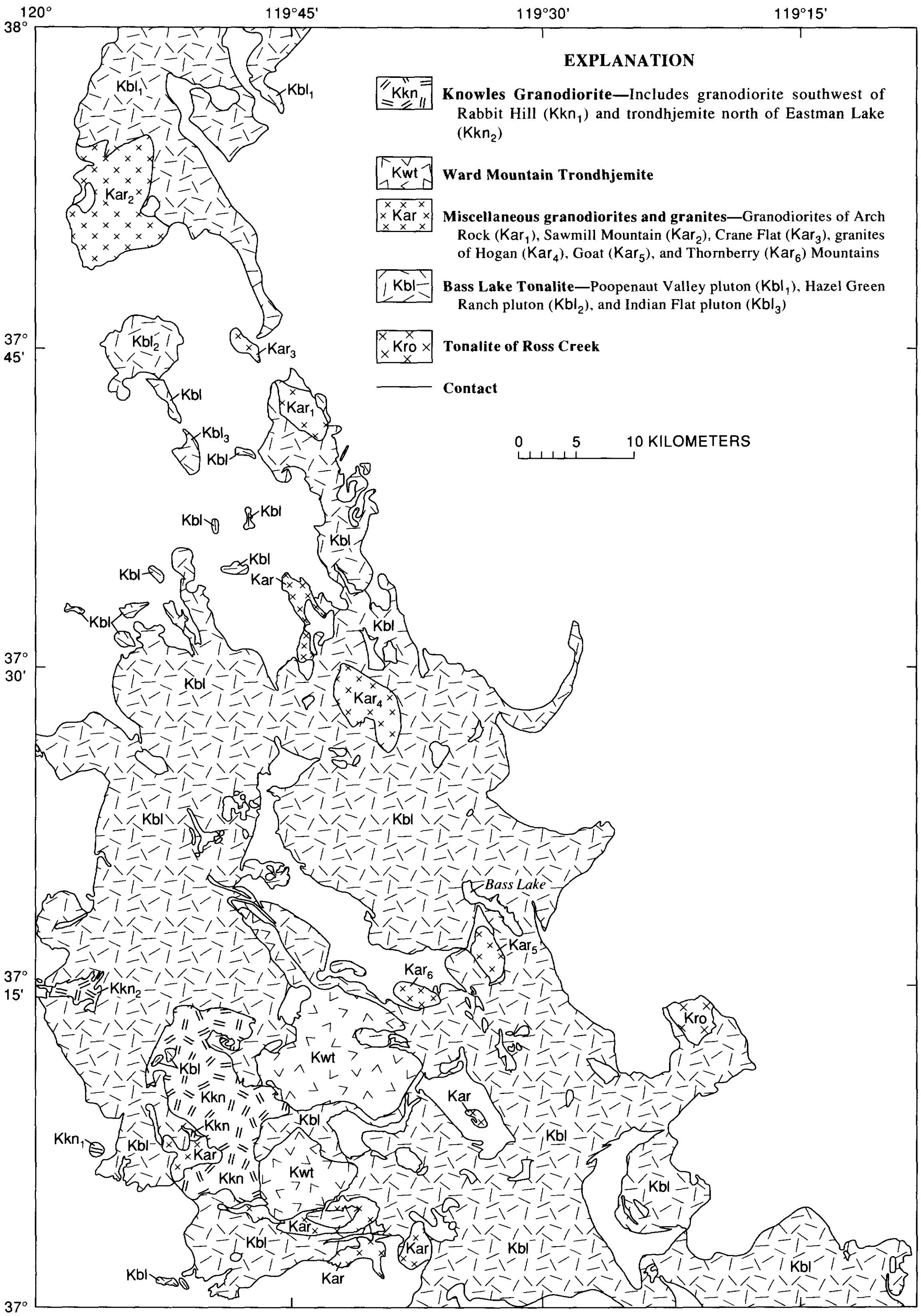


Figure 63.—Generalized geology of the Fine Gold Intrusive Suite.

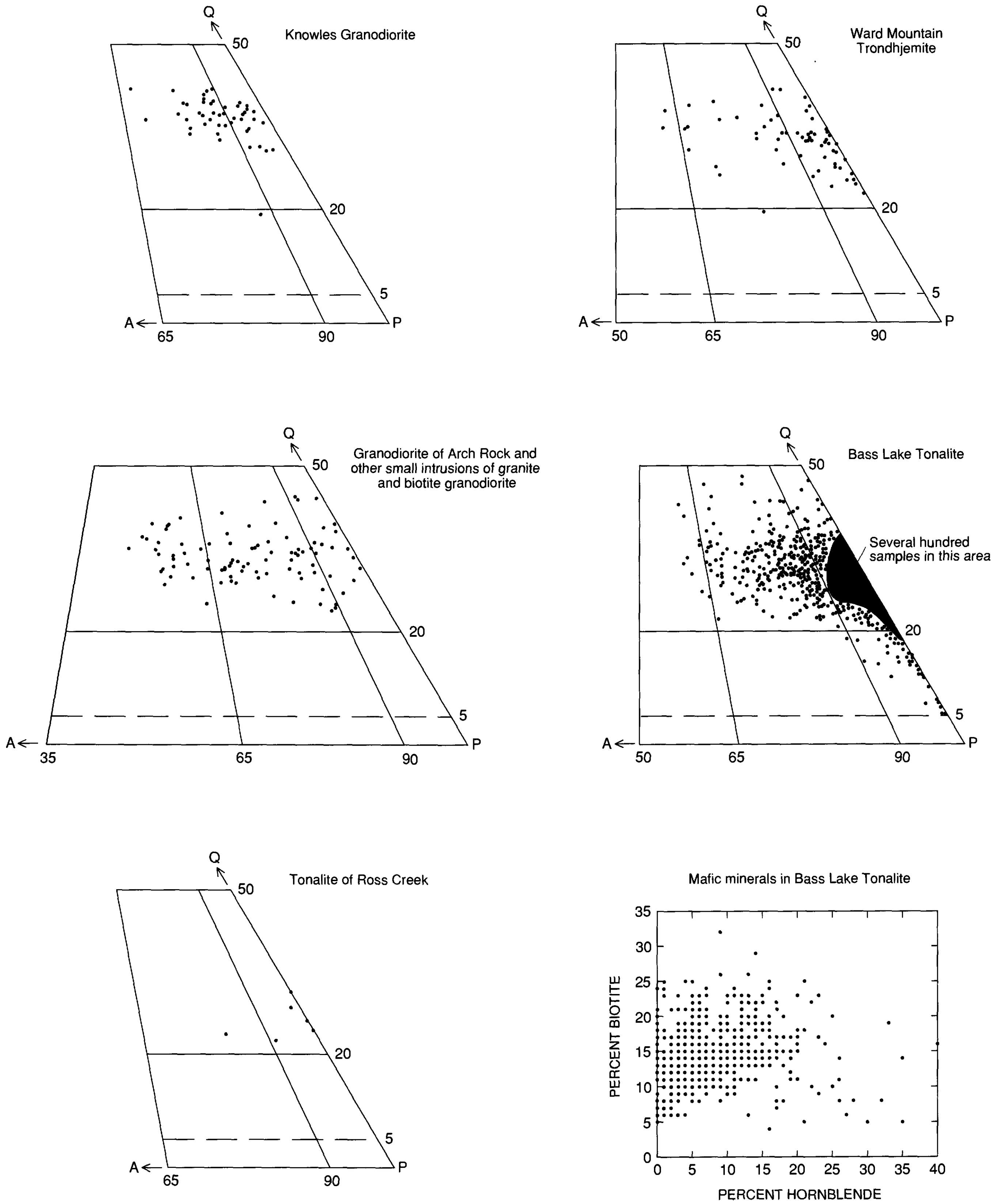


FIGURE 64.—Modal Q-A-P plots of the Fine Gold Intrusive Suite and modal abundances of biotite and hornblende in the Bass Lake Tonalite. See figure 6 for Q-A-P classification diagram.

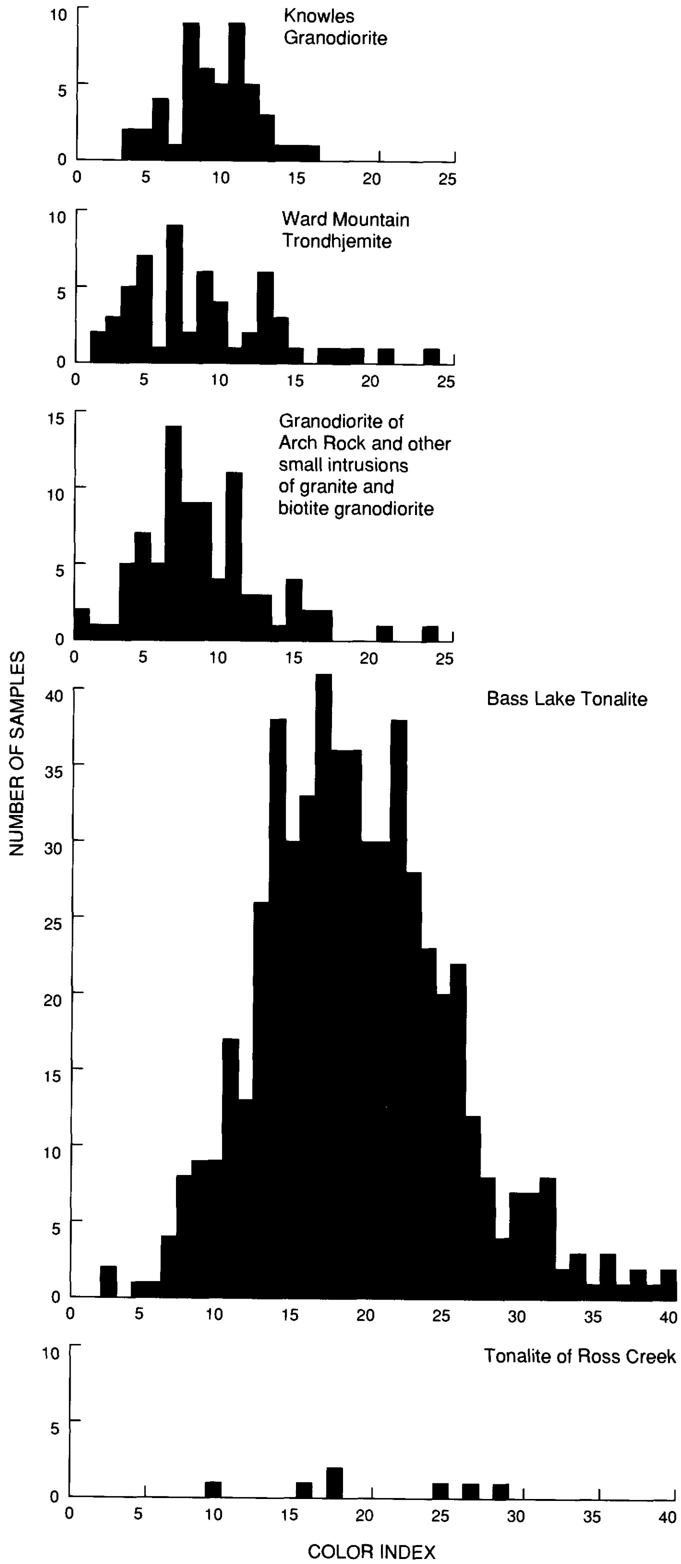


FIGURE 65.—Color indices (volume percent mafic minerals) of the Fine Gold Intrusive Suite.

The Bass Lake is gneissic northeast, east, and southeast of two plutons of the younger Ward Mountain Trondhjemite. A secondary foliation concentric to the trondhjemite plutons and a radial lineation suggest that the deformation in the Bass Lake was caused by the intrusion of the trondhjemite as a highly viscous crystal mush in a late stage of crystallization. This event was referred to as the "Blackhawk deformation" by Bateman, Busacca, and Sawka (1983).

The large size of the Bass Lake Tonalite suggests that it is not a single intrusion, but the absence of obvious contacts or abrupt compositional changes or discontinuities in foliation patterns seems to refute this. Variations in the modal abundance of minerals in conjunction with foliation patterns and the distribution of metamorphic remnants suggest that at a slightly higher level of erosion the tonalite would have appeared to be a series of discrete plutons separated by metamorphic rocks. Steeply dipping magmatic foliations show that the exposed rocks are not at the top of the intrusion, but extensive masses of metamorphic rock within the tonalite suggest that the roof lay not far above. This interpretation is supported by the presence of numerous small outcrops of the tonalite within the southern part of the western metamorphic belt, north and northeast of Mariposa. Undoubtedly the tonalite underlies the metamorphic rocks in this area but is exposed only in the crests of domes and protrusions into the roof rocks.

A 20-km segment of a seismic-reflection traverse (C.M. Wentworth, Jr., unpub. data, 1985) follows roads that extend northeast from the western metamorphic belt through Raymond to the Coarsegold roof pendant. Within 5 km of the western metamorphic belt a prominent nearly horizontal reflector at a depth of about 10 km extends about 5 km from the contact with metamorphic rocks, and a second, shallower reflector dips northeastward from a ~3-km depth at the contact with metamorphic rocks to a ~6-km depth, 2 km to the northeast. Along the central and northeastern part of the traverse, several seismically reflective surfaces at depths of 3 to 8 km dip gently to the southwest. The significance of these surfaces has not been determined, but they suggest the presence of cumulate layers within the tonalite.

The Bass Lake Tonalite intrudes metamorphic rocks, small masses of diorite and gabbro, the Jurassic tonalite of Granite Creek and granite of Woods Ridge, the tonalite of Millerton Lake, and the tonalite of Ross Creek. It is intruded by granitoids assigned to the Shaver Intrusive Suite and the intrusive suite of Yosemite Valley.

U-Pb ages for 13 samples of the Bass Lake Tonalite range from 124 to 105 Ma and average 114 Ma, which is considered the optimum age of the Bass Lake (Stern and others, 1981; Dodge and Calk, 1986). Nevertheless, a large and complex intrusion such as the Bass Lake Tonalite probably cooled to below the blocking temperature for U-Pb determinations over a period of several million years. K-Ar age determinations on biotite range from 114 to 91 Ma and on hornblende from 118 to 102 Ma (Bateman and others, 1976; Evernden and Kistler, 1970; R.W. Kistler, written commun., 1976).

GRANODIORITE OF ARCH ROCK AND OTHER SMALL INTRUSIONS OF GRANITE AND BIOTITE GRANODIORITE (Kar)

The granodiorite of Arch Rock and several other intrusions that are compositionally and texturally similar but may not be exact equivalents of the Arch Rock intrude the Bass Lake Tonalite (fig. 63). The rocks are medium to coarse grained and mostly range in composition from biotite granite to biotite granodiorite and tonalite (fig. 64); a few are hornblende-biotite trondhjemite. The color index generally is 4 to 12 but is higher or lower in a few rocks (fig. 65). The U-Pb age of the granodiorite of Sawmill Mountain is 116 Ma. The granodiorite of Arch Rock has been dated by the K-Ar method on biotite at 95 Ma, but it is intruded by the ~103-Ma (U-Pb) El Capitan Granite, and so its true age must be older. The granite of Hogan Mountain has a fine-grained xenocrystic texture that continues into adjacent parts of the Bass Lake Tonalite; this indicates release of pressure during crystallization and fluxing of both itself and the surrounding tonalite with released water. Obviously this pluton was emplaced at a shallow depth and may have erupted. The age of this pluton is uncertain.

The tonalite south of the Experimental Range, which lies along the south side of the southern pluton of the Ward Mountain Trondhjemite, is composed mostly of distinctive fine-grained, light-colored tonalite that contains biotite but little or no hornblende. All modes plot in the tonalite field on a Q-A-P diagram (fig. 64). This pluton is intruded by the Ward Mountain Trondhjemite.

WARD MOUNTAIN TRONDHJEMITE (Kwt)

The Ward Mountain Trondhjemite is formally named herein for exposures at Ward Mountain, its type locality. It forms two subcircular plutons in the southwest corner of the map area. These rocks have previously been referred to as the leucotonalite of Ward Mountain (Bateman and Busacca, 1982) and as the plagiogranite of Ward Mountain (Stern and others, 1981). Most of the trondhjemite is gneissic. Remnants of undeformed and slightly deformed rock are medium grained, equigranular, and light colored. Biotite is the only mafic silicate present and generally constitutes less than 15 percent of the rock, averaging about 10 percent (fig. 65). Most samples contain less than 5 percent alkali feldspar, most of which occurs as interstitial stringers; where a larger amount of alkali feldspar is present, it forms subequant grains.

Deformed trondhjemite, which constitutes the bulk of the rock, is fine grained and gneissic, and when viewed in thin section it can be seen to have a granoblastic texture. Gneissic foliation is pervasive and generally dips gently east in the eastern margins of the two plutons and steeply east in the western margins. The two plutons are interpreted to be the upward projections of an intrusive body that is continuous at depth. The attitudes of contacts and gneissic foliations in the trondhjemite and of gneissic foliations and lineations in the adjacent metamorphic rocks and in the Bass Lake Tonalite indicate that this composite body is asymmetric; its east flanks dip gently east and its west flanks are vertical or dip steeply east. This unit is believed to have been emplaced as a viscous, largely crystallized mush, which deformed the adjacent rocks as well as itself (Bateman, Busacca, and Sawka, 1983).

The Ward Mountain Trondhjemite intrudes the Bass Lake Tonalite and the pluton of biotite tonalite along the south side of the southern pluton and is intruded by the Knowles Granodiorite. A concordant U-Pb age of 115 Ma was determined by Stern and others (1981) and a K-Ar biotite age of 106 Ma by Naeser and others (1971).

KNOWLES GRANODIORITE (Kkn)

The Knowles Granodiorite, herein formally named for the settlement of Knowles, is light gray, medium grained, and equigranular. In previous publications, this rock was informally designated the granodiorite of Knowles (Stern and others, 1981; Bateman, Busacca, and Sawka, 1983).

The area in the vicinity of the Raymond quarry is designated the type locality. The name is formalized herein because of the economic importance of the rock in this area. For many years it was quarried in large quantities for use as a building stone; the Raymond quarry at the north end of the intrusion is still being operated on a reduced scale. Such notable structures as the Campanile and several other buildings on the Berkeley campus of the University of California, the Fairmont Hotel in San Francisco, and the Los Angeles City Hall are constructed of the Knowles Granodiorite.

The granodiorite appears compositionally and texturally uniform, but a modal plot on a Q-A-P diagram (fig. 64) indicates moderate compositional variation across the tonalite and granodiorite fields. The rock can be seen in thin section to have a hypidiomorphic-granular texture and to lack granoblastic mortar that would indicate deformation during late stages of crystallization or afterward. Tiny flakes of biotite are present in amounts ranging from 5 to 14 percent, and a sprinkling of muscovite generally also is present. Most of the alkali feldspar is interstitial, as it is in other low-potassium rocks of the western foothills. The rock appears structureless in most outcrops, but a faint magmatic foliation is visible locally. The Knowles Granodiorite is considered correlative with a body of similar rock north of Eastman Lake—variously called the trondhjemite north of Eastman Lake (fig. 63), the plagiogranite north of Buchanan Lake (Bateman and Sawka, 1981), and the plagiogranite north of Eastman Lake (Bateman and others, 1982)—and with a small stock southwest of Rabbit Hill, called the granodiorite east of Hensley Lake (Bateman and Sawka, 1981; Bateman and others, 1982). These small plutons are shown to be parts of the Knowles Granodiorite on plate 1 and in figure 63.

The Knowles Granodiorite intrudes the Bass Lake Tonalite and the Ward Mountain Trondhjemite and truncates gneissic structures formed as a result of emplacement of the Ward Mountain Trondhjemite. A U-Pb age of 111.5 Ma (Stern and others, 1981) is mildly discordant but is supported by K-Ar biotite ages of 113, 113, 110, 109, and 107 Ma (Evernden and Kistler, 1970; Bateman and Sawka, 1981). As the Knowles Granodiorite is the youngest intrusion in the southwestern part of the map area, the K-Ar ages are probably fairly reliable.

INTRUSIVE SUITE OF YOSEMITE VALLEY

The intrusive suite of Yosemite Valley (fig. 66), called the Yosemite Valley granitoid sequence by Stern and others (1981), comprises several felsic intrusions that have U-Pb isotopic ages of about 103 Ma. These intrusions include the El Capitan Granite, the granite of Rancheria Mountain, and the correlative Taft Granite and leucogranite of Ten Lakes. The suite is approximately coeval with the Shaver Intrusive Suite farther south, but the two suites are considered separate because no lithodemes common to both suites have been identified. In particular, the intrusive suite of Yosemite Valley lacks a hornblende-bearing granodiorite unit comparable to the Dinkey Creek Granodiorite of the Shaver Intrusive Suite. The relative ages of the El Capitan Granite and the granite of Rancheria Mountain are uncertain, but the Taft Granite and correlative leucogranite of Ten Lakes are younger than both.

EL CAPITAN GRANITE (Kec)

The El Capitan Granite (Calkins, 1930) is a weakly to moderately megacrystic, leucocratic biotite granite. The principal pluton, which crosses Yosemite Valley in the face of El Capitan, is more than 30 km long and averages about 5 km in width. Several plutons in the northern part of the suite, which are herein included in the El Capitan Granite, previously were referred to by different names (fig. 66). These include the granodiorites of Double Rock and of Mount Hoffman (Kistler, 1973; Bateman, Kistler, and others, 1983), the granite of Gray Peak (Peck, 1980, 1983), and the granites of Bald Mountain and of Swamp Lake (Dodge and Calk, 1986, 1987).

Typical El Capitan Granite is light gray and medium to coarse grained. Locally megacrystic or seriate, it contains 1- to 2-cm alkali feldspar megacrysts. Biotite, generally in small books, is the principal mafic mineral and locally is accompanied by small amounts of hornblende. The composition is variable, and modes on a Q-A-P diagram are scattered across both the granite and granodiorite fields; a few plot in the tonalite field (fig. 67). The color index also is variable, ranging from 1 to about 20, but most values range from 4 to 9. Primary magmatic foliation is only weakly discernible, doubtless because of the paucity of mafic minerals. However, a secondary foliation that flagrantly transgresses contacts can be seen locally.

The El Capitan Granite intrudes the Bass Lake Tonalite and the granodiorite of Arch Rock of the Fine Gold Intrusive Suite and the granite porphyry of Star Lakes; it is intruded by all the granitoids that border it on the east. Its age relative to the similar granite of Rancheria Mountain and the granite of Shuteye Peak of the Shaver Intrusive Suite has not been established; Kistler (1973) represented the granite of Rancheria

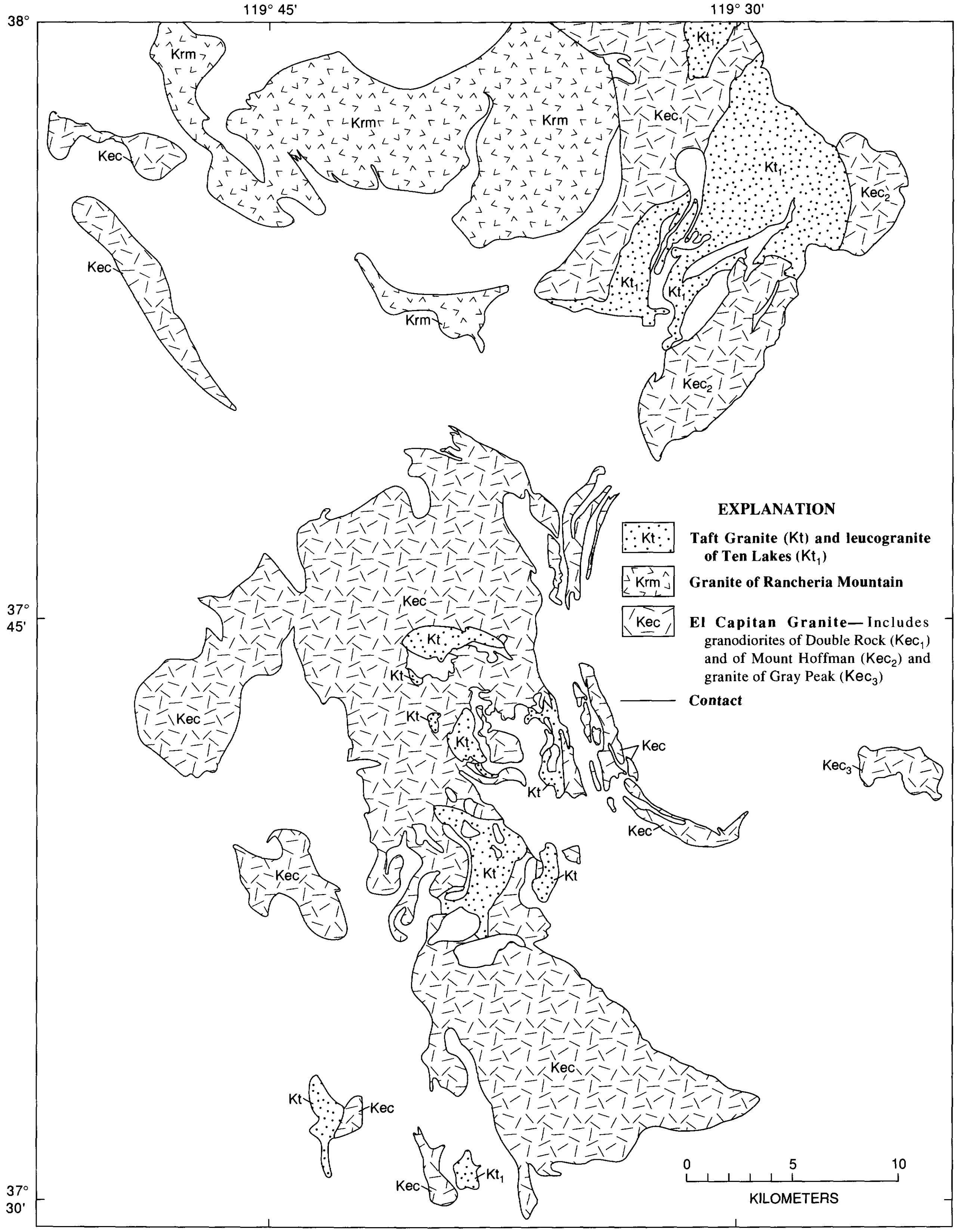


FIGURE 66.—Generalized geology of intrusive suite of Yosemite Valley.

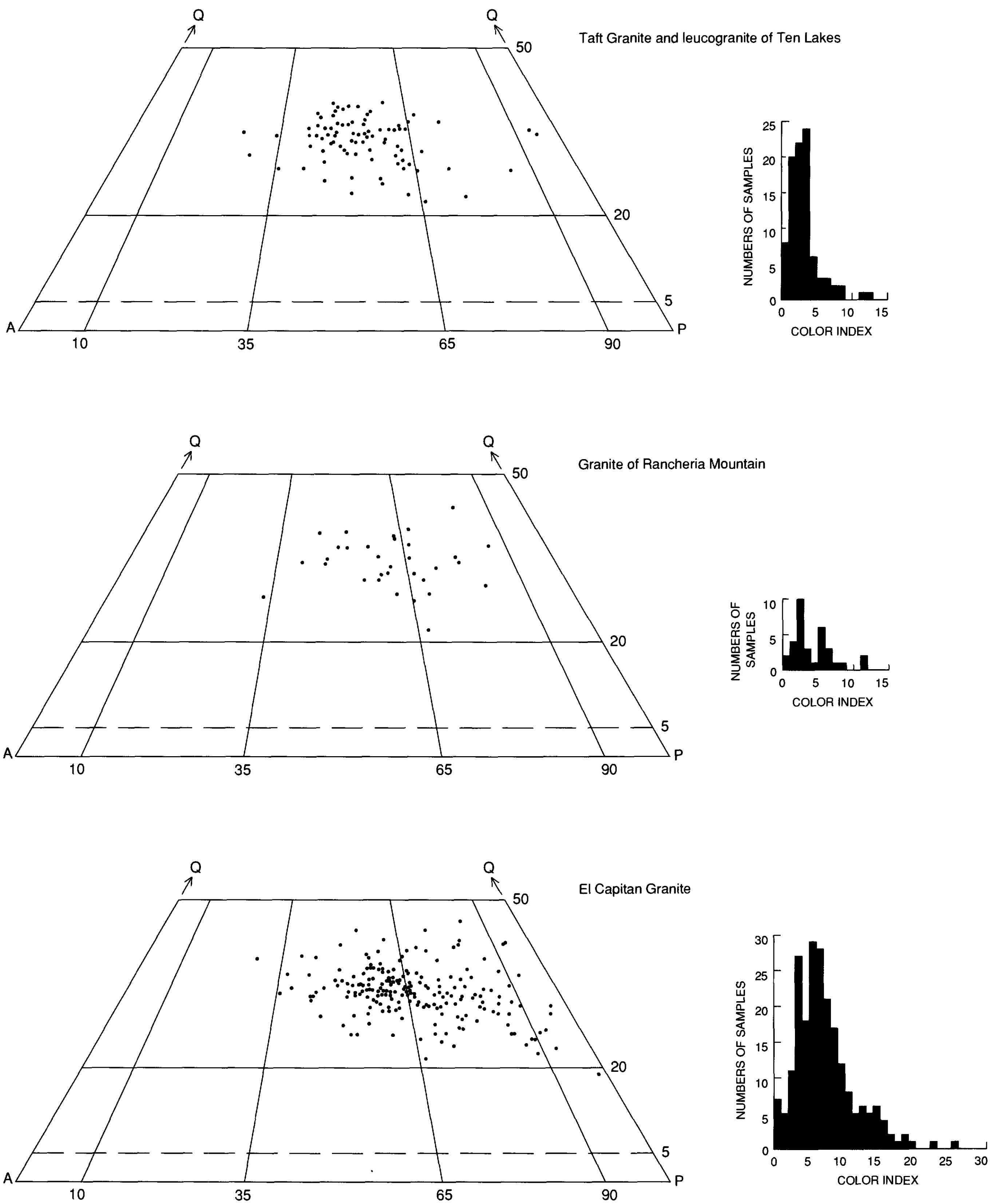


FIGURE 67.—Modes and color indices of intrusive suite of Yosemite Valley. See figure 6 for Q-A-P classification diagram.

Mountain as being younger than the El Capitan Granite. The El Capitan has yielded two concordant U-Pb ages of 103 and 102 Ma, and a discordant age of 97 Ma (Stern and others, 1981; T.W. Stern, written commun., 1983). K-Ar biotite ages of 94, 92, and 85 Ma appear to have been reduced (Curtis and others, 1958; Evernden and Kistler, 1970).

GRANITE OF RANCHERIA MOUNTAIN (Krm)

The granite of Rancheria Mountain is of variable composition and texture and in most places is virtually indistinguishable from the El Capitan Granite. It is not in contact with the main pluton of El Capitan but is in sharp contact with the Double Rock pluton, which herein is included in the El Capitan Granite. Modes of the granite of Rancheria Mountain are distributed across the granite and granodiorite fields on a Q-A-P diagram (fig. 67). The color index ranges from 1 to about 12 but more commonly is in the lower half of this range.

The granite of Rancheria Mountain intrudes the Poopenaut Valley pluton of the Bass Lake Tonalite and is intruded by the granodiorite of Yosemite Creek, the quartz diorite of Mount Gibson, and the granodiorite of Bearup Lake.

TAFT GRANITE AND LEUCOGRANITE OF TEN LAKES (Kt)

The Taft Granite and the leucogranite of Ten Lakes are the youngest and most leucocratic of the granitoids included in the intrusive suite of Yosemite Valley. Widely separated plutons assigned to these units are composed of rock of the same appearance and composition. Typical rock is medium grained, very light gray, and generally has a color index less than 5. On a Q-A-P diagram, most modes plot in the granite field, but a few plot in adjacent parts of the granodiorite and syenogranite fields (fig. 67). The Taft Granite intrudes the El Capitan Granite and is intruded by granitoids assigned to the Tuolumne Intrusive Suite, by the Sentinel Granodiorite and the granodiorite of Yosemite Creek (possible early members of the Tuolumne Intrusive Suite), and by the intrusive suite of Buena Vista Crest. A discordant U-Pb age of 95 Ma was reported by Stern and others (1981) for the Taft Granite, but this age cannot represent the age of most of the unit. The determination may simply be in error, but the rock that was sampled is fine grained, unlike typical Taft, and leaves open the possibility that the sampled rock is not Taft.

SHAVER INTRUSIVE SUITE

The Shaver Intrusive Suite lies east of the Fine Gold Intrusive Suite in the southern part of the map area (pl. 1). Previously it was informally termed the Shaver granitoid sequence by Stern and others (1981). It is herein formally named for exposures around Shaver Lake, and the Shaver Lake-Dinkey Dome area is designated the type area. The oldest and most extensive unit, the Dinkey Creek Granodiorite, is intruded by numerous younger bodies of granite and granodiorite, some of which are assigned to the Shaver Intrusive Suite and others to the younger John Muir Intrusive Suite. Other units assigned to the Shaver Intrusive Suite are the granodiorite of McKinley Grove, the granodiorite of the Pick and Shovel Mine area, the granite of Ordinance Creek, the granite of Shuteye Peak, and the granites of Dinkey Dome, Short Hair Creek, Sheepthief Creek, lower Bear Creek, Mushroom Rock, and north of Snow Corral Meadow (fig. 68). Compared with the Fine Gold Intrusive Suite, the rocks of the Shaver Intrusive Suite are much richer in alkali feldspar and potassium, though not as rich as intrusive suites farther east, and commonly contain magnetite. Isotopic ages indicate that the Shaver Intrusive Suite is the approximate temporal equivalent of the intrusive suite of Yosemite Valley.

In two locations, nested intrusions assigned to the Shaver Intrusive Suite are successively younger and more leucocratic inward. One location is southeast of the Dinkey Creek roof pendant where the granodiorite of McKinley Grove intrudes the Dinkey Creek Granodiorite and is intruded by the granite north of Snow Corral Meadow in a bull's-eye pattern. The other location is west of Huntington Lake (pl. 1) where the successively younger rocks inward are the Dinkey Creek Granodiorite, the granodiorite of the Pick and Shovel Mine area, the granite of Ordinance Creek, and the granite of Mushroom Rock.

DINKEY CREEK GRANODIORITE (Kdc, Kdcm)

The Dinkey Creek Granodiorite, informally referred to as the granodiorite of Dinkey Creek in earlier reports (Huber, 1968; Bateman and others, 1971; Bateman and Wones, 1972a, b; Lockwood and Bateman, 1976), is herein formally named for Dinkey Creek. The south shore of Shaver Lake is designated the type locality. The Dinkey Creek Granodiorite ranges in composition from tonalite to granite; the average composition is granodiorite (fig. 69). The Dinkey Creek consists of two facies—equigranular and megacrystic. In most places the rock is equigranular, but in the northern part, where the rock is more felsic and inclusions of metasedimentary rocks are abundant, and in the margin of the lobe that extends toward the southwest, the rock is megacrystic and contains subhedral alkali feldspar megacrysts, as much as 2 cm long and 1 cm across, set in a medium-grained groundmass. The color index varies markedly over

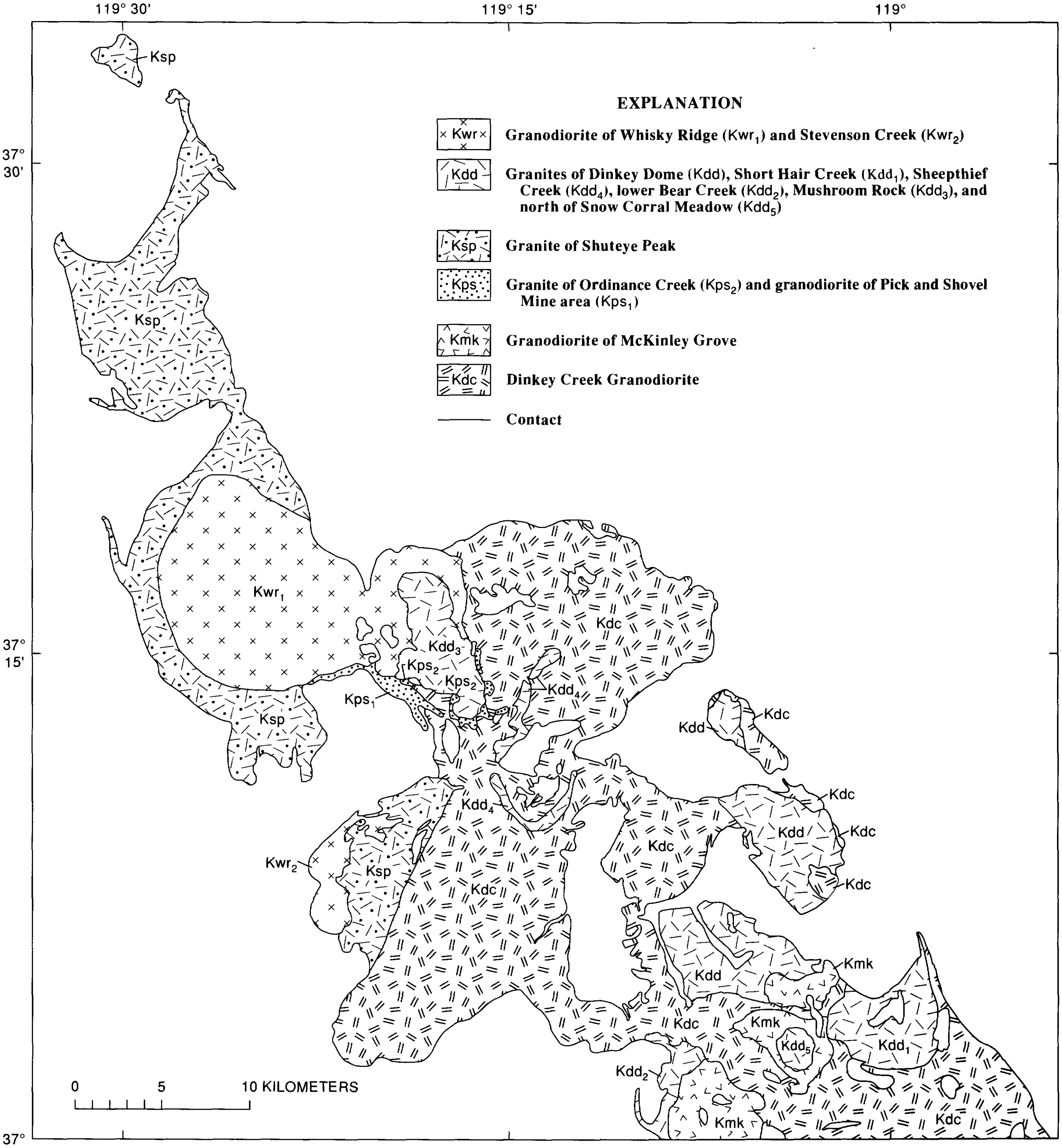


FIGURE 68.—Generalized geology of the Shaver Intrusive Suite and possible correlates, granodiorites of Whisky Ridge and Stevenson Creek.

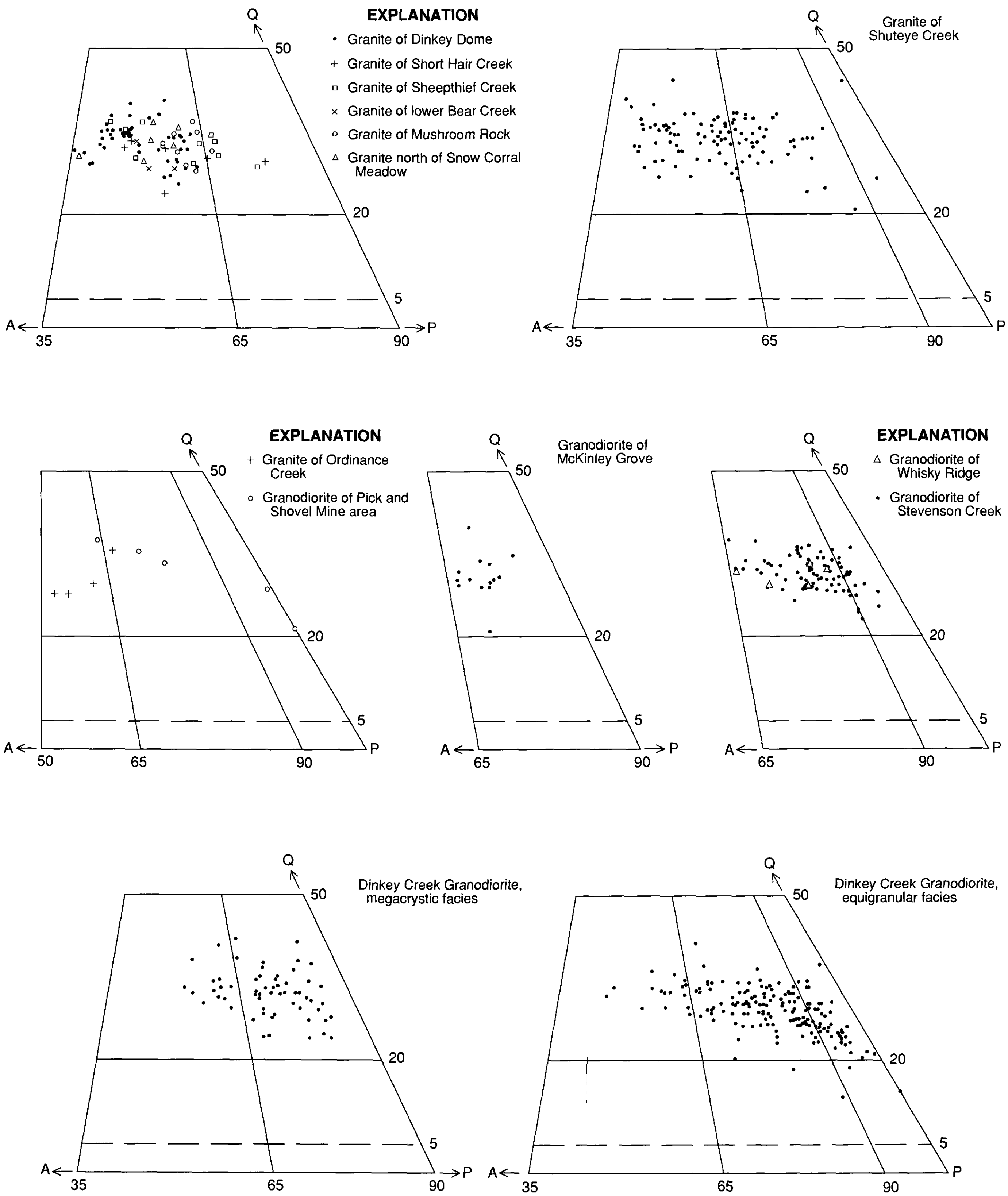


Figure 69.—Modes of the Shaver Intrusive Suite and possible correlates, granodiorites of Whisky Ridge and Stevenson Creek. See figure 6 for Q-A-P classification diagram.

short distances, ranging from less than 5, in a belt of megacrystic and associated equigranular rocks that extends northeast from the east end of Huntington Lake, to more than 25, south of the Dinkey Creek roof pendant (fig. 70). Variations in bulk specific gravity within the Dinkey Creek Granodiorite (fig. 37) closely reflect compositional variations (see section "Shaver Intrusive Suite" under "Compositional Patterns").

Typical equigranular granodiorite such as that exposed along the shores of Shaver and Huntington Lakes is strongly foliated and contains a greater density of mafic clots and inclusions than any other large body of rock within the map area. Almost all the mafic and accessory minerals, including hornblende, biotite, titanite, magnetite, apatite, and zircon, are in clusters that range from inconspicuous clots less than 2 mm across to lenticular mafic

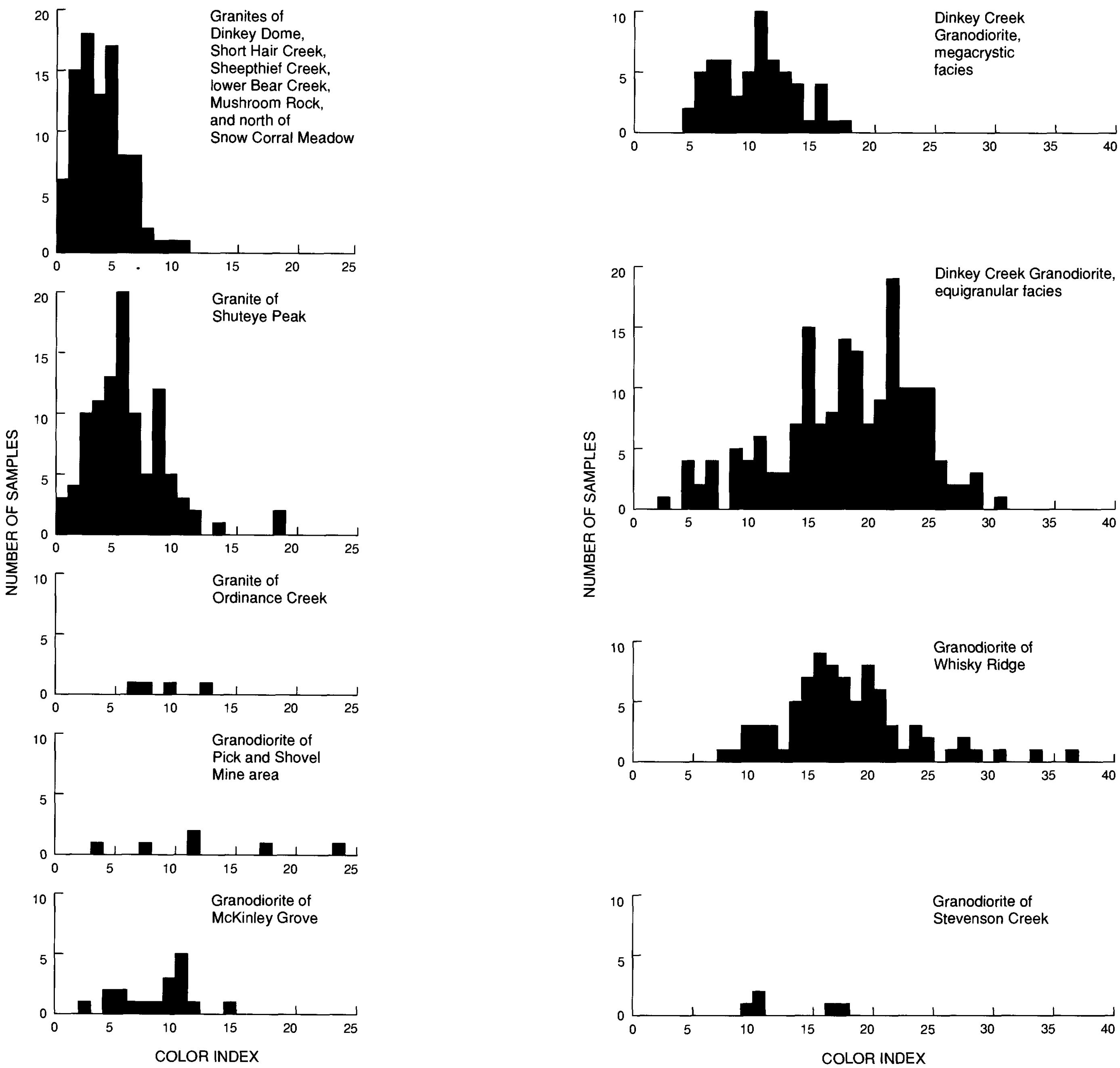


FIGURE 70.—Color indices of the Shaver Intrusive Suite and possible correlates, granodiorites of Whisky Ridge and Stevenson Creek.

inclusions as much as 30 cm long and 5 cm thick. The mafic minerals are intergrown, and individual grains are anhedral. Much of the biotite and hornblende is chloritized. Sporadically distributed euhedral hornblende prisms as much as 1 cm long and biotite books as much as 0.5 mm across contrast with the clots of small anhedral grains of these minerals. The euhedral crystals are believed to have precipitated directly from the melt phase of the magma, perhaps obtaining some constituents from the clots, which probably were formed early in the magmatic history, contemporaneously with the mafic inclusions.

The foliation in the Dinkey Creek Granodiorite dips steeply and follows the external contacts in most places. The foliation bends around most of the younger, small, leucocratic plutons but generally not around small inclusions of metamorphosed sedimentary rock. Obviously the late intrusions have forcibly deflected the foliation in the Dinkey Creek. An absence of granoblastic mortar indicates that either the Dinkey Creek was not completely crystalline when the younger intrusions were emplaced or that heat from the younger intrusion generated an interstitial melt phase in the Dinkey Creek.

The Dinkey Creek Granodiorite intrudes the Bass Lake Tonalite and is intruded by the younger units of the Shaver Intrusive Suite and by the Mount Givens Granodiorite and other units of the John Muir Intrusive Suite. A single discordant U-Pb zircon age of 104 Ma was obtained by Stern and others (1981). Although this age is discordant, it agrees well with a U-Pb age of 102 Ma for the granite of Shuteye Peak, which intrudes the Dinkey Creek Granodiorite and is considered to belong to the Shaver Intrusive Suite (Stern and others, 1981). K-Ar ages for the Dinkey Creek Granodiorite range from 101 to 80 Ma on hornblende and from 94 to 78 Ma on biotite (Kistler and others, 1965; Bateman and Wones, 1972a, b; Bateman and others, 1976).

GRANODIORITE OF MCKINLEY GROVE (Kmk)

The granodiorite of McKinley Grove forms several plutons that intrude the southeastern part of the Dinkey Creek Granodiorite. The granodiorite of McKinley Grove is megacrystic hornblende-biotite granodiorite (fig. 69) and contains conspicuous alkali feldspar megacrysts 2 to 3 cm across in a medium-grained groundmass. It has a lower color index—ranging from 5 to about 11 (fig. 70)—than adjacent equigranular Dinkey Creek Granodiorite and resembles the megacrystic facies of the Dinkey Creek. The granodiorite of McKinley Grove intrudes the equigranular facies of the Dinkey Creek Granodiorite and is intruded by the granite of Dinkey Dome and correlative granites.

GRANODIORITE OF THE PICK AND SHOVEL MINE AREA (Kps) AND GRANITE OF ORDINANCE CREEK (Ko)

The granite of Ordinance Creek and the granodiorite of the Pick and Shovel Mine area are small intrusions that lie along the contact between the older Dinkey Creek Granodiorite and the younger granite of Mushroom Rock (fig. 68). The granite of Ordinance Creek is fine- to medium-grained, leucocratic rock that is characterized by dark spots composed of glomeroporphyritic biotite set in a felsic aplitic matrix. Similar dark spots in felsic dikes that intrude the Bass Lake Tonalite are associated with miarolitic cavities and appear to represent a separate aqueous phase before crystallization was completed. The older granodiorite of the Pick and Shovel Mine area also is fine to medium grained, but its composition is highly variable and ranges from leucogranite to tonalite (fig. 69). It is strongly foliated and contains abundant mafic and metamorphic inclusions.

GRANITE OF SHUTEYE PEAK (Ksp)

The granite of Shuteye Peak forms two plutons that lie just east of the Bass Lake Tonalite of the Fine Gold Intrusive Suite. The larger and more northerly pluton has a highly irregular shape, whereas the smaller and more southerly pluton is ovoid. Typical rock is light colored, medium grained, and equigranular; locally it is weakly megacrystic and contains equant alkali feldspar megacrysts that generally are about 1 cm across. Biotite is the common mafic silicate, but a little hornblende is present in some places. The color index generally ranges from 1 to 12 and averages 6 (fig. 70). Modes plot in both the granite and granodiorite fields on a Q-A-P diagram (fig. 69). This rock resembles the El Capitan Granite of the intrusive suite of Yosemite Valley and is apparently of about the same age. It intrudes the Dinkey Creek Granodiorite and the Bass Lake Tonalite of the Fine Gold Intrusive Suite and is intruded by the granodiorite of Whisky Ridge and the unassigned granodiorites of Camino Creek, Beasore Meadow, and Grizzly Creek. It is in contact with the El

Capitan Granite of the intrusive suite of Yosemite Valley along a short span, but the relative ages of the two units have not been determined. The granite of Shuteye Peak has a concordant U-Pb age of 102 Ma (Stern and others, 1981).

GRANITES OF DINKEY DOME, SHORT HAIR CREEK, SHEEPTHIEF CREEK, LOWER BEAR CREEK, MUSHROOM ROCK, AND NORTH OF SNOW CORRAL MEADOW (Kdd)

These granites are compositionally similar leucocratic biotite granites that differ chiefly in grain size. They are generally more felsic than the granite of Shuteye Peak (fig. 70), and a much larger percentage of their modes plot in the granite field on a Q-A-P diagram (fig. 69). They are not in contact with one another or with the granite of Shuteye Peak, but they intrude all the rocks with which they are in contact. The most areally extensive and coarsest grained of these granites is the granite of Dinkey Dome; the least extensive and finest grained is the granite north of Snow Corral Meadow. The outer part of the granite of Mushroom Rock is medium grained and somewhat megacrystic, whereas the core is aplitic.

The granite of Dinkey Dome comprises three plutons that are separated from one another by younger intrusions of the John Muir Intrusive Suite and probably were emplaced as a single pluton. The rock is equigranular, generally has a color index of less than 5 (fig. 70), and is medium grained in most places. Garnet, andalusite, and sillimanite are locally present adjacent to quartz-mica schist and probably represent contamination. Pegmatitic zones within the pluton northeast of the Dinkey Creek roof pendant locally contain smoky quartz, which is much sought after by mineral collectors. These pegmatitic zones suggest that the magma was saturated with water during final consolidation and that it was emplaced at a shallow level.

K-Ar biotite ages of 88, 88, and 83 Ma for the granite of Dinkey Dome, 90 Ma for the granite of lower Bear Creek, and 87 Ma for the granite north of Snow Corral Meadows (Bateman and Wones, 1972b) suggest that these rocks are distinctly younger than the Dinkey Creek Granodiorite and the granite of Shuteye Peak. However, K-Ar ages of 83 Ma on biotite and of 88 Ma on hornblende from a sample of the Dinkey Creek Granodiorite collected in the same general area indicate that all these K-Ar ages have been reset, probably as a result of the intrusion of the younger John Muir Intrusive Suite.

POSSIBLE ADDITIONAL UNIT OF THE SHAVER INTRUSIVE SUITE

GRANODIORITES OF WHISKY RIDGE AND STEVENSON CREEK (Kwr)

The affiliations of both the granodiorite of Whisky Ridge and the granodiorite of Stevenson Creek are uncertain. The granodiorite of Whisky Ridge intrudes the more felsic granite of Shuteye Peak, contrary to the usual mafic-to-felsic order of emplacement within intrusive suites; consequently its assignment to the Shaver Intrusive Suite is questionable. The principal reason for suggesting its assignment to the Shaver Intrusive Suite is a concordant U-Pb zircon age of 103 Ma which is the same age (within the limits of analytical error) as those for the Dinkey Creek Gran-odiorite and the granite of Shuteye Peak. The granodiorite of Whisky Ridge forms a subcircular pluton that is bounded on three sides by the granite of Shuteye Peak. It closely resembles the Bass Lake Tonalite but has the average composition of granodiorite rather than tonalite (fig. 69). It is medium grained and equigranular, has a color index that generally ranges from 10 to 25 and averages 17 (fig. 70), and contains euhedral hornblende and biotite crystals as well as small clots of anhedral mafic minerals and accessories.

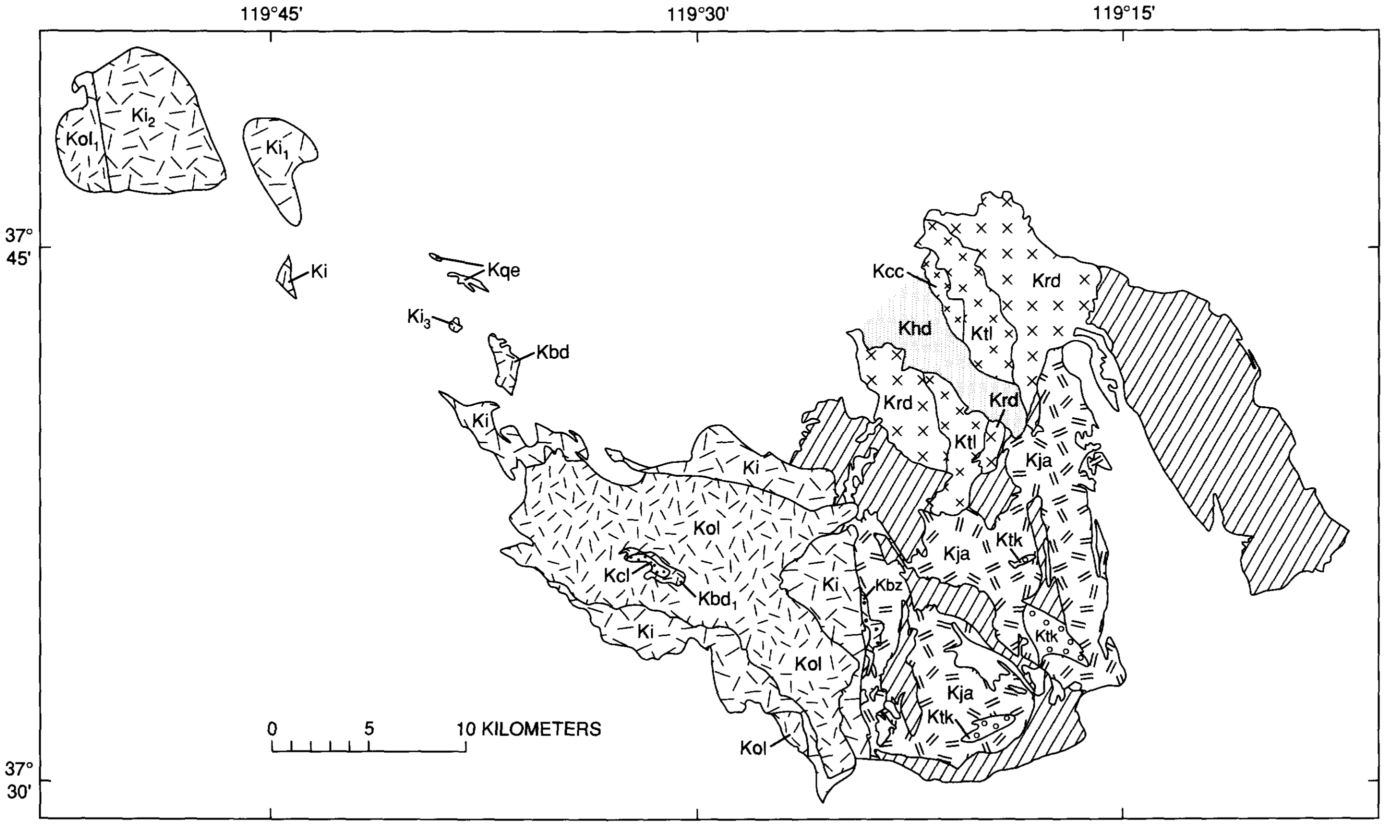
The granodiorite of Stevenson Creek forms an arcuate pluton farther south, west of Shaver Lake, that is composed of medium-grained hornblende-biotite granodiorite, somewhat finer grained than adjacent Bass Lake Tonalite. Its intrusive relations are ambiguous. Narrow dikes from the north end intrude the granite of Shuteye Peak, whereas the southern part appears gradational with the Bass Lake Tonalite. Exposures are discontinuous, and it is possible that two different units have been included in the granodiorite of Stevenson Creek—one younger than the granite of Shuteye Peak and possibly of the same age as the granodiorite of Whisky Ridge, and the other a facies of the Bass Lake Tonalite.

INTRUSIVE SUITE OF BUENA VISTA CREST

The intrusive suite of Buena Vista Crest lies mostly east of the principal pluton of the El Capitan Granite, but some masses lie within and north of that pluton. The suite comprises six successively younger map units (pl. 1; fig. 71): (1) the quartz diorite dikes in El Capitan, (2) the granodiorites of Illilouette Creek and Tamarack Creek, the tonalite of Crane Creek, and the Leaning Tower Granite, (3) the granodiorite of Ostrander Lake, (4) the granodiorite of Breeze Lake, (5) the Bridalveil Granodiorite, and (6) the granite of Chilnualna Lake. This suite was formerly referred to as the intrusive sequence of Buena Vista Crest (Peck, 1980) and the Buena Vista granitoid sequence (Stern and others, 1981).

QUARTZ DIORITE DIKES IN EL CAPITAN (Kqe)

Several quartz diorite dikes that intrude the Taft and El Capitan Granites in the face of El Capitan in Yosemite Valley arouse interest because most bodies of gabbro, diorite, and quartz diorite in the Sierra Nevada have been interpreted to be older than the enclosing granitoid. However, dio-



EXPLANATION

Khd — Half Dome Granodiorite

Intrusive suite of Washburn Lake

Kcc — Granite porphyry of Cony Crags

Ktl — Granite of Turner Lake

Krd — Granodiorite of Red Devil Lake

Intrusive suite of Merced Peak

Ktk — Leucogranites of Timber Knob and Norris Creek

Kja — Granodiorite of Jackass Lakes

Intrusive suite of Buena Vista Crest

Kcl — Granite of Chilnualna Lake

Kbd — Bridalveil Granodiorite and granodiorite of Horse Ridge (Kbd_1)

Kbz — Granodiorite of Breeze Lake

Kol — Granodiorite of Ostrander Lake and Hodgdon Ranch pluton (Kol_1)

Ki — Granodiorites of Illilouette Creek and Tamarack Creek (Ki_1), tonalite of Crane Creek (Ki_2), and Leaning Tower Granite (Ki_3)

Kqe — Quartz diorite dikes in El Capitan

Minarets sequence—Includes hypabyssal intrusions

Contact

FIGURE 71.—Generalized geology of intrusive suites of Washburn Lake, Merced Peak, and Buena Vista Crest.

rite magma can be hot enough to mobilize adjacent, more leucocratic granitic rock and obscure evidence of the younger age of the diorite. D.L. Peck (written commun., 1983) considered these dikes to be early intrusions of the intrusive suite of Buena Vista Crest and unrelated to the intrusive suite of Yosemite Valley.

GRANODIORITES OF ILLILOUETTE CREEK AND TAMARACK CREEK, TONALITE OF CRANE CREEK, AND LEANING TOWER GRANITE (Ki)

Except for the quartz diorite dikes in El Capitan, this composite map unit includes the oldest and most mafic rocks of the intrusive suite of Buena Vista Crest. The granodiorite of Illilouette Creek is by far the largest intrusion included in this map unit. It forms several discontinuous bodies that are peripheral to the granodiorite of Ostrander Lake. The other plutons lie farther northwest; the tonalite of Crane Creek and the granodiorite of Tamarack Creek are north of the Merced River. D.L. Peck (written commun., 1983) considered the Leaning Tower Granite, which is a dike complex, to be correlative with the granodiorite of Illilouette Creek. Typical granodiorite of Illilouette Creek is dark, medium-grained, equigranular hornblende-biotite granodiorite and hornblende tonalite that contain 15 to 50 percent biotite and hornblende (figs. 72, 73). Sparse euhedral crystals of hornblende are as long as 10 cm. Plots of modes on a Q-A-P diagram show that the other rocks included in this map unit have somewhat-different compositions. Whereas modes of most samples from the granodiorite of Illilouette Creek plot in the granodiorite field, those of the tonalite of Crane Creek plot in the tonalite field, and those of the Leaning Tower Granite plot in or near the granite field (fig. 72).

The granodiorite of Illilouette Creek intrudes the granite of Shuteye Peak of the Shaver Intrusive Suite and the El Capitan and Taft Granites of the intrusive suite of Yosemite Valley. It is intruded by the granodiorites of Ostrander Lake and of Breeze Lake of the intrusive suite of Buena Vista Crest, by the Mount Givens Granodiorite of the John Muir Intrusive Suite, and by the Sentinel Granodiorite. It has a discordant U-Pb age of 100 Ma (Stern and others, 1981). The tonalite of Crane Creek is intruded by the Hodgdon Ranch pluton of the granodiorite of Ostrander Lake and intrudes the Poopenaut Valley pluton of the Bass Lake Tonalite. The granodiorite of Tamarack Creek intrudes the El Capitan Granite and the unaffiliated tonalite of Aspen Valley.

GRANODIORITE OF OSTRANDER LAKE (Kol)

The granodiorite of Ostrander Lake is the most extensive unit of the intrusive suite of Buena Vista Crest. It is medium-grained, equigranular hornblende-biotite granodiorite and granite. This rock is distinctly lighter colored than the granodiorite of Illilouette Creek, and on a Q-A-P diagram modes plot in the granodiorite and adjacent parts of the granite fields (fig. 72). Most samples contain 5 to 15 percent anhedral hornblende and biotite. The granodiorite of Ostrander Lake intrudes the granodiorite of Illilouette Creek, the El Capitan and Taft Granites of the intrusive suite of Yosemite Valley, and the granodiorite of Jackass Lakes of the intrusive suite of Merced Peak. It is intruded by the Mount Givens Granodiorite of the John Muir Intrusive Suite and by the Sentinel Granodiorite. It has two discordant U-Pb ages of 112 and 107 Ma. Both of these ages are inconsistent with intrusive relations, which indicate an age less than both that of the ~102-Ma El Capitan Granite and that of the ~98-Ma granodiorite of Jackass Lakes. The Hodgdon Ranch pluton, 30 km northwest of the main mass of granodiorite of Ostrander Lake, is composed of similar rock and is tentatively correlated with the granodiorite of Ostrander Lake.

GRANODIORITE OF BREEZE LAKE (Kbz)

The granodiorite of Breeze Lake consists of a small, elongate pluton that lies between the granodiorites of Jackass Lakes and Illilouette Creek and intrudes both of these units. The rock is light-gray biotite granodiorite with a seriate texture. Plagioclase phenocrysts as much as 8 mm long are set in a fine-grained groundmass. Alkali feldspar occurs as small interstitial grains and as poikilitic crystals as much as 1 cm across.

BRIDALVEIL GRANODIORITE AND GRANODIORITE OF HORSE RIDGE (Kbd)

The Bridalveil Granodiorite forms a complex of gently dipping dikes in the vicinity of Bridalveil Falls, and the granodiorite of Horse Ridge is a separate unit that forms a small pluton in the core of the granodiorite of Ostrander Lake. These rocks are mostly fine-grained granodiorite. Modes of a few samples of the granodiorite of Horse Ridge, which generally is a little darker than the Bridalveil Granodiorite, plot in the tonalite and quartz diorite fields on a Q-A-P diagram (figs. 72, 73).

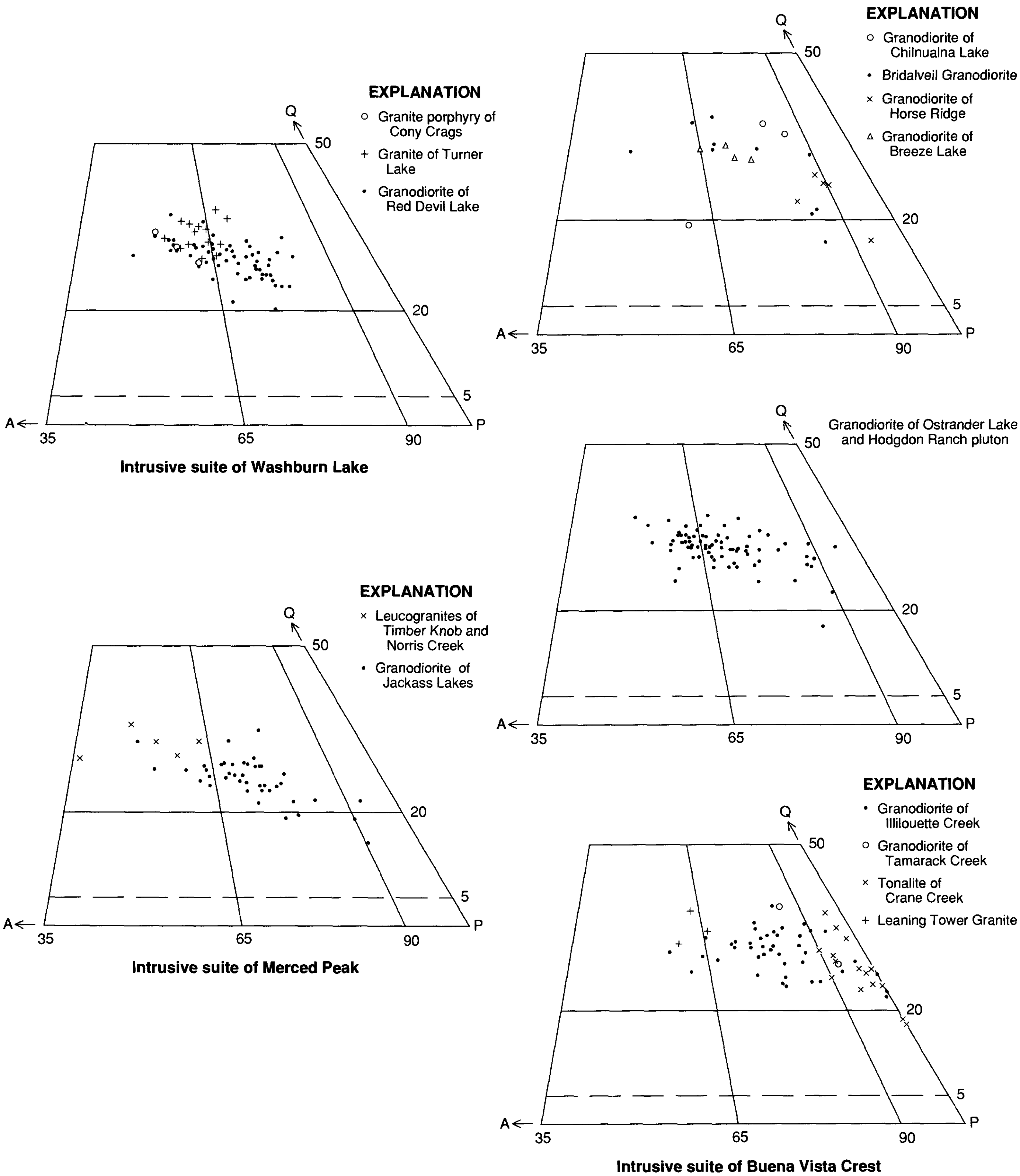


FIGURE 72.—Modes of intrusive suites of Washburn Lake, Merced Peak, and Buena Vista Crest. See figure 6 for Q-A-P classification diagram.

GRANITE OF CHILNUALNA LAKE (Kcl)

The granite of Chilnualna Lake forms a small body of fine-grained granite and granodiorite that is enclosed in the granodiorite of Horse Ridge in the core of the intrusive suite of Buena Vista Crest.

INTRUSIVE SUITE OF MERCED PEAK

The intrusive suite of Merced Peak, formerly called the intrusive sequence of Merced Peak by Peck (1980) and the Merced Peak granitoid sequence by Stern and others (1981), lies within and

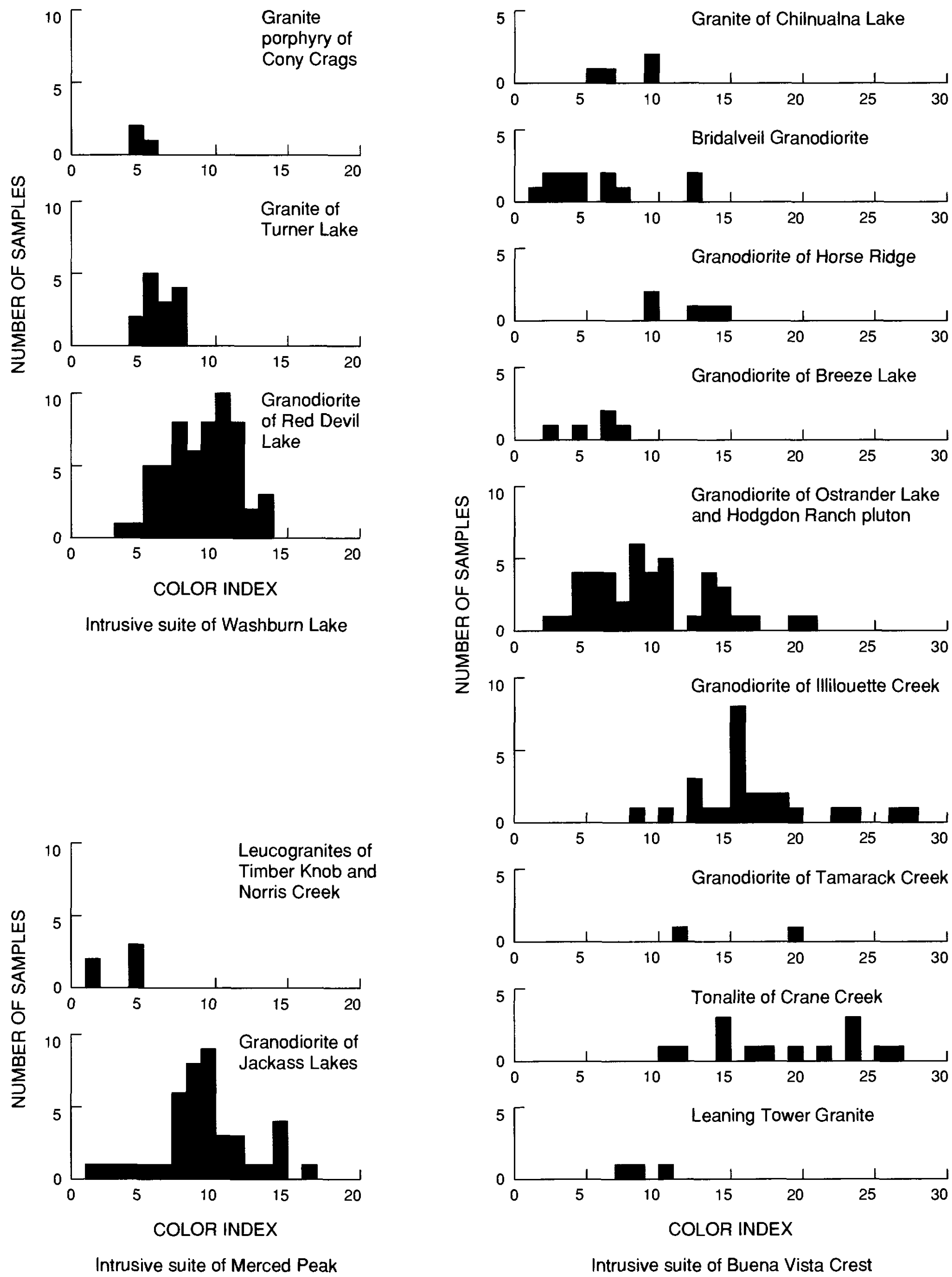


FIGURE 73.—Color indices of intrusive suites of Washburn Lake, Merced Peak, and Buena Vista Crest.

intrudes the volcanogenic Minarets sequence. It comprises the granodiorite of Jackass Lakes and the leucogranites of Timber Knob and Norris Creek (Peck, 1980). The granodiorite of Jackass Lakes makes up most of the suite; the younger leucogranites of Timber Knob and Norris Creek together underlie an area less than 10 km^2 (fig. 71). The older leucogranite porphyries of Red Peak and Post Peak, which Peck (1980) assigned to this suite, are included with other hypabyssal intrusions in the probably cogenetic Minarets sequence.

GRANODIORITE OF JACKASS LAKES (Kja)

The granodiorite of Jackass Lakes complexly intrudes the Minarets sequence and contains many bodies of metamorphic rock ranging from tiny fragments to large roof pendants. In some areas fragments of metamorphic rocks are so abundant as to form intrusion breccias. Typical rock is light-gray, medium-grained granodiorite containing about 10 percent biotite and smaller amounts of hornblende. The mafic minerals occur both as discrete subhedral grains and as anhedral grains in small clots. Plots of modes on a Q-A-P diagram extend across the granodiorite field and into adjacent parts of the tonalite and granite fields (fig. 72). Some felsic variants contain abundant alkali feldspar megacrysts.

The granodiorite of Jackass Lakes intrudes the granodiorite of Illilouette Creek of the intrusive suite of Buena Vista Crest and is intruded by the granodiorites of Ostrander Lake and of Breeze Lake, also belonging to the intrusive suite of Buena Vista Crest, and by the granodiorite of Red Devil Lake and the granite of Turner Lake of the slightly younger intrusive suite of Washburn Lake. Thus the granodiorite of Jackass Lakes is intermediate in age relative to units assigned to the intrusive suite of Buena Vista Crest; it has a discordant U-Pb age of 98 Ma (Stern and others, 1981).

LEUCOGRANITES OF TIMBER KNOB AND NORRIS CREEK (Ktk)

These rocks include small bodies, stockworks, and dike complexes of very light gray, fine-grained, equigranular biotite leucogranite. Modes plot in the granite field on a Q-A-P diagram (fig. 72). These rocks intrude only the granodiorite of Jackass Lakes and are the youngest plutonic rocks in the area in which they occur.

INTRUSIVE SUITE OF WASHBURN LAKE

The intrusive suite of Washburn Lake, formerly called the intrusive sequence of Washburn Lake by Peck (1980) and the Washburn granitoid sequence by Stern and others (1981), comprises the granodiorite of Red Devil Lake, the granite of Turner Lake, and the granite porphyry of Cony Crags. This suite is arranged in a crudely nested pattern with successively more felsic units toward its interior, but it has been disrupted and split into two parts by a tongue of the Half Dome Granodiorite of the Tuolumne Intrusive Suite (fig. 71).

GRANODIORITE OF RED DEVIL LAKE (Krd)

The granodiorite of Red Devil Lake is the outermost unit and the most mafic rock of the intrusive suite of Washburn Lake. It consists of three disconnected bodies. The rock is zoned from inclusion-rich, mafic hornblende-biotite granodiorite in the outer margins to biotite granite at the inner contacts with the granite of Turner Lake. Most modes plot in the granodiorite field, but those near inner contacts plot in the granite field (fig. 72). The granodiorite of Red Devil Lake intrudes the granodiorite of Jackass Lakes and is intruded by the granite of Turner Lake and by the Half Dome Granodiorite of the Tuolumne Intrusive Suite. It has a concordant U-Pb age of 98 Ma (Stern and others, 1981).

GRANITE OF TURNER LAKE (Ktl)

The granite of Turner Lake consists of porphyritic biotite granite and felsic granodiorite. Tabular alkali feldspar phenocrysts 1 to 2 mm long are set in a medium-grained groundmass. The rock intrudes the granodiorite of Red Devil Lake and is intruded by the granite porphyry of Cony Crags and the Half Dome Granodiorite.

GRANITE PORPHYRY OF CONY CRAGS (Kcc)

The granite porphyry of Cony Crags forms a small, elongate pluton at the center of the intrusive suite of Washburn Lake. Sparse phenocrysts of plagioclase, quartz, and alkali feldspar are set in a fine-grained matrix. The rock intrudes the granite of Turner Lake and is intruded by the Half Dome Granodiorite.

RELATIONS OF THE INTRUSIVE SUITES OF BUENA VISTA CREST, MERCED PEAK, AND WASHBURN LAKE TO ONE ANOTHER AND TO THE MINARETS SEQUENCE

The intrusive suites of Merced Peak, Buena Vista Crest, and Washburn Lake have many features in common. They are small compared to other intrusive suites, are closely associated spatially, have similar isotopic ages, and all intrude the only slightly older volcanic and hypabyssal Minarets sequence. Because these features suggest a close, probably genetic relation, their names are not formalized.

Field relations along contacts show that the intrusive suite of Washburn Lake is younger than the granodiorite of Jackass Lakes of the intrusive suite of Merced Peak but that the intrusive suites of Merced Peak and Buena Vista Crest overlap in age. The ages of all three intrusive suites are bracketed between the concordant U-Pb 98-Ma age for the granodiorite of Red Devil Lake and the 100-Ma age for the Minarets sequence (Fiske and Tobisch, 1978; Peck, 1980; Kistler and Swanson, 1981; Stern and others, 1981). Whether these three suites of essentially the same age should be distinguished or included in a single intrusive suite is questionable; clear guidelines for assigning plutons and lithodemes to intrusive suites are still evolving. Nevertheless, their separation is consistent with the separation of the coeval Shaver Intrusive Suite and the intrusive suite of Yosemite Valley and the approximately coeval Tuolumne and John Muir Intrusive Suites.

TUOLUMNE INTRUSIVE SUITE

The Tuolumne Intrusive Suite is the best known and most firmly established intrusive suite in the Sierra Nevada (Calkins, 1930; Bateman and Chappell, 1979). It is also one of the best exposed suites because it is located in high glaciated country in the eastern part of Yosemite National Park. It underlies more than 1,200 km^2 of the High Sierra, but about one-quarter of the suite lies north of the map area.

The units that form the Tuolumne Intrusive Suite were emplaced in a series of surges (fig. 33) and are concentrically arranged—the oldest and most mafic units compose the margins, and the youngest and most felsic unit forms the core. Although contacts between units are sharp and the rocks in contact are distinctly different, progressive inward changes in texture and composition take place within units as well as at contacts (see section "Tuolumne Intrusive Suite" under "Patterns Within Intrusive Units"). Dark, generally fine-grained, well-foliated rocks in the margins give way inward to progressively lighter colored, coarser grained, poorly foliated rocks. Where a magmatic surge removed little of the adjacent older unit, textural and compositional changes across the contact are slight, but where it was transgressive and removed a significant part of an older unit, compositional and textural changes are significant.

The units from margin to core and from oldest to youngest are (1) the correlative granodiorite of Kuna Crest (in the east), tonalites of Glen Aulin and Glacier Point (in the west), and the granodiorite of Grayling Lake (in the south), (2) the Half Dome Granodiorite, (3) the Cathedral Peak Granodiorite, and (4) the Johnson Granite Porphyry (fig. 32). Offset contacts of units of the intrusive suite of Washburn Lake indicate that the tongue of megacrystic Half Dome Granodiorite that extends southeast from the east side of the Cathedral Peak Granodiorite split the intrusive suite of Washburn Lake and forcibly spread it apart. Outward-dipping schlieren (fig. 22) and streaky zones of alternating dark- and light-colored rocks in the margins of the suite indicate that shearing movements and magma mixing occurred during consolidation.

GRANODIORITES OF KUNA CREST AND GRAYLING LAKE AND TONALITES OF GLEN AULIN AND GLACIER POINT (Kk)

The units that form the margins of the Tuolumne Intrusive Suite are discontinuous, and different segments have been given different names. The granodiorite of Kuna Crest lies along the east side, the tonalites of Glen Aulin and Glacier Point lie along the west side, and the granodiorite of Grayling Lake lies along the south side (fig. 32). These rock masses may have been continuous when they were emplaced but later were torn apart and partly reincorporated in magma during the surge that produced the Half Dome Granodiorite.

The rocks in this unit are dark gray and equigranular. Inward from the margins of the suite, the average composition changes from quartz diorite to granodiorite (fig. 74) and the color index decreases from about 30 to about 10 (fig. 75). Because of the equant shapes of constituent grains, foliation is not readily visible in the tonalite of Glen Aulin, but in the granodiorite of Kuna Crest it is shown by lensoid inclusions. In the tonalite of Glen Aulin, the grain size increases inward from less than 3 mm at the outer contact to about 4 mm at the inner contact with the Half Dome Granodiorite. In the outer part of the tonalite of Glen Aulin, the principal minerals are plagioclase, hornblende, and biotite. Some hornblende crystals contain pyroxene cores. Magnetite and titanite grains are less than 0.3

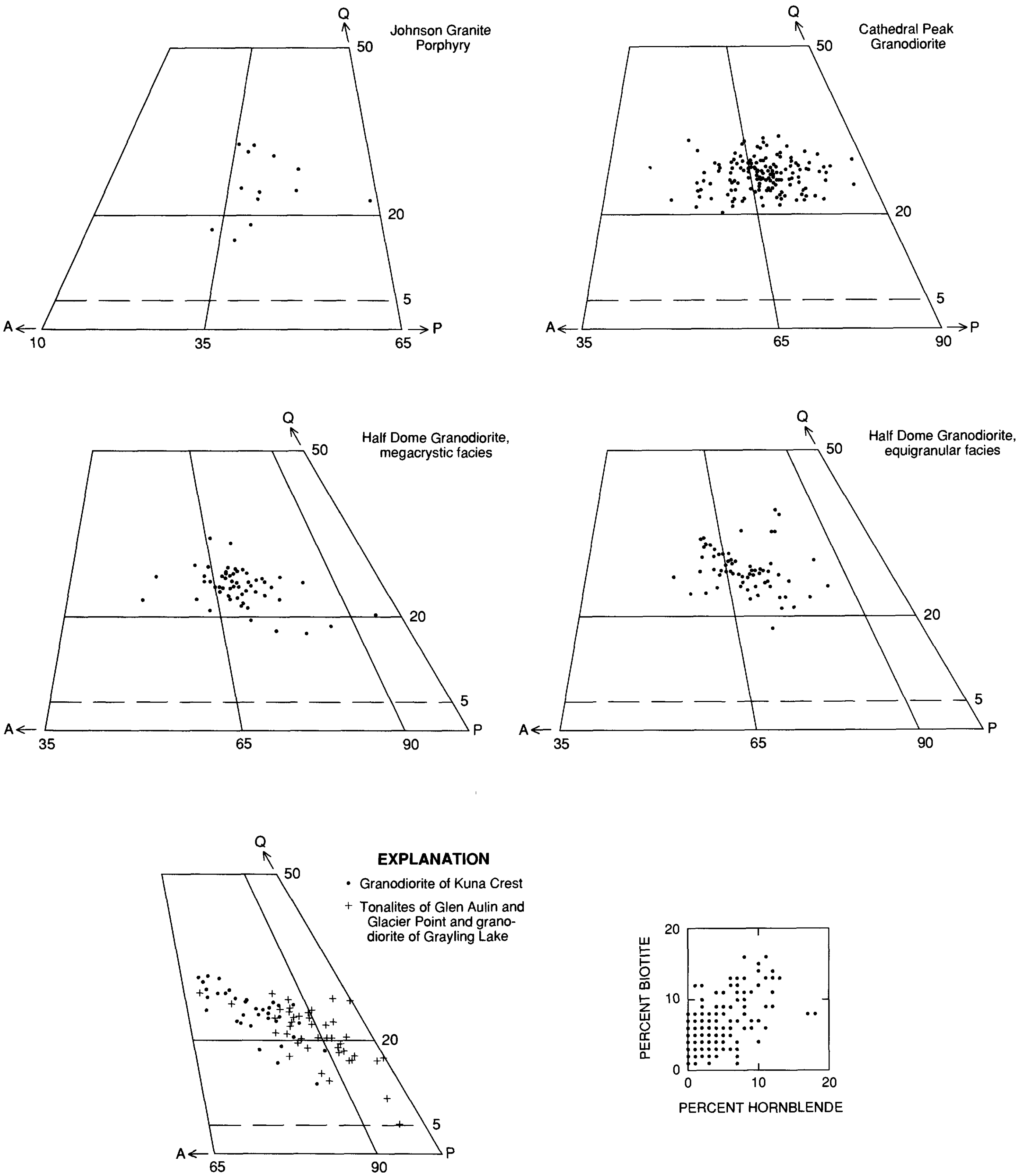


Figure 74.—Modes of the Tuolumne Intrusive Suite and plot showing amounts of hornblende and biotite in suite. See figure 6 for Q-A-P classification diagram.

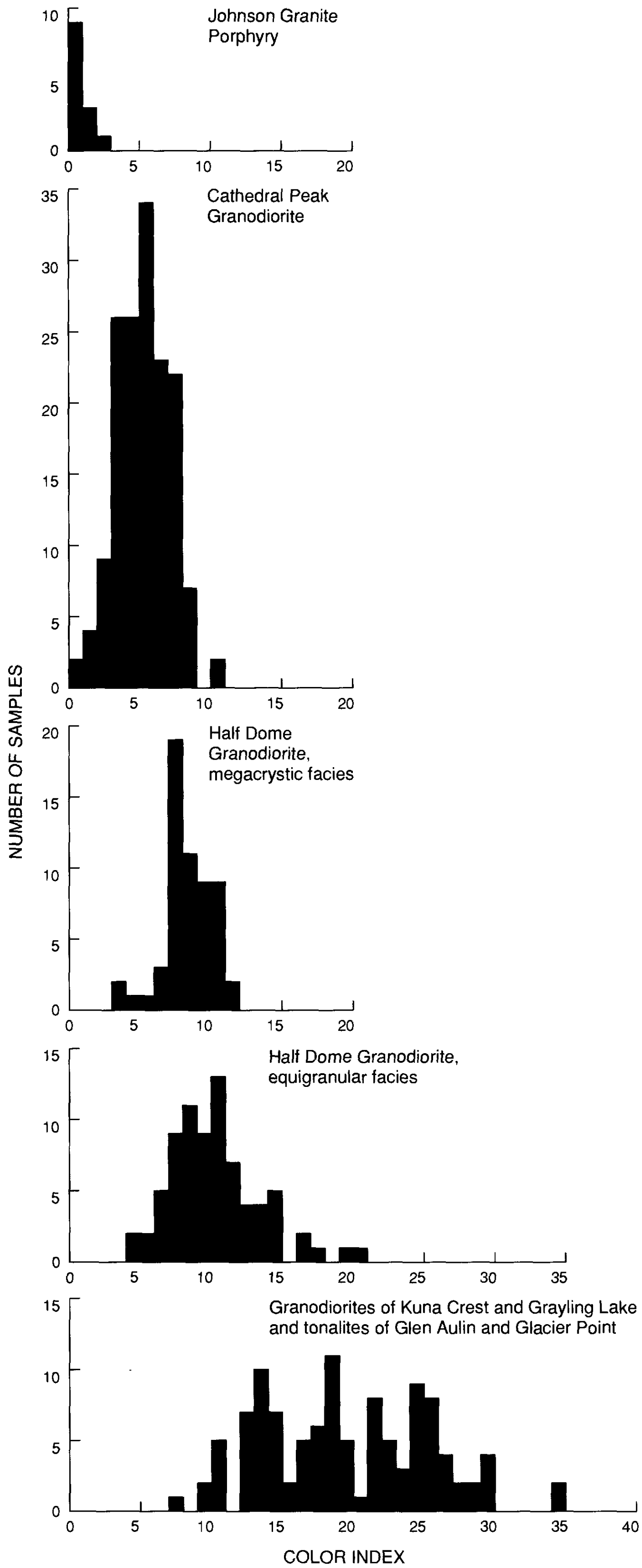


FIGURE 75.—Color indices of the Tuolumne Intrusive Suite.

mm across; some magnetite is present as equant crystals, and some titanite occurs as skeletal crystals. Apatite occurs as tiny (0.2–0.5-mm) prismatic crystals. Only small amounts of interstitial quartz and alkali feldspar are present in the outer part of the Glen Aulin, but both minerals increase in abundance inward and form nearly equant grains at the inner contact with the Half Dome Granodiorite.

These rocks in the margins of the Tuolumne Intrusive Suite intrude all the plutonic rocks with which they are in contact that do not belong to the suite, which indicates that the Tuolumne Intrusive Suite is the youngest in the region. They are intruded by the Half Dome and Cathedral Peak Granodiorites of the Tuolumne Intrusive Suite. The tonalite of Glen Aulin has a concordant U-Pb age of 88 Ma and the granodiorite of Kuna Crest a concordant U-Pb age of 91 Ma (Stern and others, 1981).

HALF DOME GRANODIORITE (Khd, Khdm)

The Half Dome Granodiorite is much coarser grained than the outer parts of the masses of tonalite and granodiorite just discussed but only slightly coarser grained near the contacts with those units. Although generally sharp, contacts with the tonalites and granodiorites are elusive because differences in composition and texture at the contacts are slight; careful examination of outcrops is generally required to locate contacts accurately. However, north of the Tioga Pass, the Half Dome Granodiorite sharply truncates foliation in the granodiorite of Kuna Crest.

The Half Dome Granodiorite includes an outer, equigranular facies and an inner, megacrystic facies. In most places the two facies are gradational through a few tens or hundreds of meters, but beginning 2 km north of Tenaya Lake and extending northward for a distance of about 6 km the contact between the two facies is sharp, and dikes of the megacrystic facies intrude the equigranular facies. Most modes of samples of both the equigranular and megacrystic facies plot in the granodiorite field on Q-A-P diagrams, but some plot in adjacent parts of the granite and quartz monzodiorite fields (fig. 74). The color index is generally within the range of 5 to 15 in the equigranular facies and 7 to 12 in the megacrystic facies (fig. 75).

Inward from the outer contact, both the size and perfection of crystal shapes continue to increase. In the inner part of the equigranular facies, discrete, well-formed hornblende prisms are as much as 5 mm across and 1.5 cm long, biotite books are as much as 1 cm across, and titanite wedges are as much as 3 mm across. Although hornblende and biotite crystals are larger and better formed inward across the equigranular facies of the Half Dome, both minerals decrease in abundance inward, hornblende decreasing from about 8 to 2 percent and biotite from 11 to 4 percent. Plagioclase remains constant at about 40 percent, but both quartz and alkali feldspar increase inward for about 1 km, then remain constant farther inward, both at about 25 percent.

Tabular alkali feldspar megacrysts about 1 cm thick and 2 to 3 cm long appear in the transition zone between the two facies and increase inward to the contact with the Cathedral Peak Granodiorite, where they make up about 10 percent of the rock. Accurate measurement of the abundance of alkali feldspar megacrysts is difficult, even when counts are made on large, glacially polished surfaces, because the texture of the rock is seriate. No sharp cutoff between megacryst and groundmass alkali feldspar exists. Although megacrysts increase in abundance inward, the total alkali feldspar content remains nearly constant.

The Half Dome Granodiorite intrudes the rocks in the margins of the Tuolumne Intrusive Suite and is intruded by the Cathedral Peak Granodiorite. The Half Dome has not been dated isotopically by the U-Pb method, but concordant U-Pb ages of 88 and 91 Ma for the granodiorite of Kuna Crest, 88 Ma for the tonalite of Glen Aulin, and of 86 Ma for the Cathedral Peak Granodiorite bracket its age closely (Stern and others, 1981).

CATHEDRAL PEAK GRANODIORITE (Kcp)

The Cathedral Peak Granodiorite attracts much attention because it contains blocky alkali feldspar megacrysts as much as 5 cm across in a medium-grained groundmass. Near the contact of the Cathedral Peak Granodiorite with the megacrystic facies of the Half Dome Granodiorite, megacrysts are crowded together in swarms, so that it is highly impractical to accurately determine their abundance. However, about 1 km inward from the contact the Cathedral Peak contains about 12 percent megacrysts. The size and abundance of megacrysts decrease inward toward the contact with the Johnson Granite Porphyry, where only a few scattered megacrysts half the size of those near the outer contact are present. Except for the inward decrease in abundance of megacrysts and an absence of hornblende in the innermost parts, the modal composition of the Cathedral Peak Granodiorite is fairly constant. Hornblende, biotite, and titanite form smaller crystals and are distinctly less abundant than in the Half Dome. Modal plots on a Q-A-P diagram range across the boundary between the granodiorite and granite fields (fig. 74). The color index generally ranges from 1 to 10 and peaks at 6 (fig. 75). Biotite is the chief mafic mineral, and the rock contains only traces of hornblende. The Cathedral Peak intrudes the Half Dome Granodiorite and is intruded by the Johnson Granite Porphyry. It has a concordant U-Pb age of 86 Ma (Stern and others, 1981).

JOHNSON GRANITE PORPHYRY (Kjp)

The Johnson Granite Porphyry forms a central body surrounded by a network of dikes in the core of the Tuolumne Intrusive Suite. Contacts with the adjacent Cathedral Peak Granodiorite are sharp, and dikes of the Johnson intrude the Cathedral Peak in an intricate pattern. The Johnson Granite Porphyry differs significantly from the other units of the Tuolumne Intrusive Suite in having a fine-grained groundmass, a porphyritic texture, miarolitic cavities, and a color index of about 1 (fig. 75). Sporadically distributed alkali feldspar megacrysts, some with a border of medium-grained rock identical to the groundmass of the contiguous Cathedral Peak, and scattered angular fragments of plagioclase and quartz approximately 2 to 4 mm across are set in an extremely fine-grained groundmass composed of quartz, alkali feldspar, and plagioclase grains less than 1 mm across. The fine-grained groundmass indicates that the magma was quenched, the angular plagioclase and quartz fragments indicate comminution, and the miarolitic cavities indicate the presence of a separate fluid phase. The most likely cause of quenching is release of pressure, which would diminish the solubility of volatiles in the melt and raise the crystallization temperature. If an eruption occurred, it probably was caused by the gradual increase of volatiles in the melt phase of the magma as a result of the crystallization of anhydrous and nearly anhydrous minerals until the volatiles content exceeded the amount soluble in the melt. The appearance of this separate fluid phase would quickly increase the volume of the system to a size beyond which the wall and roof rocks could no longer adjust without fracturing (Burnham, 1972).

The Johnson contains more modal alkali feldspar than the contiguous Cathedral Peak Granodiorite (fig. 84). Chemically it contains more K_2O and less CaO, Fe_2O_3, FeO, MgO, TiO_2, P_2O_5, and MnO; minor elements include small amounts of V, Zn, and Sr. These chemical characteristics extend into adjacent parts of the Cathedral Peak Granodiorite, showing that it too was affected by circulating volatiles.

POSSIBLE OLDER UNITS OF THE TUOLUMNE INTRUSIVE SUITE

The affiliations of the granodiorite of Yosemite Creek and the Sentinel Granodiorite are uncertain. Although no isotopic ratios have been determined, these Late Cretaceous granodiorites appear to be of about the same age as the adjacent Tuolumne Intrusive Suite and may actually be early units of that suite. Because of uncertainties concerning their ages and contact relations with units of the Tuolumne Intrusive Suite, they are herein left unassigned.

GRANODIORITE OF YOSEMITE CREEK (Kyc)

The granodiorite of Yosemite Creek, called the Yosemite Creek Granodiorite of Rose by Kistler (1973) and Bateman, Kistler, and others (1983), is a dark-gray medium- to coarse-grained rock of highly variable composition. On a Q-A-P diagram, modes are widely scattered across the granite, granodiorite, tonalite, and quartz diorite fields, and the color index shows a correspondingly wide range (fig. 76). Locally, the rock contains plagioclase phenocrysts. It intrudes the El Capitan Granite and the granite of Rancheria Mountain of the intrusive suite of Yosemite Valley and is intruded by the Sentinel Granodiorite.

SENTINEL GRANODIORITE (Kse)

The Sentinel Granodiorite as originally mapped by F.C. Calkins (Calkins, 1930; see also Calkins and others, 1985) included the tonalite of Glacier Point of this report. However, while mapping the geology of the Yosemite 15-minute quadrangle, D.L. Peck (written commun., 1982) mapped a contact within the original Sentinel Granodiorite and separated the tonalite of Glacier Point from the unit. Sentinel Rock, the type locality of the Sentinel Granodiorite, lies within the unit as restricted (pl. 1) Typical Sentinel resembles the Half Dome Granodiorite in being equigranular and containing well-formed crystals of hornblende and biotite and abundant wedge-shaped crystals of sphene. On a Q-A-P diagram, modes plot across the granite, granodiorite, and tonalite fields but are most abundant in the granodiorite field (fig. 76). The color index ranges from 4 to about 26 and peaks at 15 (fig. 76). The Sentinel Granodiorite intrudes the granodiorite of Yosemite Creek and rocks of the intrusive suites of Yosemite Valley and Buena Vista Crest. Its age relative to the outermost rocks of the Tuolumne Intrusive Suite has not been established.

JOHN MUIR INTRUSIVE SUITE

The John Muir Intrusive Suite, called the John Muir sequence by Bateman and Dodge (1970), is herein formally named for exposures along the John Muir trail, which extends southward from Yosemite Valley along the crest of the Sierra Nevada. The type area of the suite is the area between Rock Creek on the east and

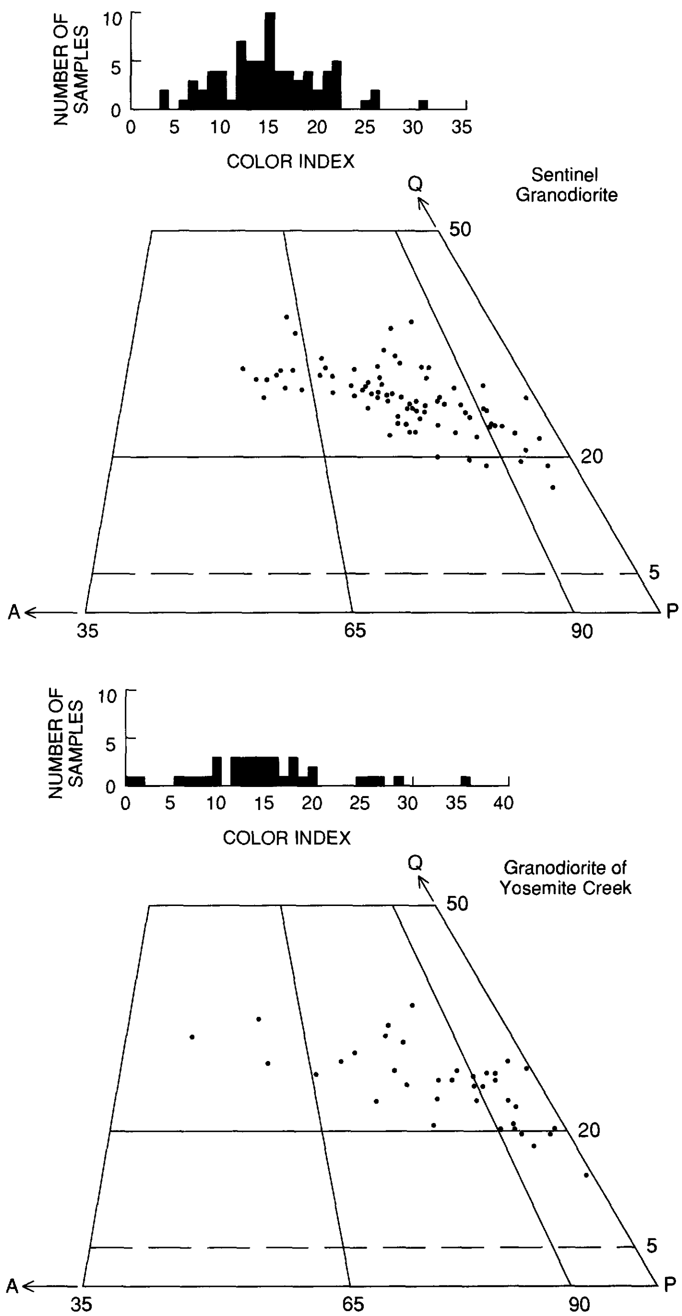


FIGURE 76.—Modes and color indices of the Sentinel Granodiorite and granodiorite of Yosemite Creek. See figure 6 for Q-A-P classification diagram.

Potter Pass on the west. The units assigned to this suite are—from oldest to youngest—the Lamarck, Mount Givens, Lake Edison, and Round Valley Peak Granodiorites, the Mono Creek Granite, and the Evolution Basin Alaskite. The granite of Rock Creek Lake, a small body of granite that intrudes the north end of the Lake Edison Granodiorite, is also included. Stern and others (1981) assigned most of the units now included in the suite to their Kaiser, Powell, and Mono Pass granitoid sequences, but these divisions have proved difficult to justify and are not used in this report.

The order of emplacement within the suite is from mafic to felsic. With the exception of the Mount Givens Granodiorite, all the larger plutons have long, narrow shapes and strike N. 45°–50° W., suggesting that the magmas from which they solidified were unusually mobile and that they may have been emplaced during a time of regional extension. The Mount Givens Granodiorite, the largest intrusion included in the suite, is partly separated from the others by the Mount Goddard roof pendant. It is less elongate, and its long axis trends a little more toward the north than the others. All the U-Pb isotopic ages for units of the suite are about 90 Ma (Stern and others, 1981).

LAMARCK GRANODIORITE (Klk)

The Lamarck Granodiorite, the oldest unit of the John Muir Intrusive Suite, forms a thin lenticular pluton more than 60 km long and 10 km wide near the middle. Generally, the rock is homogeneous in composition and texture, but the core of the thickened middle and the southeast end, which is almost detached from the main mass, are leucocratic and weakly megacrystic. Most modes plot in the granodiorite field on a Q-A-P diagram, but those of samples from the core and the south end plot in the granite field (fig. 77). The color index ranges from 4 to 24 and averages about 16, but the average for the leucocratic southern part is only about 6 (fig. 79).

Typical rock is medium grained and seriate or, less commonly, equigranular. The average grain size is 3 to 4 mm, but grains of alkali feldspar commonly are as much as 1 cm across. The texture is typically hypidiomorphic-granular, indicating that the rock has not been significantly deformed since it solidified. Biotite and hornblende are generally evenly distributed and occur both in clusters and in discrete euhedral to subhedral crystals. Plagioclase commonly is zoned from about An_{45} to about An_{30} but is more sodic in the felsic south end of the pluton where the zoning generally ranges from An_{27} to An_{17}.

Commonly, the rock has a conspicuous planar foliation that is most clearly expressed by abundant lenticular mafic inclusions but which also is expressed by planar orientation of biotite and hornblende. The foliation is most apparent in the margins of the intrusion and is less obvious inward as the rock becomes more leucocratic and the number of mafic inclusions diminishes. Most foliations are steep and subparallel to contacts with other rocks, but in the southern part of the thickened middle of the pluton they dip gently or are horizontal and define a northwest-trending antiform and synform. These foliations presumably reflect the configuration of the upper contact with roof rocks, which must have overlain this part of the Lamarck not far above the exposed level.

The Lamarck Granodiorite intrudes the unassigned Inconsolable Quartz Monzodiorite and the Tinemaha Granodiorite of the Palisade Crest Intrusive Suite; it is intruded by the Mount Givens and Lake Edison Granodiorites and the Evolution Basin Alaskite of the John Muir Intrusive Suite, and the unassigned granodiorite of Cartridge Pass. It has a discordant U-Pb age of 90 Ma (Stern and others, 1981) and K-Ar ages on hornblende of 90 and 86 Ma and on biotite of 85 and 79 Ma (Kistler and others, 1965; Evernden and Kistler, 1970). None of these ages are definitive; nevertheless the age of the Lamarck is bracketed by the ages of other intrusions with which it is in contact at about 90 Ma.

MOUNT GIVENS GRANODIORITE (Kmg, Kmgm)

The Mount Givens Granodiorite, one of the larger plutons in the Sierra Nevada, is exposed over an area of about 1,400 km^2. The pluton is elongate in a northwestern direction, having a length of 80 km and a width that averages 17 km across the southern two-thirds of the body but bulges to almost twice that distance at the north end. The modes of most samples of the Mount Givens plot on a Q-A-P diagram in the granodiorite and granite fields, but a few plot in the tonalite field (fig. 78). The Mount Givens consists of two facies, an equigranular granodiorite facies and a megacrystic facies that is composed of about equal amounts of granite and granodiorite. The equigranular facies forms the central and southern parts of the intrusion, and the megacrystic facies forms most of the bulbous head at the north end. However, equigranular to weakly megacrystic rock also forms several elliptical domains in the central and southern part of the pluton. The color index ranges from about 5 to 24 in the equigranular granodiorite and from about 1 to 12 in the megacrystic granite and granodiorite (fig. 79).

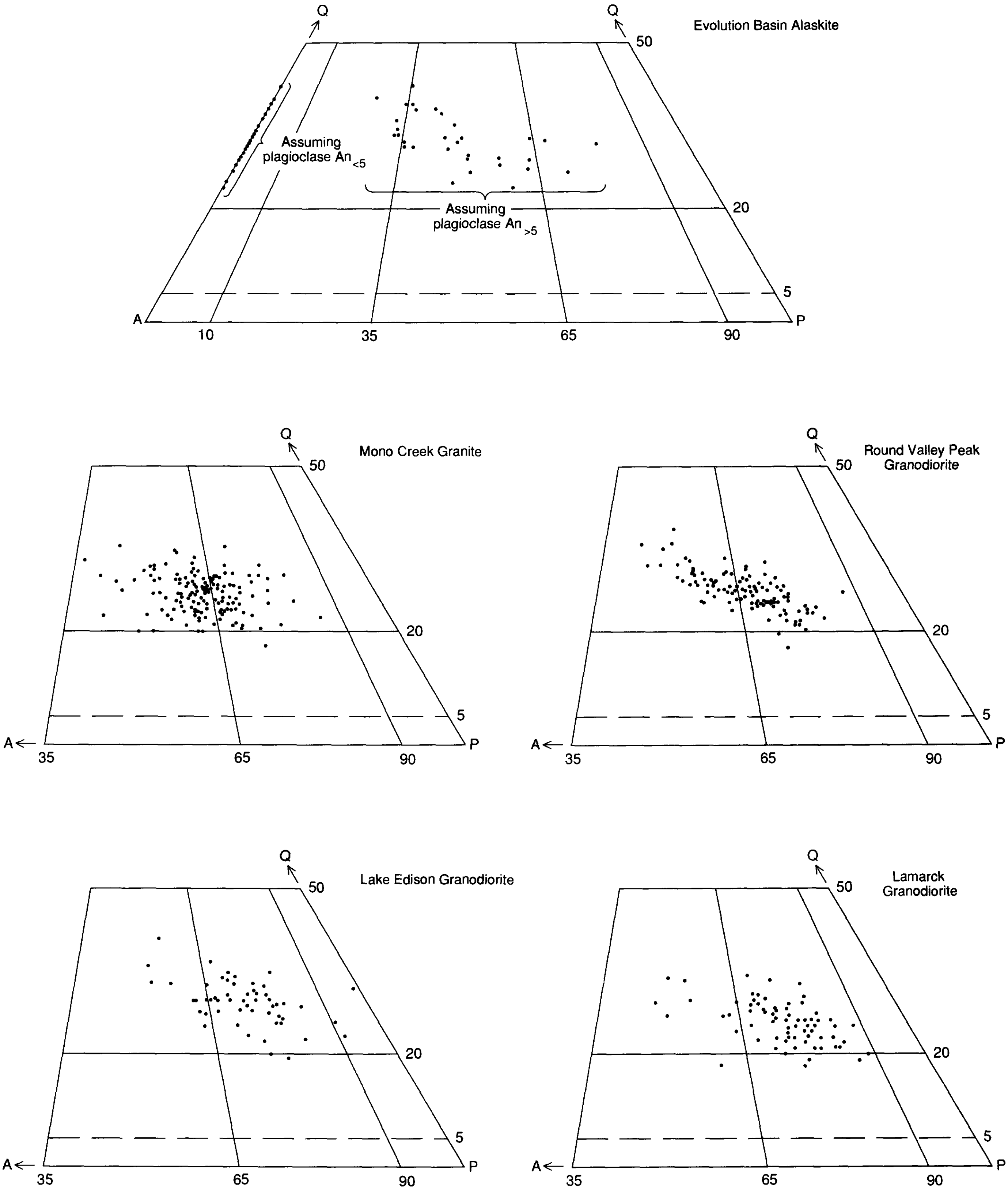


FIGURE 77.—Modes of the John Muir Intrusive Suite east of the Mount Givens Granodiorite. See figure 6 for Q-A-P classification diagram.

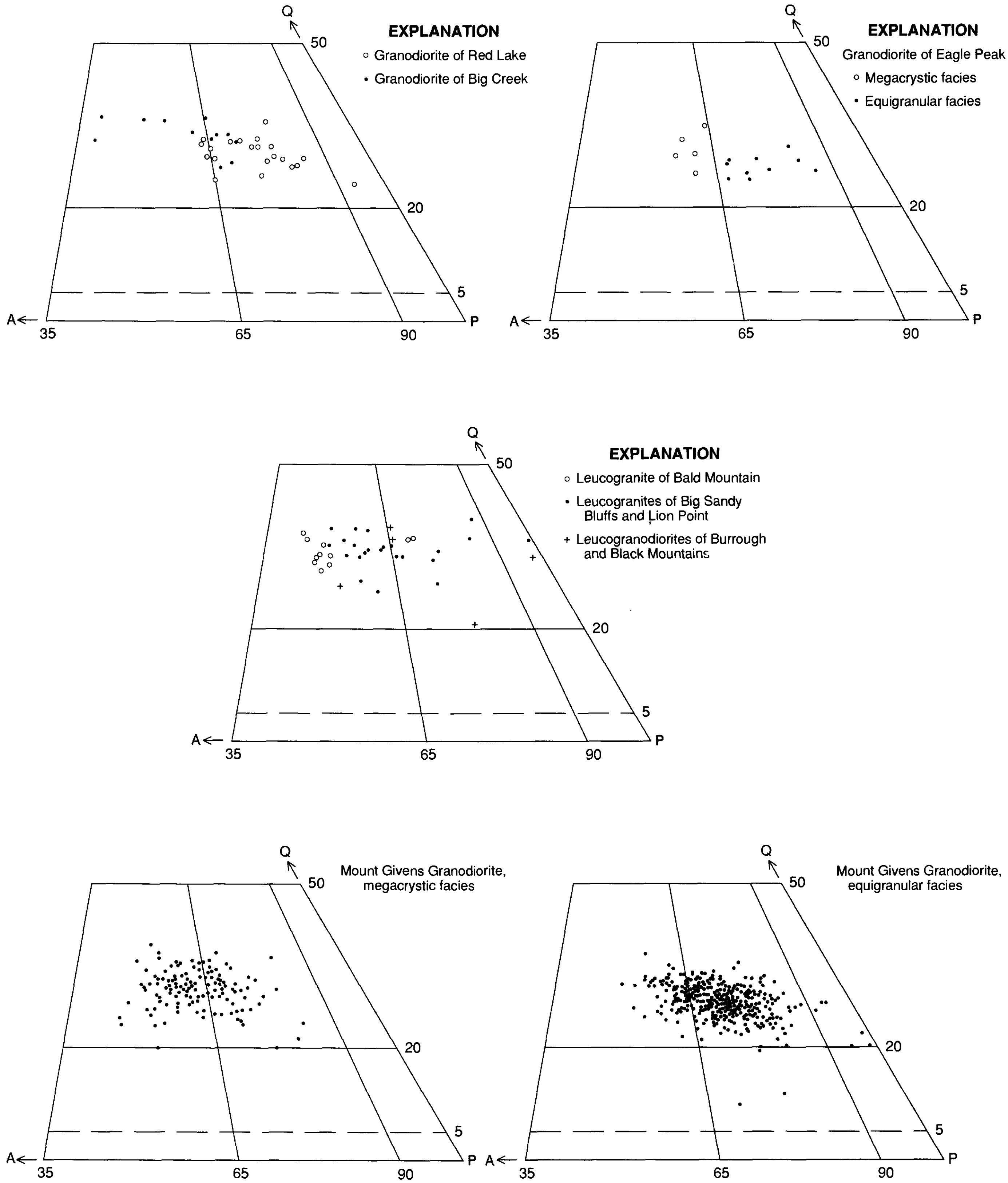


FIGURE 78.—Modes of the Mount Givens Granodiorite and small intrusions of the John Muir Intrusive Suite west of the Mount Givens Granodiorite. See figure 6 for Q-A-P classification diagram.

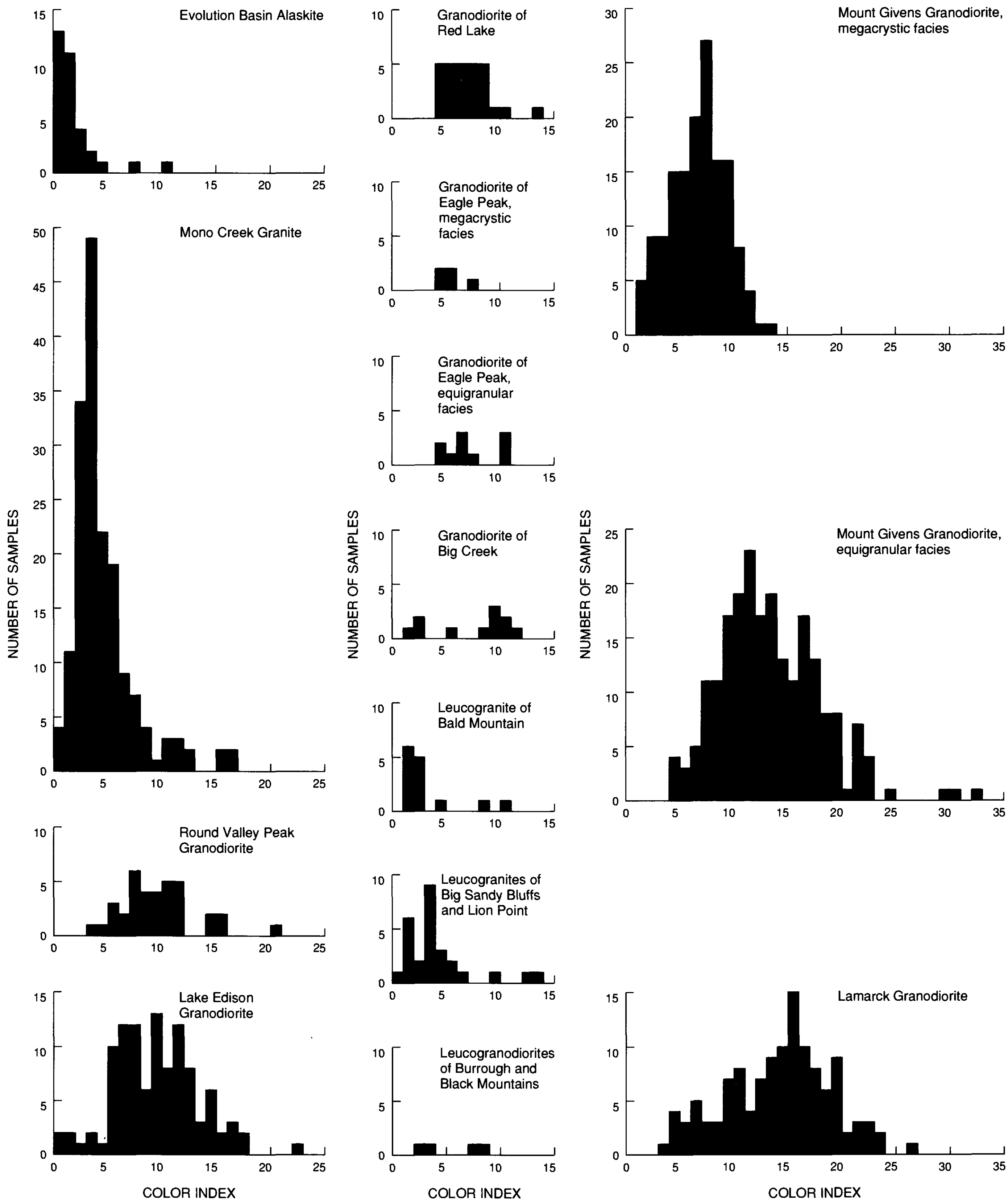


FIGURE 79.—Color indices of the John Muir Intrusive Suite.

Variations in modal composition and specific gravity within the pluton are shown in figure 40. The more mafic rocks, mostly granodiorite, are medium grained, equigranular, and generally have a strong primary foliation that is expressed by the preferred orientation of minerals and by lens-shaped mafic inclusions; the megacrystic rocks are weakly foliated or unfoliated. The different compositional and textural variants generally grade into one another. Samples collected along a traverse from the margin to the core of the bulbous head of the pluton (fig. 39) and studied in detail by Bateman and Nokleberg (1978) show that equigranular biotite-hornblende tonalite in the margin grades inward through equigranular hornblende-biotite tonalite and granodiorite to megacrystic biotite granodiorite and granite. This inward gradation is interrupted at a horseshoe-shaped internal intrusive contact where the megacrystic granite and granodiorite are intruded by fine-grained and equigranular granodiorite that grades inward to megacrystic granite. Details of the compositional zoning in the bulbous head are given in the section "Mount Givens Granodiorite" under "Patterns Within Intrusive Units."

Hornblende and biotite in the hornblende-bearing rocks are generally subhedral or anhedral, and alkali feldspar megacrysts in the megacrystic rocks are subhedral. Some of the biotite and hornblende is present in discrete grains, but much of it occurs in clots with titanite, magnetite, and apatite. All gradations between tiny clots consisting of only a few grains of mafic and accessory minerals to mafic inclusions as much as 1 m in longest dimension can be seen.

The distribution of rock compositions suggests that the pluton solidified inward from the margins toward the granite domains, which are composed of lower temperature mineral assemblages. The most extensive areas of granite are in the bulbous head, but several smaller areas are present in the southern part of the pluton. The rounded shape of the bulbous head probably resulted from the lower density and higher viscosity of the siliceous magma; this felsic magma, whose viscosity approached that of the wallrocks, ballooned to form a near-equant magma cell (Ramberg, 1970). A new surge of relatively femic magma then intruded the bulbous head and fractionated inward to form granitic rock.

The Mount Givens Granodiorite intrudes all the granitoids along its west and north borders except the granodiorites of Red Lake and Eagle Peak, which are considered members of the John Muir Intrusive Suite. On the east it is intruded by the Lake Edison Granodiorite of the John Muir Intrusive Suite. The Mount Givens is in contact with the Lamarck Granodiorite only along a short span that is mostly covered with glacial deposits; nevertheless, Lockwood (1975) found evidence that the Mount Givens intrudes the Lamarck. U-Pb ages for two samples of the Mount Givens are 93 and 88 Ma (Stern and others, 1981). The 93-Ma age is slightly discordant and the 88-Ma age is concordant. K-Ar ages are 89 and 88 Ma on hornblende and 89, 87, and 84 Ma on biotite (Kistler and others, 1965). These determinations suggest that the true age is about 90 Ma (early Late Cretaceous), 12 m.y. younger than the optimum average age, about 102 Ma (late Early Cretaceous), for the Shaver Intrusive Suite, just to the west.

LAKE EDISON GRANODIORITE (Kle)

The Lake Edison Granodiorite is herein formally named for exposures at its type locality, which is along the northeast side of Lake Thomas Edison. This unit has previously been informally referred to as the granodiorite of Lake Edison (Lockwood and Lydon, 1975; Stern and others, 1981). The Lake Edison Granodiorite forms a pluton that is more than 50 km long; it is narrow in the middle and swollen at both ends. A narrow tongue, previously designated the Morgan Creek mass (Bateman, 1965b), which extends from the southern bulge northward along the west side of the Pine Creek septum, was cut off from the main mass by intrusion of the younger Mono Creek Granite. The Morgan Creek mass is of particular interest because it is genetically related to the important skarn-type tungsten deposits of the Pine Creek septum.

Typical rock of the Lake Edison Granodiorite is fine- to medium-grained, equigranular hornblende-biotite granodiorite, distinctly finer grained than the Lamarck and Mount Givens Granodiorites. The rock exhibits good foliation that parallels adjacent contacts and is vertical or steeply dipping. Titanite is generally abundant. The color index generally ranges from 1 to 18, somewhat lower than the range in the Lamarck, and about the same as in the Mount Givens (fig. 79). The bulge at the north end of the pluton is concentrically zoned, and the color index decreases from more than 14 in the margins to less than 8 in the core. The bulge at the south end also is zoned, but laterally rather than concentrically; hornblende is absent in the bulge, and the color index decreases somewhat irregularly toward the east from more than 10 to less than 6. This bulge was originally called the Basin Mountain mass of the Tungsten Hills Quartz

Monzonite (Bateman, 1965b) (herein called the Tungsten Hills Granite), but later mapping and isotopic ages show that the bulge is part of the Lake Edison Granodiorite and much younger than the Triassic Tungsten Hills Granite (Lockwood and Lydon, 1975; Stern and others, 1981).

The bulges at the two ends of the Lake Edison Granodiorite reflect the presence of leucocratic granite, which was emplaced as relatively low-temperature, viscous magma that reduced the viscosity contrast between the magma and the wallrocks and favored a more equidimensional form. The bulge at the north end of the pluton crowded the Mount Givens Granodiorite westward and deflected its foliation (pl. 1).

The Lake Edison Granodiorite intrudes the Mount Givens and Lamarck Granodiorites, also of the John Muir Intrusive Suite, and the unassigned leucogranite of Graveyard Peak; it is intruded by the Round Valley Peak Granodiorite and the Mono Creek Granite of the John Muir Intrusive Suite. It has a concordant U-Pb age of 90 Ma (Stern and others, 1981) and two K-Ar ages on biotite of 82 and 77 Ma and one on hornblende of 85 Ma (Kistler and others, 1965; Evernden and Kistler, 1970). Apparently it is only slightly younger than the Lamarck and Mount Givens Granodiorites.

GRANITE OF ROCK CREEK LAKE (Krc)

The granite of Rock Creek Lake forms a small oval mass of light-colored, medium- to coarse-grained biotite granite in the core of the zoned northern bulge of the Lake Edison Granodiorite. It appears to be a late leucocratic fractionate of the Lake Edison Granodiorite which moved slightly upward and intruded the Lake Edison.

ROUND VALLEY PEAK GRANODIORITE (Krv)

The Round Valley Peak Granodiorite forms three plutons that lie east of the Lake Edison Granodiorite and are generally separated from it by the younger Mono Creek Granite and by older granitoids and country rocks; the Round Valley Peak and Lake Edison Granodiorites are in contact only at the south end of the largest and southernmost pluton, the Round Valley Peak pluton. This pluton is an arcuate body that encloses a lobe of the younger Mono Creek Granite on the west. Before intrusion of the Mono Creek, this pluton probably was elongate parallel to the long axes of the Lamarck and Lake Edison Granodiorites. The other two plutons of the Round Valley Peak Granodiorite lie farther northwest and intrude the Mount Morrison roof pendant.

Typical Round Valley Peak Granodiorite is notably equigranular and medium grained. Though generally finer grained, it closely resembles the Lamarck Granodiorite and the equigranular facies of the Half Dome Granodiorite of the Tuolumne Intrusive Suite. Biotite and hornblende are evenly distributed in discrete euhedral to subhedral crystals, and plagioclase commonly is in equant to slightly elongate grains. Some plagioclase grains contain mottled cores, in the compositional range An_{49-45}, which are surrounded by progressively zoned plagioclase that ranges in composition from An_{38} in the inner part to An_{21} in the margins. On a Q-A-P diagram the plot of modes is centered on the granodiorite field but extends into both the tonalite and granite fields (fig. 77). The color index ranges from 5 to 15 and averages about 9 (fig. 79).

The Round Valley Peak pluton is strongly foliated and compositionally zoned. In the eastern margin, adjacent to older rocks, abundant strongly flattened mafic inclusions that appear elongate in outcrop give the rock a strong foliation. The rock in the margins of the pluton is relatively fine grained, the average grain size being only about 2 mm. With distance from the older rocks, toward the younger Mono Creek Granite, the rock is progressively more leucocratic, coarser grained, and contains fewer and more rounded inclusions. In the smaller masses of the Round Valley Peak Granodiorite that are enclosed in the Mount Morrison roof pendant, the rock is similar but finer grained, the average grain size ranging from 1 to 2 mm, and contains few mafic inclusions (Rinehart and Ross, 1964).

The Round Valley Peak Granodiorite intrudes the Lake Edison Granodiorite, also of the John Muir Intrusive Suite, and the Triassic Wheeler Crest Granodiorite of the Scheelite Intrusive Suite and is intruded by the Mono Creek Granite. It has a discordant U-Pb age of 89 Ma (Stern and others, 1981) and K-Ar ages on biotite of 89 Ma and on hornblende of 84 Ma (Kistler and others, 1965).

MONO CREEK GRANITE (Kmo)

The Mono Creek Granite, previously called the quartz monzonite of Mono Recesses by Lockwood and Lydon (1975), is herein formally named for exposures in the recesses along the tributaries to Mono Creek, which are designated the type locality. Most of the Mono Creek Granite forms a single pluton, but a small pluton called the McGee Creek mass (Bateman, 1965b) crops out farther south along the range front, east of Mount Humphreys; another small mass, called the quartz monzonite of Turret Peak (Lockwood and Lydon, 1975), lies southwest of the south end of the main mass. The Mono Creek Granite resembles the Cathedral Peak Granodiorite of the Tuolumne Intrusive Suite but generally is a little finer grained.

Typical Mono Creek Granite is megacrystic. Blocky alkali feldspar megacrysts, averaging 1.4×2.5 cm as seen in outcrops, are set in a medium-grained groundmass. The color index of most samples is in the range of 2 to 16 (fig. 79). On a Q-A-P diagram, modes plot in both the granite and granodiorite fields (fig. 77). A weak foliation parallel to the long axis of the pluton is indicated chiefly by the preferred orientation of the megacrysts. Mafic inclusions are scarce, but inclusions of wallrocks are present locally, adjacent to metamorphic and mafic igneous rocks.

Compositional zoning is not readily apparent in the field or in modal plots (Lockwood, 1975) but is indicated by small but systematic changes in specific gravity (fig. 80). The specific gravity of rocks in the margins is greater than 2.65, whereas in the core area it ranges from 2.63 to 2.64. Zonation is also shown by the abundance of alkali feldspar megacrysts (fig. 8). In small outcrops the megacrysts appear to be distributed irregularly, but by determining the average percentages in large outcrops, Lockwood (1975) demonstrated that in the central and southern parts of the main pluton the proportion of megacrysts decreases systematically inward from about 10 percent in the margins to as little as 3 percent in the core. An exception to this pattern is a protrusion of equigranular leucogranite from the east side of the pluton. A northwest-trending zone of high specific-gravity values and abundant alkali feldspar megacrysts crosses the west side of this protrusion (figs. 8, 80). As occurred in the bulges at the two ends of the Lake Edison Granodiorite, the viscous leucocratic magma in this protrusion has bent the surrounding Round Valley Peak Granodiorite and older metamorphic and plutonic rocks eastward many kilometers. The abundance of megacrysts is loosely related to the total abundance of alkali feldspar, which ranges irregularly from 20 to 35 percent, and shows no relation to altitude (fig. 8).

In the vicinity of the Pine Creek septum, the pluton is bordered by a swarm of pegmatite and aplite dikes that dip gently into the pluton. The dikes extend as much as 2 km from the pluton but thin with distance; those that reach the Pine Creek septum pinch out in marble that lies along the west side of the pendant. Matching irregularities in the wallrocks of the dikes clearly indicate that the dikes were emplaced by dilation of feather joints that were produced by upward movement and expansion of the viscous Mono Creek magma (Bateman, 1965b). Kerr (1946) suggested that these dikes carried tungsten from the last-crystallizing core of the Mono Creek Granite to form the tungsten-rich skarn deposits in the west side of the Pine Creek roof septum, but this suggestion is refuted by the fact that steeply dipping to vertical skarn ore bodies are cut and physically displaced by the gently dipping dikes.

The Mono Creek Granite intrudes the Round Valley Peak, Lake Edison, and Mount Givens Granodiorites, also of the John Muir Intrusive Suite. It also intrudes the unassigned granite of Chickenfoot Lake, the leucogranite of Graveyard Peak, and the granodiorites of King and Fish Creeks and of Margaret Lake. It has a discordant U-Pb age of 88 Ma and a concordant U-Pb age of 76 Ma (Stern and others, 1981) and K-Ar ages on biotite of 82 to 79 Ma and on hornblende of 79 Ma (Evernden and Kistler, 1970).

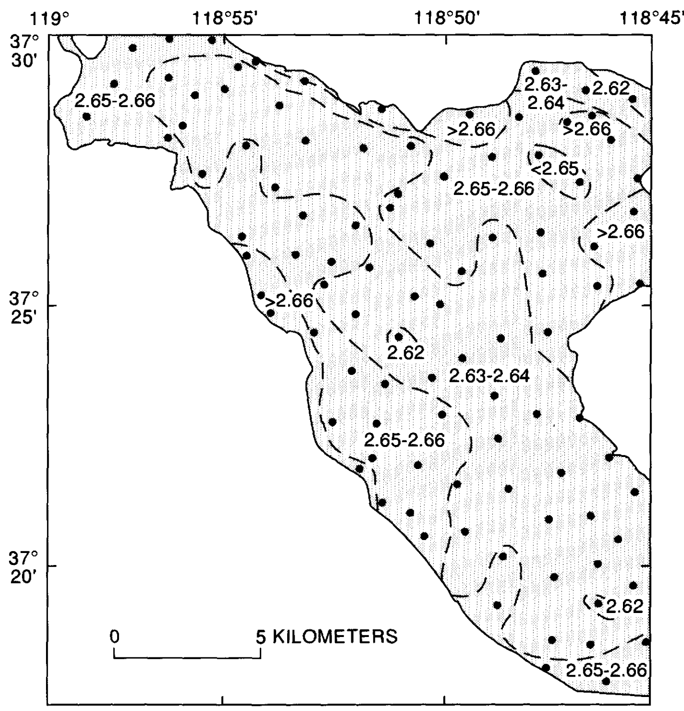


FIGURE 80.—Variations in specific gravity within central and southern parts of the Mono Creek Granite. Dots show sample localities; dashed lines separate areas of different specific gravity. Modified from Lockwood (1975).

EVOLUTION BASIN ALASKITE (Kev)

The Evolution Basin Alaskite, called the alaskite of Evolution Basin and LeConte Canyon by Bateman and Moore (1965) and the leucogranite of Evolution Basin by Stern and others (1981), is herein formally named for exposures in Evolution Basin, which is designated the type locality. It lies between the south half of the Lamarck Granodiorite on the east and the Goddard septum on the west. Like the Lamarck, it forms an elongate body but is only about 20 km long and averages about 5 km in width. A spur of metamorphic rock from the Goddard septum very nearly separates it into two plutons, a northern one in Evolution Basin and a southern one in LeConte Canyon. Contacts dip outward at 60° to 70° in the northern part, and a few observations suggest this is also true of the southern part. These outward-dipping contacts indicate that the rocks at the exposed level are in the upper part of the alaskite and that the roof was not far above.

The rock in both parts is extremely felsic, and the average color index is about 2 (fig. 79). Two chemically analyzed samples from the northern part of the alaskite have normative plagioclase compositions of $An_{1.1}$ and $An_{1.7}$ (Bateman, Dodge, and Bruggman, 1984); such compositions indicate that this rock is a hypersolvus granite and was emplaced at a shallow depth. In hand specimen the rock appears to be medium grained and equigranular but in thin section can be seen to be fine grained. Grains have ragged boundaries and in places are separated by very fine grained granoblastic intergrowths. Quartz generally exhibits strong undulatory extinction, and the apparently large quartz grains seen in hand specimens actually are composed of differently oriented segments that meet along crenulate boundaries. The sparse mafic minerals are chiefly biotite, titanite, and opaque minerals.

Although the two parts of the alaskite appear very similar, the modal plagioclase content is generally less than 35 percent in the northern part and more than 35 percent in the southern part. No samples from the southern part have been analyzed chemically, but the higher plagioclase content suggests earlier crystallization at higher temperatures.

The alaskite intrudes the Lamarck Granodiorite and a body of sheared, fine-grained hornblende-quartz syenite that has a U-Pb age of 157 Ma (Chen and Moore, 1982). It also truncates mafic dikes in the quartz syenite, which presumably belong to the ~148-Ma Independence dike swarm. The alaskite is intruded by the unassigned granodiorite of Cartridge Pass and by dikelike bodies of fine-grained aplitic granite. K-Ar biotite ages of 85, 80, and 79 Ma were reported by Evernden and Kistler (1970).

SMALL INTRUSIONS WEST OF THE MOUNT GIVENS GRANODIORITE

Three groups of small intrusions that lie west of the Mount Givens Granodiorite are tentatively assigned to the John Muir Intrusive Suite. One group consists of the granodiorites of Red Lake, Eagle Peak, and Big Creek; the second group includes only the leucogranite of Bald Mountain; and the third group consists of the leucogranites of Big Sandy Bluffs and Lion Point and the leucogranodiorites of Burrough and Black Mountains (fig. 81).

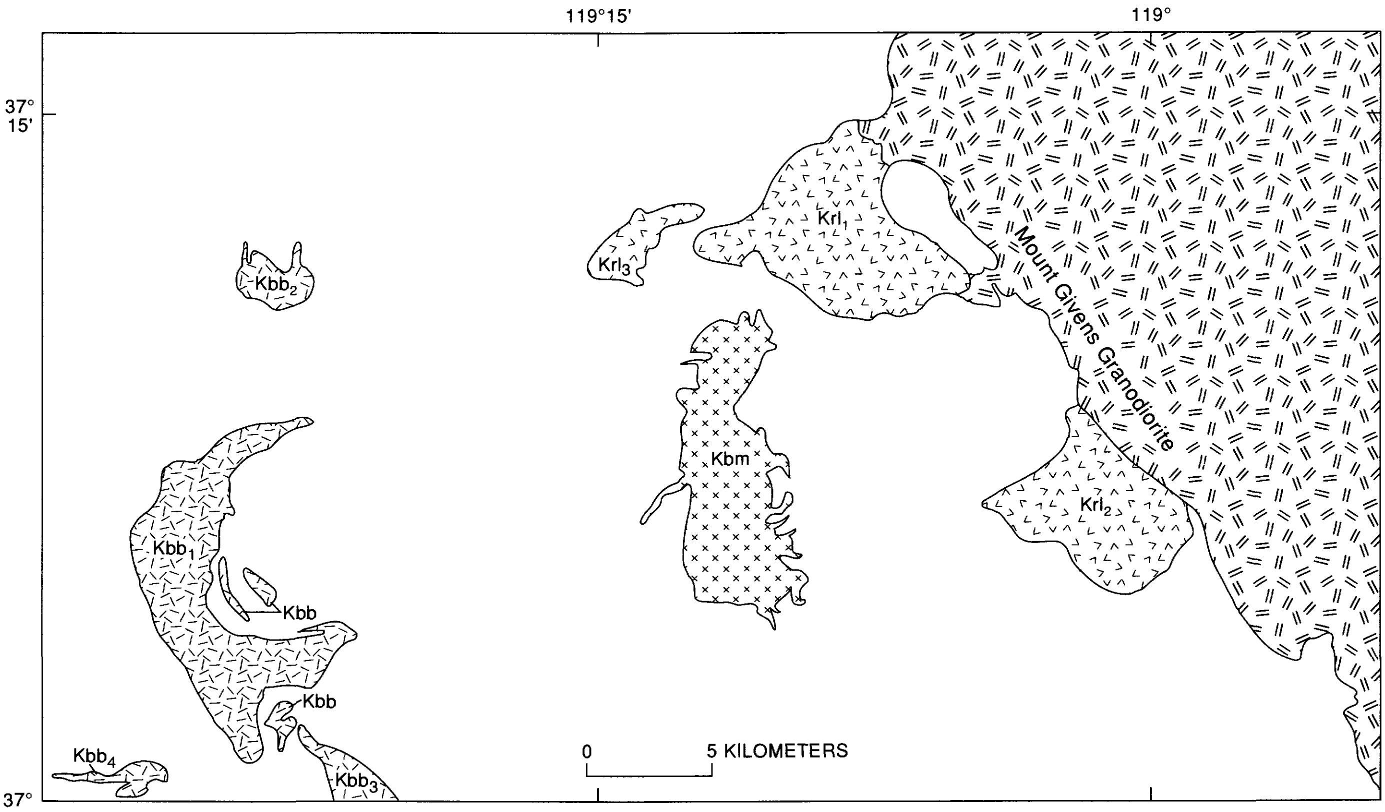
GRANODIORITES OF RED LAKE, EAGLE PEAK, AND BIG CREEK (Krl)

The granodiorites of Red Lake and Eagle Peak were the subject of detailed studies by Noyes and others (Noyes, Frey, and Wones, 1983; Noyes, Wones, and Frey, 1983) in which the source and mode of emplacement of the two granodiorites were compared. The granodiorite of Red Lake forms a triangular pluton that lies along the west side of the Mount Givens Granodiorite and intrudes the Dinkey Creek Granodiorite and the granite of Dinkey Dome. Except in a lobe at the west end, where the rock is hydrothermally altered, the granodiorite of Red Lake is uniformly light gray, fine grained, and equigranular; the color index ranges from 5 to 9 (fig. 79). Mafic minerals include both hornblende and biotite, and opaque minerals and titanite are the most common accessory minerals. The rock appears uniform throughout the pluton. Also, the plot of modes on a Q-A-P diagram shows only moderate spread, and most modes plot within the granodiorite field (fig. 78). The western and central parts of the pluton are massive, but the northern and southeastern parts are foliated, generally parallel to the closest external contact.

The granodiorite of Red Lake is in contact with the Mount Givens Granodiorite at two places along its northeast side. Along a more northern contact, the Red Lake clearly intrudes the Mount Givens but appears gradational along the southern contact. The granodiorite of Red Lake is also in contact with and intrudes the Dinkey Creek Granodiorite of the

Shaver Intrusive Suite. The contact between the granodiorite of Red Lake and the Dinkey Creek is well exposed in roadcuts along State Highway 163 just south of the Big Creek crossing. Along the northwest side of the highway, flat-lying peripheral dikes of the granodiorite of Red Lake can be seen to intrude the Dinkey Creek Granodiorite. Two K-Ar biotite ages of 90 and 84 Ma were determined for the granodiorite of Red Lake (Bateman and Wones, 1972b).

The granodiorite of Eagle Peak forms a pluton that lies southeast of the granodiorite of Red Lake and, like the Red Lake, lies along the southwest side of the Mount Givens Granodiorite. Most of the Eagle Peak is fine-grained equigranular granodiorite; the pluton is concentrically zoned outward from a core of megacrystic granite. Both hornblende and biotite are present. The color index ranges from 5 to 10 and generally is a little higher in the margins than in the megacrystic core (fig. 79). The contact between the megacrystic core and equigranular rock is completely gradational, but the gradation takes place within a very narrow transitional zone, generally less than 1 m wide. The groundmass in the megacrystic core resembles the equigranular rock in the margins. Megacrysts are generally tabular and range from 2 to 7 cm in the long dimension. The pluton is conspicuously foliated. Locally, the foliation parallels the margins of the pluton,



EXPLANATION

Krl — **Granodiorites of Red Lake(Krl_1), Eagle Peak(Krl_2), and Big Creek(Krl_3)**

Kbm — **Leucogranite of Bald Mountain**

Kbb — **Leucogranite, undivided**—Leucogranites of Big Sandy Bluffs(Kbb_1) and Lion Point(Kbb_2), and leucogranodiorites of Burrough(Kbb_3) and Black(Kbb_4) Mountains

Contact

FIGURE 81.—Generalized geology of intrusions west of the Mount Givens Granodiorite, which are tentatively assigned to the John Muir Intrusive Suite.

but over most of the pluton it strikes northwest and is vertical or steeply dipping. At the northwest end, the foliation appears to continue into a mass of Dinkey Creek Granodiorite. This relation suggests that some of the foliation formed in response to regional stress after most of the pluton had solidified. The granodiorite of Eagle Peak is the youngest plutonic unit in the area and intrudes the Mount Givens Granodiorite of the John Muir Intrusive Suite and the Dinkey Creek Granodiorite, the granodiorite of McKinley Grove, and the granites of Dinkey Dome and of Short Hair Creek of the Shaver Intrusive Suite. A single K-Ar biotite age of 89 Ma was reported by Bateman and Wones (1972b).

The granodiorite of Big Creek forms a small pluton that lies just west of the granodiorite of Red Lake (fig. 81). It is enclosed in the Dinkey Creek Granodiorite and intrudes both the Dinkey Creek and the granite of Shuteye Peak. The rock commonly is medium-grained equigranular granodiorite, but locally it grades to megacrystic granite. Modes plot on a Q-A-P diagram across both the granodiorite and granite fields (fig. 78). The color index is mostly in the range of 8 to 12 but is as low as 2 in a few samples (fig. 79). The pluton is elongate toward the northeast, an unusual trend in the Sierra Nevada. The central part and the northern margin are massive, whereas the southeastern margin and both the east and west ends are foliated parallel to the external contacts of the pluton.

A K-Ar biotite age of 88 Ma may be close to the emplacement age of the granodiorite, as the pluton is not in contact with any younger rocks (Bateman and Wones, 1972b). The similar isotopic ages of the granodiorites of Red Lake and Eagle Peak indicate that they were emplaced at about the same time. Nevertheless, Noyes, Frey, and Wones (1983) concluded, chiefly from studies of minor-element distributions, that the parent magmas for the two granodiorites were generated from different source materials.

Leucogranite of Bald Mountain (Kbm)

The leucogranite of Bald Mountain forms an elongate pluton that intrudes only the Dinkey Creek Granodiorite. The rock is light-colored, equigranular biotite granite, having a color index that generally ranges from 2 to 3 (fig. 79) and a uniform texture and composition. Most of the modes form a tight cluster in the granite field on a Q-A-P diagram (fig. 78). Typical rock is massive and structureless, but faint foliation was observed near the north end, locally along the west side of the pluton where the rock is unusually mafic (possibly because of contamination by the Dinkey Creek Granodiorite), and in the south end where small scattered mafic inclusions are present. Along the east side and south end of the pluton numerous gently dipping felsic dikes emanating from the leucogranite of Bald Mountain penetrate the adjacent Dinkey Creek Granodiorite. Bateman and Wones (1972b) reported a K-Ar biotite age of 91 Ma for this intrusion.

Leucogranites of Big Sandy Bluffs and Lion Point and leucogranodiorites of Burrough and Black Mountains (Kbb)

These bodies of light-colored, equigranular, fine- to medium-grained biotite leucogranite and leucogranodiorite lie considerably west of other rocks assigned to the John Muir Intrusive Suite; the principal reason for including them is a concordant U-Pb age of 93 Ma determined for the granite of Big Sandy Bluffs (Stern and others, 1981). However, they generally contain less alkali feldspar than other rocks of the suite, and future reassignment may be required. The color index of all the rocks in this unit is generally less than 6 (fig. 79) but is higher along some contacts with the Bass Lake Tonalite. Modes of samples are scattered across the granite and granodiorite fields (fig. 78).

GRANITIC ROCKS OF THE WEST SLOPE OF THE SIERRA NEVADA NOT ASSIGNED TO INTRUSIVE SUITES

GRANITOIDS OF JURASSIC AGE IN THE NORTHWEST CORNER OF THE MAP AREA

The tonalite of Granite Creek and the granodiorite of Woods Ridge, two granitoids in the northwest corner of the map area (pl. 1), have Middle Jurassic and Late Jurassic ages, respectively. The tonalite of Granite Creek has been ductilely deformed, presumably during the Nevadan orogeny, whereas the younger granodiorite of Woods Ridge is undeformed except locally, close to contacts with other rock units.

Tonalite of Granite Creek (Jgc)

The tonalite of Granite Creek was called the quartz diorite of Granite Creek by Stern and others (1981) and was included in their Jawbone granitoid sequence. Because the other units of this sequence lie west of the map area (pl. 1) and are so similar in composition and texture to the tonalite of Granite Creek as to suggest that they may actually belong to the same lithodeme, the tonalite of Granite Creek is not assigned to an intrusive suite in this report.

Typical rock is dark-gray, medium-grained, equigranular hornblende-biotite tonalite with gneissic foliation and conspicuous lineation. The foliation in the tonalite is continuous with that in the country rocks and reflects regional deformation. Individual mineral grains generally are smeared and broken, but locally euhedral hornblende and biotite remain intact. On a Q-A-P diagram, most modes plot in the tonalite field, but some plot in adjacent parts of the quartz diorite and granodiorite fields (fig. 82). The color index of most samples ranges from 16 to 30 and averages about 27 (fig. 83). The tonalite of Granite Creek intrudes only Paleozoic(?) metamorphic rocks and is intruded by the granite of Woods Ridge and the granodiorite of Sawmill Mountain. It has a concordant U-Pb age of 163 Ma and a slightly discordant age of 166 Ma (Stern and others, 1981). The rock that forms the Standard pluton west of the map area near Sonora is identical and has yielded a U-Pb age of 164 Ma (Stern and others, 1981).

GRANITE OF WOODS RIDGE (Jwr)

The granite of Woods Ridge (pl. 1) consists of medium- to light-gray, conspicuously megacrystic, splotchy-textured rock. Alkali feldspar forms erratically distributed megacrysts of variable size and shape. Most larger megacrysts are tabular, but some smaller crystals appear to have blocky shapes, probably because of their diverse orientations relative to the exposed surfaces. Biotite is in tiny flakes. On a Q-A-P diagram (fig. 82), modes scattered across the granodiorite and granite fields reflect the erratic distribution of alkali feldspar megacrysts. The color index ranges from 4 to 14 and averages about 8 (fig. 83). The granite of Woods Ridge intrudes the tonalite of Granite Creek and is intruded by the Bass Lake Tonalite. The granite of Woods Ridge has a concordant U-Pb age of 151 Ma (Stern and others, 1981).

TONALITE OF MILLERTON LAKE (Kml)

The tonalite of Millerton Lake is mostly south of the map area, but three lobes extend northward into the map area (pl. 1). The westernmost lobe is 3 km east of the intersection of California State Highways 41 and 145, the middle lobe is about 7 km farther east along the shores of Millerton Lake, and the easternmost lobe extends eastward from Marshal Station to the south flank of Black Mountain. The rock in these lobes closely resembles the Bass Lake Tonalite, and the eastern and middle lobes were included in the tonalite of Blue Canyon on the geologic maps of the Millerton Lake and Raymond quadrangles (Bateman and Busacca, 1982; Bateman and others, 1982). The rock south of Black Mountain was suspected of being older than rocks herein included in the Bass Lake Tonalite and was designated the tonalite south of Black Mountain on the geologic maps of the Shaver and Millerton Lake quadrangles (Lockwood and Bateman, 1976; Bateman and Busacca, 1982). When Saleeby and Sharp (1980) reported isotopic ages for similar rocks just south of the map area that were older than any for the Bass Lake Tonalite, a sample from the north shore of Millerton Lake was collected and isotopically dated by T.W. Stern. This sample yielded a concordant U-Pb age of 134 Ma (table 1)—distinctly older than the Fine Gold Intrusive Suite.

The tonalite of Millerton Lake is undeformed and exhibits mineral foliation parallel to the margins of the lobes. The rock contains almost no alkali feldspar and on a Q-A-P diagram plots along the Q-P sideline of the quartz diorite and tonalite fields (fig. 82). The tonalite of Millerton Lake intrudes Triassic metamorphic rocks and small bodies of diorite. It is intruded by the Bass Lake Tonalite and by granite herein included with the granodiorite of Arch Rock of the Fine Gold Intrusive Suite.

UNASSIGNED GRANITIC ROCKS OF THE YOSEMITE AREA

GRANITE PORPHYRY OF STAR LAKES (Ksl)

The granite porphyry of Star Lakes forms two masses that lie at the northwest end of the Mount Givens Granodiorite (pl. 1). The larger porphyry mass has an irregular but roughly ellipsoidal shape and underlies an area of about 6 km^2; the smaller porphyry mass is elongate and lies between the El Capitan Granite on the south and the granodiorite of Illilouette Creek and the granite of Shuteye Peak on the north. The porphyry ranges widely in composition from leucogranite to granodiorite (fig. 82). It intrudes the El Capitan Granite and has yielded a Rb-Sr whole-rock age of 108±4 Ma; it therefore cannot be included with younger porphyries assigned to the volcanogenic Minarets sequence.

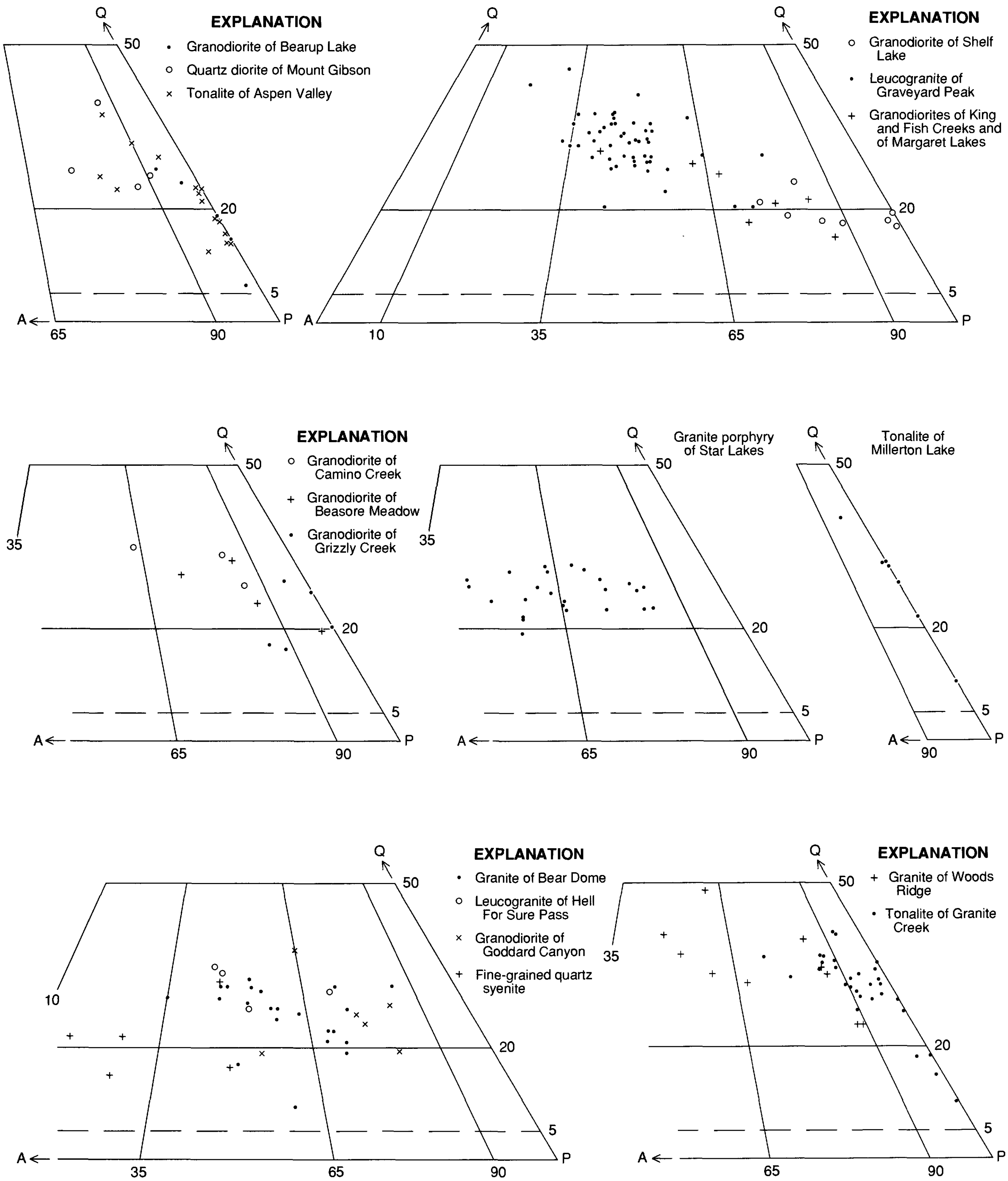


FIGURE 82.—Modes of unassigned intrusions in western and central Sierra Nevada. See figure 6 for Q-A-P classification diagram.

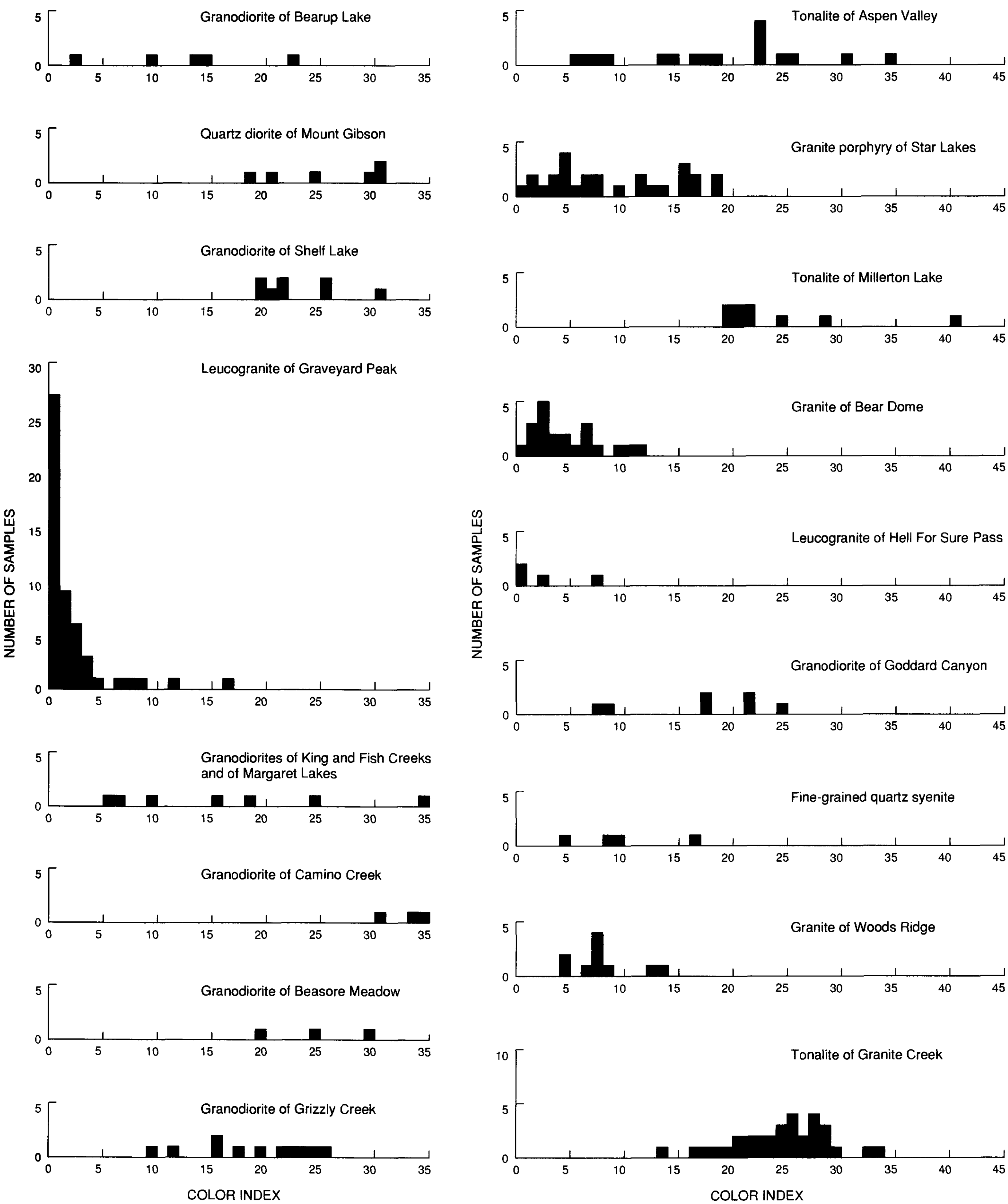


FIGURE 83.—Color indices of unassigned intrusions in western and central Sierra Nevada.

Tonalite of Aspen Valley (Kav)

The tonalite of Aspen Valley forms a small pluton with an irregular shape in the northwestern part of the map area (pl. 1). It is composed of dark, medium-grained biotite-hornblende tonalite and quartz diorite. Kistler (1973) referred to this rock as the quartz diorite of the South Fork of the Tuolumne River. Modes plot in the granodiorite, tonalite, and quartz diorite fields on a Q-A-P diagram (fig. 82), and the color index varies widely, ranging from 5 to 35 (fig. 83). The pluton intrudes the Bass Lake Tonalite of the Fine Gold Intrusive Suite and the El Capitan Granite of the intrusive suite of Yosemite Valley and is intruded by the Sentinel Granodiorite and by the granodiorite of Tamarack Creek of the intrusive suite of Buena Vista Crest.

Quartz diorite of Mount Gibson (Kgi)

The quartz diorite of Mount Gibson lies along the north border of the map area (pl. 1) and extends northward. Typical rock is dark, medium-grained pyroxene-bearing biotite-hornblende quartz diorite and tonalite (fig. 82). The color index ranges from about 18 to 31 (fig. 83). It intrudes the Poopenaut Valley pluton of the Bass Lake Tonalite of the Fine Gold Intrusive Suite and the granite of Rancheria Mountain of the intrusive suite of Yosemite Valley and is intruded by the granodiorite of Bearup Lake.

Granodiorite of Bearup Lake (Kbu)

The granodiorite of Bearup Lake forms a small lenticular pluton that lies along the south margin of the quartz diorite of Mount Gibson (pl. 1). Typical rock is equigranular and is characterized by euhedral hornblende and biotite. The color index ranges widely from about 3 to 23 (fig. 83). The Bearup Lake intrudes the quartz diorite of Mount Gibson and the granodiorite of Rancheria Mountain.

GRANITOIDS AT THE NORTHWEST END OF THE MOUNT GIVENS GRANODIORITE

Granodiorites of Beasore Meadow and Grizzly Creek (Kbe)

The granodiorites of Beasore Meadow and Grizzly Creek form separate plutons that lie on opposite sides of a tongue of the granite of Shuteye Peak (pl. 1). Both the Beasore Meadow and the Grizzly Creek are medium-grained, equigranular hornblende-biotite granodiorite and tonalite (fig. 82), but they differ somewhat in composition and appearance. The granodiorite of Grizzly Creek has a color index that ranges from 10 to 26 (fig. 83) and is characterized by conspicuous rounded quartz phenocrysts averaging about 5 mm across. The granodiorite of Beasore Meadow generally is darker than typical granodiorite of Grizzly Creek and locally contains abundant mafic inclusions (Huber, 1968). Both granodiorites intrude the granite of Shuteye Peak of the Shaver Intrusive Suite.

Granodiorite of Camino Creek (Kca)

The granodiorite of Camino Creek (pl. 1) is fine- to medium-grained rock of variable composition and texture. Typical rock is hornblende-biotite granodiorite but ranges from granodiorite to granite (fig. 82). It commonly contains abundant inclusions of metamorphic rocks, locally in advanced stages of assimilation. It intrudes the granite of Shuteye Peak and the granodiorite of Beasore Meadow.

SHEARED GRANITOIDS THAT INTRUDE THE GODDARD SEPTUM

Granitic rocks of four different compositions intrude the Goddard septum. They are, from north to south, the granite of Bear Dome, the granodiorite of Goddard Canyon, the leucogranite of Hell For Sure Pass, and unnamed fine-grained quartz syenite (pl. 1). The granodiorite of Goddard Canyon, the leucogranite of Hell For Sure Pass, and the fine-grained quartz syenite are pervasively sheared. Indicating a period of regional deformation, this shearing is continuous into the metamorphic rocks of the roof pendant. Because the granitoids have undergone regional deformation and are intruded by mafic dikes thought to be part of the Jurassic Independence dike swarm and because Chen and Moore (1982) reported a concordant U-Pb age of 157 Ma for the largest mass of fine-grained quartz syenite, these granitoids are assigned to the Jurassic.

Fine-grained quartz syenite (Jqs)

This fine-grained xenocrystic rock forms several long, thin bodies, the largest of which is 10 km long and 1 km wide (pl. 1). Modes are widely scattered on a Q-A-P diagram and plot in the quartz syenite, syenogranite, quartz monzonite, and monzonite fields (fig. 82); the color index is highly variable (fig. 83), probably because of strong alteration. The fine grain size and ragged texture show this to be a hypabyssal rock, and its composition indicates secondary enrichment in potassium. The rock is intruded by the Evolution Basin Alaskite of the John Muir Intrusive Suite. It has a concordant U-Pb age of 157 Ma (Chen and Moore, 1982).

Granodiorite of Goddard Canyon (Jgd)

The granodiorite of Goddard Canyon forms an elongate pluton about 10 km long and as much as 1 km wide and two much smaller plutons north of the main pluton. Presumably the largest pluton was emplaced in an elongate form, but undoubtedly shearing has modified its shape. The rock is strongly altered and contains much chlorite, epidote, and other secondary minerals; the rock is so altered that during mapping it was referred to as "the cruddy granodiorite." The few available modes show a wide compositional range on a Q-A-P diagram (fig. 82). This rock is in contact with the leucogranite of Hell For Sure Pass, but strong shearing of both rocks precludes determining which rock is older.

Leucogranite of Hell For Sure Pass (Jh)

The leucogranite of Hell For Sure Pass (pl. 1) is a medium-grained equigranular rock with a color index that is commonly less than 4 (fig. 83). Near Hell For Sure Pass the rock is ribboned with mafic dikes, and both the leucogranite and the dikes are sheared. It is intruded by the Mount Givens Granodiorite.

Granite of Bear Dome (Jbd)

The granite of Bear Dome forms an elliptical body about 11 km long and 6 km wide in the northern part of the Goddard septum (pl. 1). The rock that forms the margins is fine grained and aplitic, whereas the rock in the interior is medium grained. In some places the contact between fine- and medium-grained rock is gradational, but in others the medium-grained rock intrudes the fine-grained rock. On a Q-A-P diagram, more modes plot in the granite field than in the granodiorite field (fig. 82); the color index ranges from 1 to 12 (fig. 83). An observation by Lockwood and Lydon (1975) that the contact of the fine-grained rock with volcanic rocks is gradational suggests that the granite of Bear Dome is the hypabyssal equivalent of some of the metavolcanic rocks. The presence of mafic dikes, presumably of the Jurassic Independence dike swarm, suggests that the Bear Dome is of Jurassic age.

GRANITOIDS BETWEEN THE GODDARD SEPTUM AND THE RITTER RANGE ROOF PENDANT

Granodiorites of King and Fish Creeks and of Margaret Lakes (Kkf)

The granodiorites of King and Fish Creeks and of Margaret Lakes form three plutons that extend toward the southeast from near the southeast corner of the Ritter Range roof pendant (pl. 1). The plutons are separated from one another by the younger Mount Givens Granodiorite and the leucogranite of Graveyard Peak but originally formed a single continuous pluton. The rock in these plutons ranges from fine to coarse grained, is of varied composition and texture (fig. 82), and is strongly contaminated with partially to almost completely assimilated inclusions of metavolcanic rock. Swarms of both mafic and felsic dikes are present locally. The granodiorite of Margaret Lakes generally is sheared. The originally continuous pluton intruded only metavolcanic rocks of the Ritter Range roof pendant and has been intruded by the Mount Givens and Lake Edison Granodiorites, the Mono Creek Granite of the John Muir Intrusive Suite, and the leucogranite of Graveyard Peak.

Leucogranite of Graveyard Peak (Kgp)

The leucogranite of Graveyard Peak (pl. 1) is medium grained equigranular, plots mostly in the granite field on a Q-A-P diagram (fig. 82), and has a very low color index that only locally exceeds 3 (fig. 83). This very irregularly shaped pluton lies between and is intruded by the Lake Edison Granodiorite and the Mono Creek Granite. It is also intruded by small bodies of granite of unknown affiliation, by many small bodies of diorite of variable composition and texture, and by numerous felsic dikes. It intrudes the granodiorite of Margaret Lakes. It has a concordant U-Pb age of 99 Ma (Stern and others, 1981), which suggests that the leucogranite may be cogenetic with the Minarets volcanic sequence and (or) with the intrusive suites of Merced Peak, Buena Vista Crest, and Washburn Lake.

Granodiorite of Shelf Lake (Ksf)

The granodiorite of Shelf Lake (pl. 1) consists of several small bodies of granodiorite that intrude the leucogranite of Graveyard Peak and numerous small bodies of diorite that also intrude the leucogranite of Graveyard Peak. It is intruded by the Lake Edison Granodiorite and Mono Creek Granite. Typical rock is medium-grained hornblende-biotite granodiorite, quartz monzodiorite, and quartz diorite (fig. 82). Hornblende forms prisms that contain pyroxene cores. Mafic inclusions are abundant in the southern part of the largest pluton.

GRANITOIDS SOUTHWEST OF THE GODDARD SEPTUM (Kub, Ktu, Kmr, Kfp, Kwd)

Southwest of the Goddard septum, along and close to the south boundary of the map area (pl. 1), is a group of five intrusions that cannot at present be assigned to intrusive suites (fig. 84). The intrusive relations at contacts permit assigning relative ages to four of these intrusions. In order of decreasing age these intrusions are (1) the granodiorite of upper Blue Canyon (Kub), a medium-grained granodiorite, (2) the granite of Tunemah Lake (Ktu), a felsic biotite granite, (3) the granodiorite of Mount Reinstein (Kmr), a hornblende-biotite granodiorite containing 16 to 23 percent mafic minerals, and (4) the granite of Finger Peak (Kfp), a medium-grained light-colored biotite granite. The map pattern (fig. 84) suggests that the granodiorite of upper Blue Canyon and the granite of Tunemah Lake may belong to the same intrusive suite and that the granodiorite of Mount Reinstein and the leucogranite of Finger Peak both may belong to a younger intrusive suite. However, the spatial relations of the Mount Givens Granodiorite to the granodiorite of Mount Reinstein and the granite of Finger Peak may suggest otherwise. The Mount Givens truncates foliation in the Mount Reinstein and Finger Peak, but a tongue of Finger Peak penetrates 10 km into the Mount Givens. Additional study of the relations in this area is needed. The fifth pluton, the granodiorite of the White Divide (Kwd), is separated from other plutonic rocks within the map area (pl. 1) by a thin offshoot from the Goddard septum, but south of the map area this pluton intrudes and is intruded by several plutonic bodies (Moore, 1978). Modes of samples from these intrusions and their color indices are shown in figure 85. Petrographic descriptions are not available, and no isotopic ages have been determined.

EASTERN SIERRA NEVADA AND THE BENTON RANGE

SCHEELITE INTRUSIVE SUITE

The Scheelite Intrusive Suite is herein formally named for the mining settlement of Scheelite along lower Pine Creek, and the type area is along Pine Creek canyon east of the Pine Creek septum. Stern and others (1981) called this suite the Scheelite granitoid sequence. The suite is of Late Triassic age and is the oldest intrusive suite in the central Sierra Nevada. It occupies much of the eastern escarpment of the Sierra Nevada north of Big Pine and crops out discontinuously through Cenozoic volcanic and sedimentary cover as far east as the Benton Range. The suite has not been identified in the White Mountains. It undoubtedly continues northward beyond the map area (pl. 1). The suite consists of the Wheeler Crest Granodiorite, the granite of Lee Vining Canyon, and the Tungsten Hills Granite. Scattered outcrops indicate that within the map area the suite—mostly the Wheeler Crest Granodiorite—underlies an area of about 3,000 km². The Tungsten Hills Granite underlies an area of about 250 km² at the south end of the suite, and the granite of Lee Vining Canyon, about 200 km² at the northwest end.

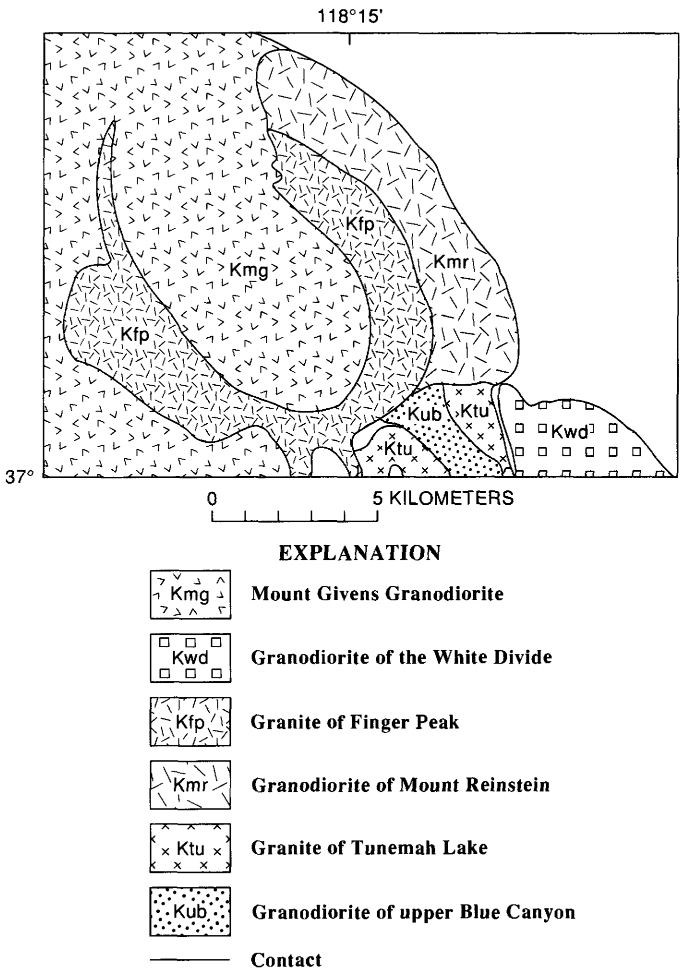


FIGURE 84.—Generalized geology of unassigned granitoids southwest of Goddard septum.

Scattered U-Pb ages for these rocks allow the possibility that the Wheeler Crest Granodiorite is a few million years older than both the Tungsten Hills Granite and the granite of Lee Vining Canyon. The most likely age of the Wheeler Crest Granodiorite is about 214 Ma, whereas the two granites have maximum U-Pb ages of about 200 Ma. However, a dike that cuts the granite of Lee Vining Canyon has yielded a U-Pb age of 210 Ma (Chen and Moore, 1979). Although the two granites are about 50 km apart, they have similar compositions, textures, isotopic ages, and intrusive relations. Additional isotopic dating is required to resolve the uncertainty of the true ages of these rocks.

WHEELER CREST GRANODIORITE (Ŧwc)

Outcrops of the major granitoid of the Scheelite Intrusive Suite in the eastern escarpment of the Sierra Nevada were formally called the Wheeler Crest Quartz Monzonite by Bateman (1961), and this name was retained in other early publications (Rinehart and Ross, 1964; Bateman, 1965b; Huber and Rinehart, 1965); identical rocks just east of the Sierra Nevada already informally had been called the quartz monzonite of Wheeler Crest (Rinehart and Ross, 1957), and those in the Benton Range, farther to the east, had been called the granodiorite of the Benton Range by Rinehart and Ross (1957) and later by Krauskopf and Bateman (1977). The division between the names Wheeler Crest and Benton Range originally was made in the Casa Diablo Mountain quadrangle, where Rinehart and Ross (1957) assigned discontinuous outcrops in the southwestern part of the quadrangle to the quartz monzonite of Wheeler Crest and outcrops in the northeastern part to the granodiorite of the Benton Range even though they considered the two units equivalent. The formal name Wheeler Crest Granodiorite is herein applied to all these rocks. The change in the lithologic designation from quartz monzonite to granodiorite is required because modal plots show the median composition of the unit to be granodiorite (fig. 85).

Typical Wheeler Crest Granodiorite is medium grained and megacrystic and has a color index that ranges from 2 to 21 and averages about 10 (fig. 85). Tabular alkali feldspar megacrysts 1 to 3 cm across and 2 to 10 cm long are embedded in a groundmass in which the felsic minerals are 2 to 4 mm across. In a few places the rock grades from megacrystic to equigranular by a decrease in the abundance of megacrysts. The texture of the equigranular rock is identical to that of the groundmass of the megacrystic rock. Small clusters of ragged grains of biotite, hornblende, apatite, titanite, allanite, zircon, opaque minerals, chlorite, and epidote are ubiquitous. Rocks having a color index less than 10 commonly contain little or no hornblende, and rocks in which the color index is greater than 10 generally contain more hornblende than biotite. Compositional zoning in the plagioclase is generally within the range An_{40-20}, but the ranges for individual grains differ widely; mottled cores of An_{45-40} or even more calcic plagioclase are present in some grains.

In most places the rock lacks foliation. Lensoid mafic inclusions are uncommon, and tabular and prismatic minerals show no preferred orientation. However, in the east face of Wheeler Ridge and on both sides of Pine Creek canyon the rock is foliated, lensoid mafic inclusions are common, and megacrysts tend to be parallel. Nevertheless, granoblastic mortar that borders larger grains indicates that some of the foliation may be secondary, probably the result of forcible intrusion by units of the Cretaceous John Muir Intrusive Suite.

The Wheeler Crest Granodiorite is the oldest granitoid that has been identified in the central Sierra Nevada. It intrudes only metamorphic rocks. Small bodies of diorite and gabbro originally interpreted to be older than the Wheeler Crest (Bateman, 1965b) probably are younger. It is also intruded by the Tungsten Hills Granite, also of the Scheelite Intrusive Suite, the leucogranite of Casa Diablo Mountain, the Round Valley Peak Granodiorite of the John Muir Intrusive Suite, and several small bodies of granite that probably are of Cretaceous age.

The Wheeler Crest Granodiorite has yielded concordant U-Pb ages of 217 Ma, 210 Ma (for a different size fraction), and 207 Ma and discordant ages of 201 and 161 Ma (Stern and others, 1981; Chen and Moore, 1982). Several K-Ar hornblende ages range from 211 to 98 Ma (Kistler and others, 1965; Evernden and Kistler, 1970; Crowder and others, 1973). The younger K-Ar hornblende ages were determined for samples collected from Wheeler Crest and are thought to have been reset during intrusion of the nearby Cretaceous John Muir Intrusive Suite; the three earliest K-Ar ages, 211, 211, and 199 Ma, are for samples from localities far from younger intrusions. K-Ar ages on biotite range from 215 to 71 Ma.

GRANITE OF LEE VINING CANYON (Ŧlv)

The granite of Lee Vining Canyon forms a long narrow pluton that crops out in the lower slopes of

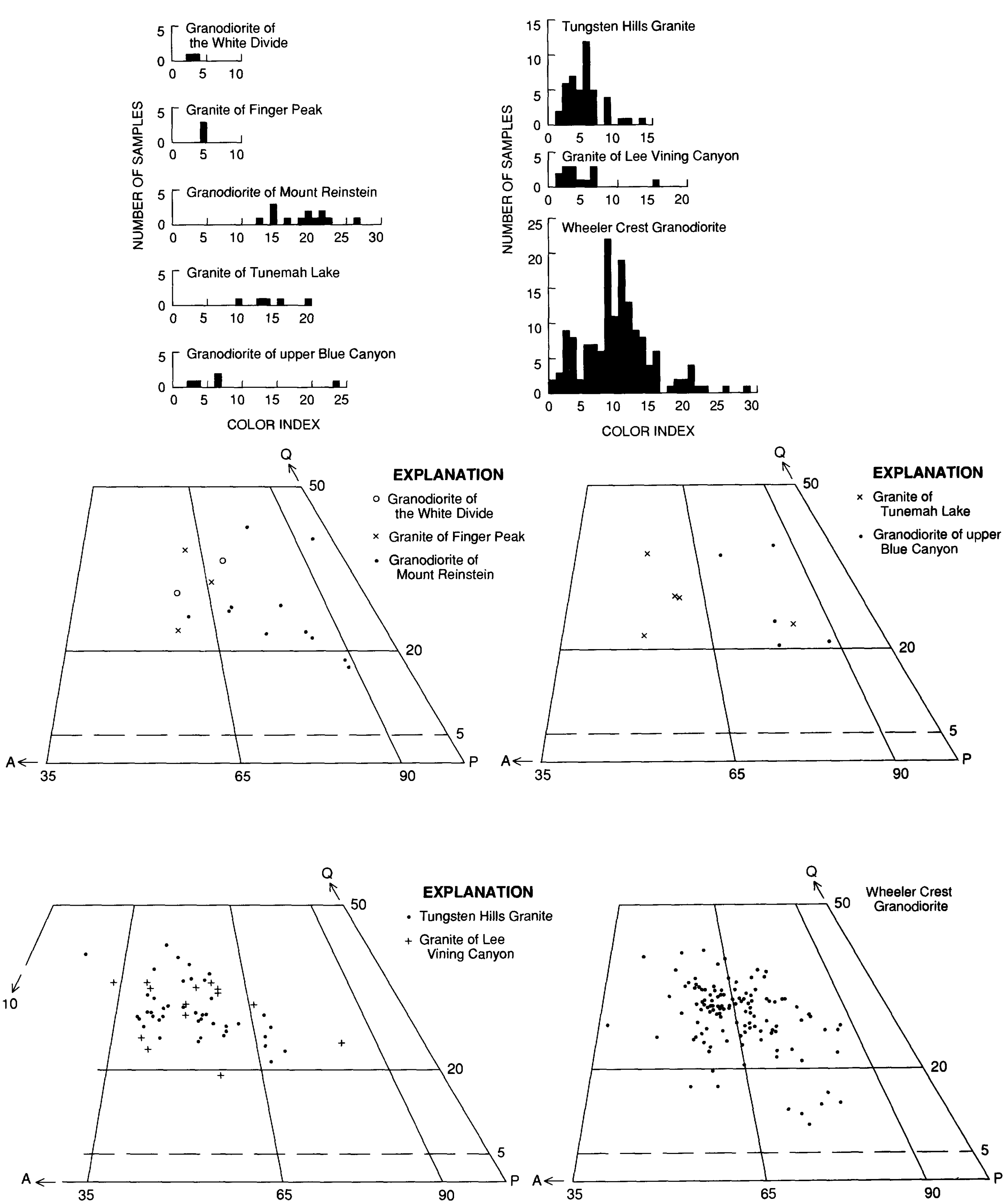


FIGURE 85.—Modes and color indices of the Scheelite Intrusive Suite and unassigned intrusions southwest of Goddard septum. See figure 6 for Q-A-P classification diagram.

the eastern escarpment of the Sierra Nevada from Mono Lake southward for about 26 km. The rock is light colored in most places but is darker in the margins exposed along Lee Vining Canyon and west of Grant Lake. The texture varies from medium grained and equigranular in the margins to megacrystic in the interior; the color index generally ranges from 2 to 9 (fig. 85). Biotite is the principal mafic mineral, and hornblende occurs only in trace amounts. Accessory minerals include magnetite, titanite, allanite, apatite, zircon, and garnet. Kistler (1966a) reported that adjacent to metamorphic rocks the granite is aplitic and is laden with recrystallized and partly digested inclusions.

The granite of Lee Vining Canyon is intruded by the granodiorite of Mono Dome, the leucogranite of Ellery Lake, and the quartz monzodiorite of Aeolian Buttes. Chen and Moore (1982) reported a concordant U-Pb age of 201 Ma and a discordant age of 154 Ma. However, they also reported a concordant age of 210 Ma for a dike that cuts the Lee Vining Canyon and concluded that the preferred age of the granite is about 210 Ma.

TUNGSTEN HILLS GRANITE (₮t)

The Tungsten Hills Granite (pl. 1) is part of a unit that originally was called the Tungsten Hills Quartz Monzonite (Bateman, 1965b). The excluded rocks lie west of the Pine Creek septum and have been reassigned to the Cretaceous Lake Edison Granodiorite of the John Muir Intrusive Suite. The lithologic designation of the Tungsten Hills is herein changed to reflect the current IUGS classification (Streckeisen, 1973), as shown in figure 85.

Typical Tungsten Hills Granite is medium-grained, seriate or weakly megacrystic, light-colored rock with a color index in the range of 2 to 9 and averaging about 5 (fig. 85). A foliation is not apparent. Mafic inclusions are scarce, and mineral grains show no evidence of preferred orientation.

Subhedral to anhedral grains of quartz, alkali feldspar, and plagioclase are present in almost equal amounts. Plagioclase is zoned from about An_{34} to An_{20} and has rims of subsolidus albite. Mottled calcic cores have not been observed. Biotite is the principal mafic mineral and locally is accompanied by minor amounts of hornblende. Common accessories are titanite, apatite, allanite, zircon, and opaque minerals; less common accessories include monazite and thorite.

The Tungsten Hills Granite intrudes the Wheeler Crest Granodiorite and is intruded by the leucogranite of Rawson Creek, the granodiorite of Coyote Flat, and small masses of fine-grained granite and aplite, all of Cretaceous age. Isotopic ages include concordant U-Pb ages of 202 and 197 Ma and a discordant age of 167 Ma (Stern and others, 1981; Chen and Moore, 1982). K-Ar biotite ages of 77 and 76 Ma obviously have been reset (Kistler and others, 1965); the true age of the Tungsten Hills Granite appears to be about 200 Ma.

PALISADE CREST INTRUSIVE SUITE

The Palisade Crest Intrusive Suite, called the Palisade Crest sequence by Bateman and Dodge (1970) and the Palisade Crest granitoid sequence by Stern and others (1981), is herein formally named for exposures along and west of the Palisade Crest, a part of the Sierra Nevada divide west of Big Pine. The type area is in the eastern escarpment of the Sierra Nevada from Big Pine Creek south to Red Mountain Creek. Within the map area the suite crops out in an area of only about 150 km^2, but it extends eastward beneath the sediments of Owens Valley and into the Inyo Mountains. Within the map area the suite comprises two intrusive units, the Tinemaha Granodiorite and the granodiorite of McMurry Meadows, and it probably also includes the leucogranites of Red Mountain and Taboose Creeks (fig. 86). Bateman and Dodge (1970) originally included the Inconsolable Quartz Monzodiorite, formerly called the Inconsolable Granodiorite (Bateman, 1965b), in the same suite, but isotopic dating by the Rb-Sr method indicates that this intrusion has a Cretaceous age (R.W. Kistler, written commun., 1983).

The granodiorite of McMurry Meadows is nested within the Tinemaha Granodiorite and is the younger unit. Although these units differ in grain size and texture, both are characterized by slightly higher contents of alkali feldspar than quartz, wide (but overlapping) ranges in the abundance of plagioclase, and the presence of augite cores in much of the hornblende. Although both rock units are called granodiorite, few modes of either unit plot in the granodiorite field on a Q-A-P diagram (fig. 86); the axes of their elongate plots pass from the quartz monzodiorite field into the granite field through the common corner of these fields with the granodiorite and quartz monzonite fields.

TINEMAHA GRANODIORITE (Jtn)

Most of the Tinemaha Granodiorite exposed within the map area (pl. 1) is in the eastern escarpment of the Sierra Nevada southwest of Big Pine, where it forms two lobes that are separated by the oval-shaped granodiorite of McMurry Meadows. However, it undoubtedly is continuous beneath the

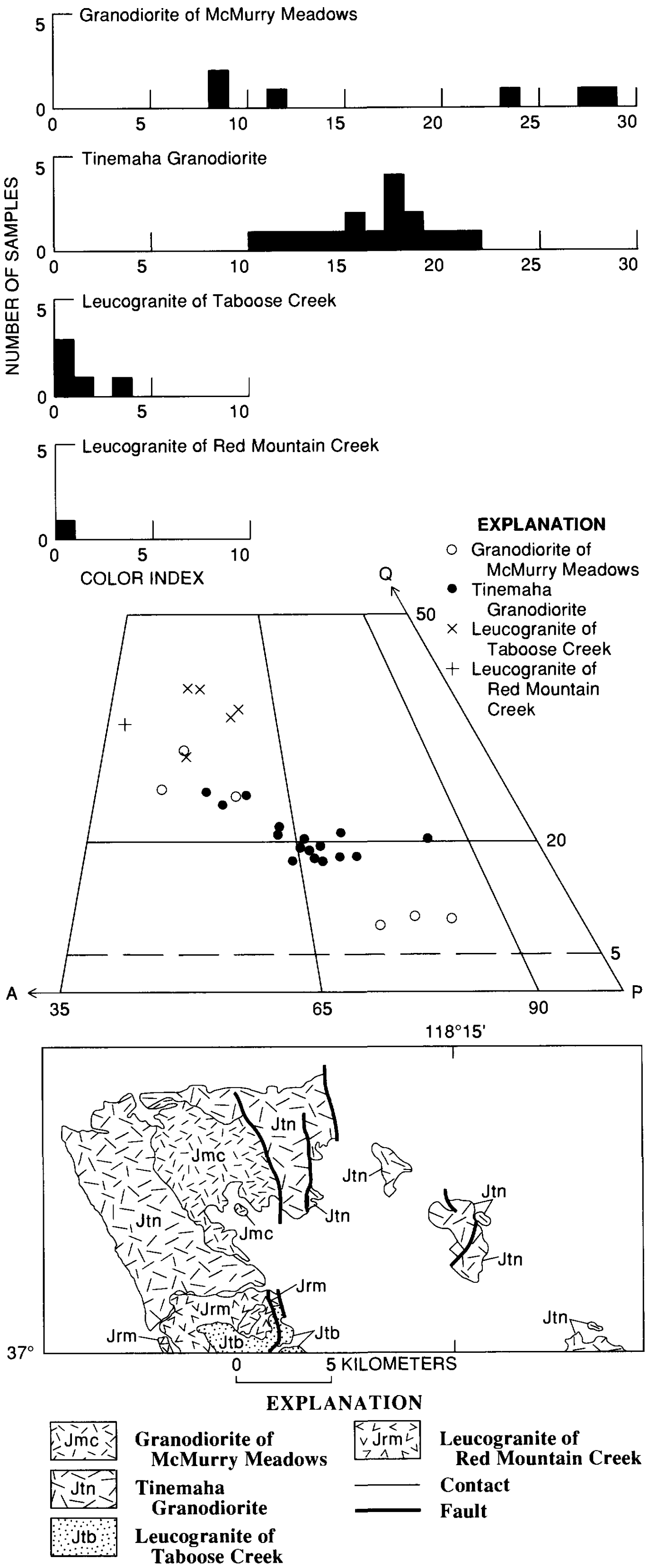


FIGURE 86.—Generalized geology, modes, and color indices of the Palisade Crest Intrusive Suite. Leucogranites of Red Mountain and Taboose Creek are tentatively assigned to the Palisade Crest Intrusive Suite. See figure 6 for Q-A-P classification diagram.

alluvial deposits of Owens Valley with exposures in Fish Springs Hill and the Inyo Mountains. The rock has much the same appearance everywhere despite a wide range of composition, especially reflected in the plagioclase content. Typical rock is seriate to weakly porphyritic and contains blocky anhedral to subhedral alkali feldspar megacrysts as much as 1.5 cm across, but in some places the rock is equigranular. The megacrysts are somewhat smaller than those in most other rocks and have poorly defined crystal faces. The groundmass is hypidiomorphic-granular. Grains of feldspar and quartz range from 2 to 4 mm across, and the color index averages about 17 (fig. 86). Quartz generally is slightly less abundant than alkali feldspar, and plagioclase is present in highly variable amounts. These relative abundances produce a modal plot that is elongate away from the plagioclase corner on a Q-A-P diagram (fig. 86). Although the rock is called granodiorite, modes plot in the monzodiorite, quartz monzonite, granodiorite, and granite fields.

Plagioclase is commonly normally zoned from about An_{40} to An_{30}, but more calcic cores and less calcic rims are common. A unique characteristic of the Tinemaha is that hornblende generally is in excess of biotite in the ratio of almost 2:1. Much of the hornblende is in euhedral or subhedral prisms, whereas biotite is rarely euhedral. Thin sections of samples from the western margin of the Tinemaha reveal augite cores in the hornblende.

The Tinemaha contains many mafic inclusions, and in most places a foliation is defined by these inclusions and by planar orientation of biotite and hornblende. In the western lobe of the Tinemaha, steeply dipping foliation in the western part and horizontal or gently dipping foliation in the eastern part (adjacent to the granodiorite of McMurry Meadows) define a foliation arch. Sparse observations within the eastern lobe suggest the presence of gently dipping foliations there as well. The flat and gently dipping foliations indicate proximity to the roof of the intrusion in the areas where they occur.

The Tinemaha Granodiorite is intruded by the cogenetic granodiorite of McMurry Meadows and possibly cogenetic leucogranites of Red Mountain Creek and Taboose Creek, by the Lamarck Granodiorite of the John Muir Intrusive Suite, and by the unassigned Inconsolable Quartz Monzodiorite and leucogranite of Rawson Creek. It has yielded U-Pb ages of 164 Ma (Chen and Moore, 1982) and 155 Ma (Stern and others, 1981), K-Ar hornblende ages of 187, 184, and 174 Ma (Kistler and others, 1965), and a Rb-Sr whole-rock age of 169±8 Ma (R.W. Kistler, written commun., 1983). The 164-Ma U-Pb age is concordant and probably better approximates the true age of the unit than either the 155-Ma U-Pb age, which is discordant, or the 169±8-Ma Rb-Sr age, which, however, permits an age of 164 Ma. Mafic dikes of the Independence dike swarm, which have yielded an age of 148 Ma (Chen and Moore, 1979), cut the Tinemaha and indicate that it is older than 148 Ma. Compositional variations within the Tinemaha Granodiorite and their origin are discussed in the section "Tinemaha Granodiorite" under "Patterns Within Intrusive Units."

GRANODIORITE OF MCMURRY MEADOWS (Jmc)

The granodiorite of McMurry Meadows forms an elliptical pluton that intrudes and is enclosed by the Tinemaha Granodiorite. It underlies an area of about 30 km^2 and is strongly zoned; effectively bimodal, the McMurry Meadows grades through a narrow transition zone from quartz monzodiorite in the margins to granite in the interior (fig. 86). Hand specimens from the margins are equigranular and contain as much as 17 percent hornblende, whereas specimens of granitic composition contain less than 1.5 percent hornblende but have conspicuous alkali feldspar megacrysts. The plagioclase in rocks of both compositions is strongly zoned from An_{50} to An_{20}. Tiny cores ranging from An_{70} to An_{50} are present in some plagioclase grains.

The granodiorite of McMurry Meadows intrudes only the Tinemaha Granodiorite and is intruded by granite dikes from the adjacent ~95-Ma granite of Rawson Creek and by small masses of diorite. The similarity of its modal and major-element compositions to those of the Tinemaha Granodiorite suggests that it crystallized from the same parent magma, somewhat modified, following a surge that produced the sharp contact between the two intrusions and a slightly reduced initial $^{87}Sr/^{86}Sr$.

POSSIBLE ADDITIONAL UNITS OF THE PALISADE CREST INTRUSIVE SUITE

LEUCOGRANITES OF RED MOUNTAIN CREEK (Jrm) AND TABOOSE CREEK (Jtb)

The closely related leucogranites of Red Mountain and Taboose Creeks (pl. 1) form a nested pair of intrusions in the eastern escarpment of the Sierra Nevada at the south boundary of the map area (fig. 86). Both leucogranites extend south of the map area, where they are cut off by younger intrusions (Moore, 1963); before these younger plutons were emplaced, the leucogranite of Taboose Creek probably was entirely enclosed within and at least partly overlain by the leucogranite of Red Mountain Creek (Bateman, 1965b).

Both leucogranites generally contain less than 1 percent biotite and accessory minerals (chiefly apatite, zircon, allanite, magnetite, and pyrite). However, the leucogranite of Red Mountain Creek is medium grained whereas the leucogranite of Taboose Creek is fine grained and has a mottled texture that may reflect loss of volatiles during crystallization. In outcrops and hand specimens of the Taboose Creek, thin intricately branching dikelets of lighter colored rock can be seen to anastomose through slightly darker rock, but the two rocks are indistinguishable under the microscope.

On the north, northwest, and southwest sides, the two leucogranite plutons are bordered by schist, and the disposition of roof remnants of schist and Tinemaha Granodiorite strongly indicates that before erosion schist overlay the leucogranites just above the present level of exposure. The enclosure of the Taboose Creek within the Red Mountain suggests that the Taboose Creek represents the residual core magma of the leucogranite of Red Mountain Creek. The leucogranite of Red Mountain Creek intrudes the Tinemaha Granodiorite and is intruded by mafic dikes of the Independence dike swarm, whose U-Pb age is 148 Ma (Chen and Moore, 1979); south of the area of figure 86, the leucogranite is intruded by younger granitoids. The Jurassic age of these younger intrusions is the only reason for considering them to be late units of the Palisade Crest Intrusive Suite.

GRANITIC ROCKS NOT ASSIGNED TO INTRUSIVE SUITES

Many masses of plutonic rock in the eastern Sierra Nevada and Benton Range have not been assigned to intrusive suites. These are (1) the sheared granites of Koip Crest and the South Fork of Bishop Creek, (2) the granite of Chickenfoot Lake, (3) the leucogranite of Casa Diablo Mountain, (4) the Inconsolable Quartz Monzodiorite, (5) the leucogranite of Rawson Creek, (6) the granodiorites of Mono Dome and Tioga Lake, (7) the granite of June Lake, (8) the quartz monzodiorite of Aeolian Buttes, the granite of Mono Lake, and the leucogranites of Ellery Lake and Williams Butte, and (9) the granodiorites of Coyote Flat and Cartridge Pass.

SHEARED GRANITES OF KOIP CREST AND THE SOUTH FORK OF BISHOP CREEK (Jsh)

Lenticular masses of gneissic granite and granodiorite crop out at Koip Crest and along the South Fork of Bishop Creek between Long Lake and

Bishop Pass. These gneissic rocks are older than the adjacent granitic rocks and were involved in an episode of regional deformation that did not affect the younger, bordering granitoids. They probably are of Middle Jurassic age.

GRANITE OF CHICKENFOOT LAKE (Jcf)

The granite of Chickenfoot Lake forms an elliptical pluton that lies a few kilometers northwest of the Pine Creek septum (pl. 1). A Rb-Sr age of 172±8 Ma (R.W. Kistler, written commun., 1985) is approximately the same as isotopic ages for the Soldier Pass Intrusive Suite in the White Mountains. Bateman (1965b) erroneously included this pluton in the Cretaceous Lamarck Granodiorite. The composition and the texture of the granite of Chickenfoot Lake vary widely. The rock in the northern part of the pluton is medium grained and weakly megacrystic, but the rock is increasingly finer grained and more mafic toward the south. On a Q-A-P diagram, most modes plot in the granite field, but the color index is higher than that of most granites, ranging from 10 to 19 (fig. 87). Mafic dikes, presumably of the Independence dike swarm, are locally abundant. Shear zones are common in the dikes as well as in the granite. The granite of Chickenfoot Lake is intruded by the Morgan Creek mass (Bateman, 1965b) of the Lake Edison Granodiorite, the Round Valley Peak Granodiorite, and the Mono Creek Granite, all of the John Muir Intrusive Suite.

LEUCOGRANITE OF CASA DIABLO MOUNTAIN (Jcd)

The leucogranite of Casa Diablo Mountain occurs in several masses that extend northward from Casa Diablo Mountain into the Benton Range and northwestward into Granite Mountain (pl. 1). The rock is medium-grained biotite leucogranite with a very low color index, mostly less than 5 (fig. 87). Texturally and compositionally, it closely resembles the leucogranites of Rawson Creek and Graveyard Peak, but it is separated from these rocks by many kilometers and has a much older concordant U-Pb age, 161 Ma (Stern and others, 1981).

INCONSOLABLE QUARTZ MONZODIORITE (Kin)

The Inconsolable Quartz Monzodiorite, formerly called the Inconsolable Granodiorite by Bateman (1965b), is herein renamed because of changes in the classification system (Streckeisen, 1973) since it was originally named. It forms an elongate pluton that lies northwest of the Tinemaha Granodiorite along the Sierran crest (pl. 1). Almost all the modes plot in the quartz monzodiorite field on a Q-A-P diagram (fig. 87). Typical rock is medium dark gray and has an average grain size of about 2 mm. In hand specimen, the rock appears equigranular, but thin sections show it to be seriate. The largest grains are plagioclase. The somber gray color is partly explained by a color index that ranges from about 12 to as much as about 31 and partly results from the prevalent gray to dark-grayish-red feldspar. In many places, especially near the southeast margin, small but conspicuously reddish plagioclase grains are scattered throughout the rock. The Inconsolable contains abundant mafic inclusions that are progressively flattened toward the margins of the monzodiorite and define a foliation that approximately parallels the long axis of the pluton.

Alkali feldspar generally is a little more abundant than quartz, which makes up less than 20 percent of the rock; plagioclase is twice as abundant as alkali feldspar, but the quite variable amount of plagioclase causes elongation of the modal plot toward the plagioclase corner of a Q-A-P diagram (fig. 87). Plagioclase generally is normally zoned in the range An_{50} to An_{30}, but even more calcic, mottled cores are common. Both quartz and alkali feldspar are interstitial to plagioclase. Biotite predominates among the mafic minerals, but both augite and hornblende are present. Because augite appears to be more abundant than hornblende, probably the water content of the magma was low. Generally augite is rimmed by hornblende, but hornblende also occurs as separate grains.

The Inconsolable Quartz Monzodiorite is intruded by the Cretaceous Lamarck Granodiorite of the John Muir Intrusive Suite and the unassigned leucogranite of Rawson Creek. The Inconsolable also is in contact with the ~164-Ma Tinemaha Granodiorite of the Palisade Crest Intrusive Suite; however, the contact is so featureless that the relative ages of the two units have not been determined by field observations. Nevertheless, the Inconsolable Quartz Monzodiorite has a Rb-Sr whole-rock age of 105±11 Ma (R.W. Kistler, unpub. data, 1983) and is therefore Cretaceous and younger than the Jurassic Tinemaha. Until this determination was made, the Inconsolable Quartz Monzodiorite was considered to be older than the Tinemaha Granodiorite (Bateman, 1965b).

LEUCOGRANITE OF RAWSON CREEK (Kra)

The leucogranite of Rawson Creek extends along the eastern escarpment of the Sierra Nevada from the Tungsten Hills southward to Big Pine Creek and

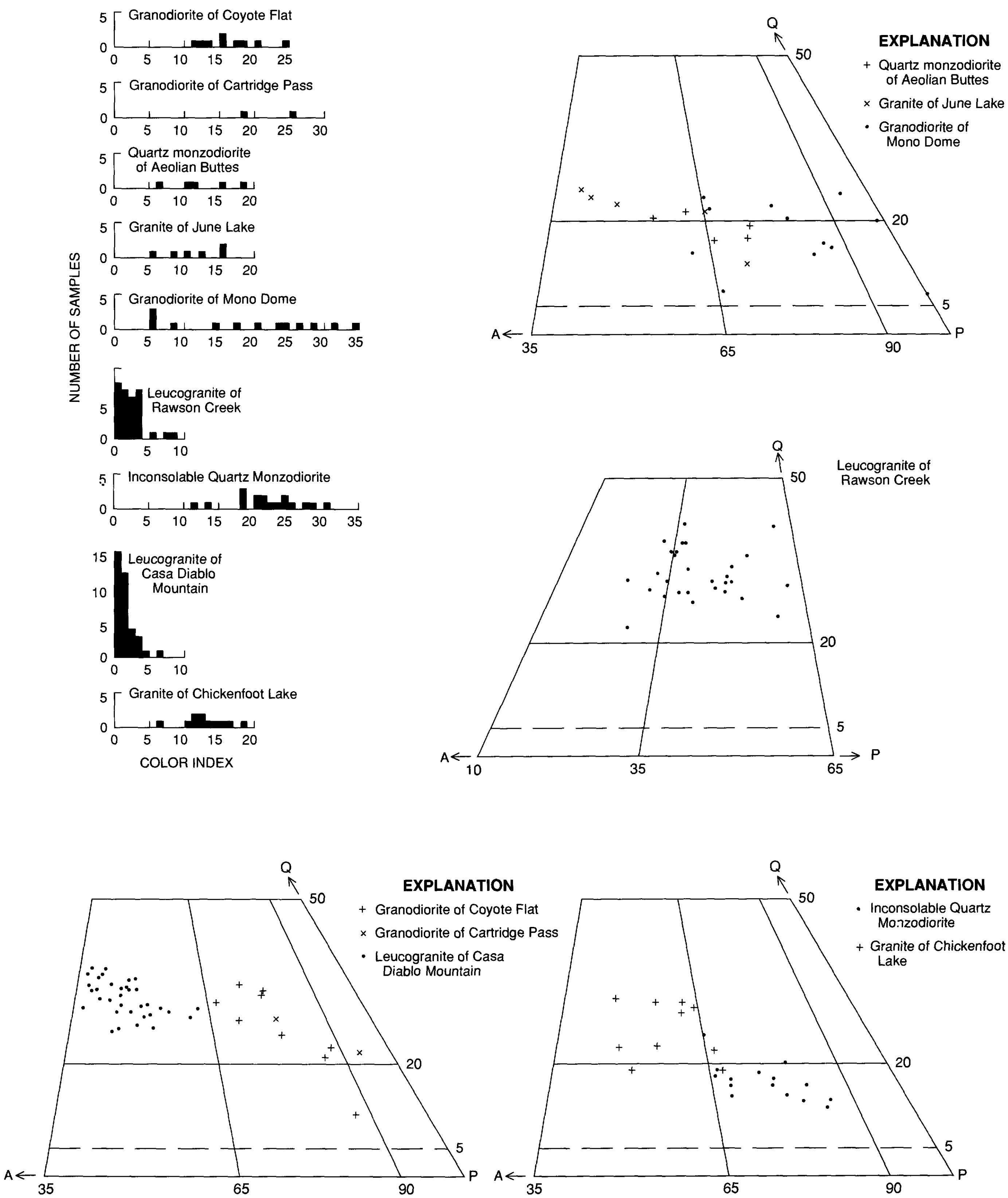


FIGURE 87.—Modes and color indices of unassigned granitic rocks of eastern Sierra Nevada and Benton Range. See figure 6 for Q-A-P classification diagram.

westward to Mount Alice and other parts of the High Sierra west of Coyote Flat (pl. 1). The rock is equigranular and medium grained; the grain size generally ranges from 3 to 4 mm. Biotite is the only mafic mineral, and rarely exceeds 4 percent of the rock (fig. 87). The plagioclase composition is variable, but most compositions, not including subsolidus albite, are within the range An_{30-10}. The leucogranite intrudes the Tungsten Hills Granite of the Scheelite Intrusive Suite, the Tinemaha Granodiorite and granodiorite of McMurry Meadows of the Palisade Crest Intrusive Suite, and the Inconsolable Quartz Monzodiorite. It is intruded by small masses of fine-grained granite and probably also by the granodiorite of Coyote Flat, although the relative ages of these two rocks has not been established by field observations. The leucogranite of Rawson Creek has a mildly discordant U-Pb age of 95 Ma (Stern and others, 1981) and K-Ar ages on biotite of 89, 87, and 83 Ma (Kistler and others, 1965).

GRANODIORITES OF MONO DOME AND TIOGA LAKE (Kmd)

The granodiorite of Mono Dome forms an elongate northwest-trending pluton in the northern margin of the map area west of Lee Vining, and the granodiorite of Tioga Lake forms a much smaller body that extends about 3 km toward the southeast from Tioga Lake (pl. 1). These are light- to dark-gray, medium-grained rocks. The Mono Dome grades from granodiorite to quartz monzodiorite (fig. 87); it contains pyroxene as well as amphibole. Much of the amphibole is altered to chlorite. The granodiorite of Mono Dome intrudes the granite of Lee Vining Canyon and is intruded by the granite of June Lake; R.W. Kistler (written commun., 1983) determined a whole-rock Rb-Sr age of 93±6 Ma for the granodiorite of Mono Dome. The granodiorite of Tioga Lake is in contact only with metamorphic rocks.

GRANITE OF JUNE LAKE (Kjl)

The granite of June Lake originally was mapped as part of the Wheeler Crest Granodiorite (Huber and Rinehart, 1965; Kistler, 1966a) but is now recognized as a separate intrusion (pl. 1). It is a medium-grained, megacrystic rock somewhat resembling the Wheeler Crest Granodiorite but with smaller alkali feldspar megacrysts, averaging only about 1 cm in greatest dimension. It intrudes the granodiorite of Mono Dome.

QUARTZ MONZODIORITE OF AEOLIAN BUTTES, GRANITE OF MONO LAKE, AND LEUCOGRANITES OF ELLERY LAKE AND WILLIAMS BUTTE (Kae)

Under this designation are included several small masses of granitic rock. The most extensive of these are the granite of Mono Lake and the leucogranite of Ellery Lake. The granite of Mono Lake and the leucogranites of Ellery Lake and Williams Butte have not been described. Exposures of contacts of the leucogranite of Ellery Lake with the granite of Lee Vining Canyon of the Scheelite Intrusive Suite and the granodiorite of Mono Dome are inadequate for determining the relative ages of the rocks; however, the leucogranite of Ellery Lake is presumed to be younger than the Lee Vining Canyon and Mono Dome rocks. The leucogranite of Williams Butte intrudes the granodiorite of Mono Dome. The granite of Mono Lake intrudes only metamorphic country rocks, and thus its age relative to other rocks included herein is unknown.

The quartz monzodiorite of Aeolian Buttes is exposed south of Mono Lake in a series of small isolated outcrops, most of which are surrounded by Cenozoic deposits. The rock is white to light gray and strongly jointed. Biotite books, hornblende needles, and large subhedra of titanite are characteristic; accessory minerals are magnetite, titanite, apatite, allanite, and epidote. On a Q-A-P diagram modes plot in adjacent parts of the quartz monzodiorite and granite fields (fig. 87). The color index ranges from 7 to 19. The rock is in contact with the granites of Lee Vining Canyon and June Lake, and at both contacts the rock is sheared. As other contacts with the granites of Lee Vining Canyon and June Lake are unsheared, it seems probable that the shearing is the result of the later emplacement of the quartz monzodiorite of Aeolian Buttes.

GRANODIORITES OF COYOTE FLAT AND CARTRIDGE PASS (Kcf)

The granodiorites of Coyote Flat and Cartridge Pass form two small subequant plutons. The Coyote Flat lies along the east side of the Bishop Creek septum. Most of the Cartridge Pass lies south of the map area (pl. 1) but is exposed along its south boundary just west of the Sierra Nevada divide. Both plutons are compositionally zoned. The margin of the Coyote Flat pluton is darker and finer grained than the interior, and the Cartridge Pass pluton is continuously zoned from granodiorite inward to a granite core (Moore, 1963). The three modally analyzed samples from the Coyote Flat pluton and the two from the Cartridge Pass pluton that plot closest to the plagioclase corner on a Q-A-P diagram are from the margin of the plutons (fig. 87). The grouping of the other Coyote Flat modes in the granodiorite field show that the composition of the interior of the pluton is nearly constant.

The rock in both plutons is light gray, equigranular, and fine to medium grained. Both biotite and hornblende are present. A striking feature of the Coyote Flat pluton is that large anhedral grains of quartz and alkali feldspar enclose all the other minerals. The plagioclase is strongly zoned. Tiny cores as calcic as An_{70} are enclosed in plagioclase that ranges in composition from An_{44} to An_{36} and that in turn grades outward toward grain margins to compositions as sodic as An_{20}.

Both plutons appear to intrude all the rock with which they are in contact. The granodiorite of Cartridge Pass intrudes the Lamarck Granodiorite and the Evolution Basin Alaskite; the granodiorite of Coyote Pass intrudes the Tungsten Hills Granite and probably the leucogranite of Rawson Creek. The granodiorite of Coyote Flat has a K-Ar biotite age of 90 Ma (Kistler and others, 1965), and the granodiorite of Cartridge Pass has an age of 81 Ma (Dodge and Moore, 1968).

WHITE AND INYO MOUNTAINS

Both Jurassic and Cretaceous granitic rocks are present in the White and northern Inyo Mountains. The Jurassic rocks range from dark-colored monzonite to granite and have alkaline affinities whereas the Cretaceous granitic rocks are leucocratic equigranular and megacrystic biotite granites. Some of the Jurassic rocks are assigned to the Soldier Pass Intrusive Suite, but some Jurassic rocks and all the Cretaceous granites are unassigned.

The plutonic rocks of the northern White Mountains are of considerable historic interest because Anderson (1937) proposed that his Boundary Peak Granite, which included all the Cretaceous equigranular granites of the northern White Mountains, granitized the overlying sedimentary rocks to produce his Pellisier granite, which included the Jurassic quartz monzonite of Mount Barcroft and the granodiorite of Cabin Creek as well as the granite of Pellisier Flats. However, Emerson (1966) convincingly showed that the granites of the northern White Mountains formed from magmas and are not granitized sedimentary rocks.

SOLDIER PASS INTRUSIVE SUITE

The Soldier Pass Intrusive Suite is herein formally named for Soldier Pass, and the area between Buckhorn Spring on the southeast side of Deep Springs Valley and Cottonwood Creek is designated the type area. This suite comprises five intrusions in the southern White and northern Inyo Mountains: the monzonite of Eureka Valley, the monzodiorite of Marble Canyon, the monzonite of Joshua Flat, the granodiorite of Beer Creek, and the Cottonwood Granite. Only three of these—the monzonite of Joshua Flat, the granodiorite of Beer Creek, and the Cottonwood Granite—crop out within the map area (pl. 1) where they underlie an area of about 25 km^2; the other two are not described in this report. The rocks of the Soldier Pass Intrusive Suite are alkali-calcic and are characterized by a high K_2O content and a relatively low quartz content. Textures in the monzonites show that in these rocks—unique among the granitoids within the map area—quartz rather than alkali feldspar was the last mineral to begin to crystallize.

On the geologic maps of the Blanco Mountain and Mount Barcroft quadrangles, the Cottonwood Granite and the granodiorite of Beer Creek both are included in the quartz monzonite of Beer Creek (Nelson, 1966a; Krauskopf, 1971). However, in 1974, while collecting samples for modal and chemical analysis, K.B. Krauskopf observed a contact in the northern part of the Blanco Mountain quadrangle between leucocratic, generally coarse-grained and megacrystic rock on the north and fine- to medium-grained, equigranular, gray rock on the south. Q-A-P plots of modal analyses of the samples he collected confirm that these are two distinct intrusions (fig. 88). In this report the equigranular rock on the south is designated the granodiorite of Beer Creek, and the megacrystic rock on the north is designated the Cottonwood Granite following earlier usage by Emerson (1966).

MONZONITE OF JOSHUA FLAT (Jj)

Within the map area (pl. 1) the monzonite of Joshua Flat occupies small areas on either side of Deep Springs Valley. The rock is weakly foliated, fine to medium grained, medium gray, and equigranular. Plagioclase (calcic oligoclase) and alkali feldspar are present in about equal amounts, and quartz is much less abundant. The feldspar grains generally are nearly equant but have ragged surfaces and edges; most of the quartz is interstitial, indicating that it was the last mineral to begin to crystallize. All samples collected from within the map area contain 5 to 10 percent quartz, and, if no other data were available, the rock would be classified as quartz monzonite (fig. 88);

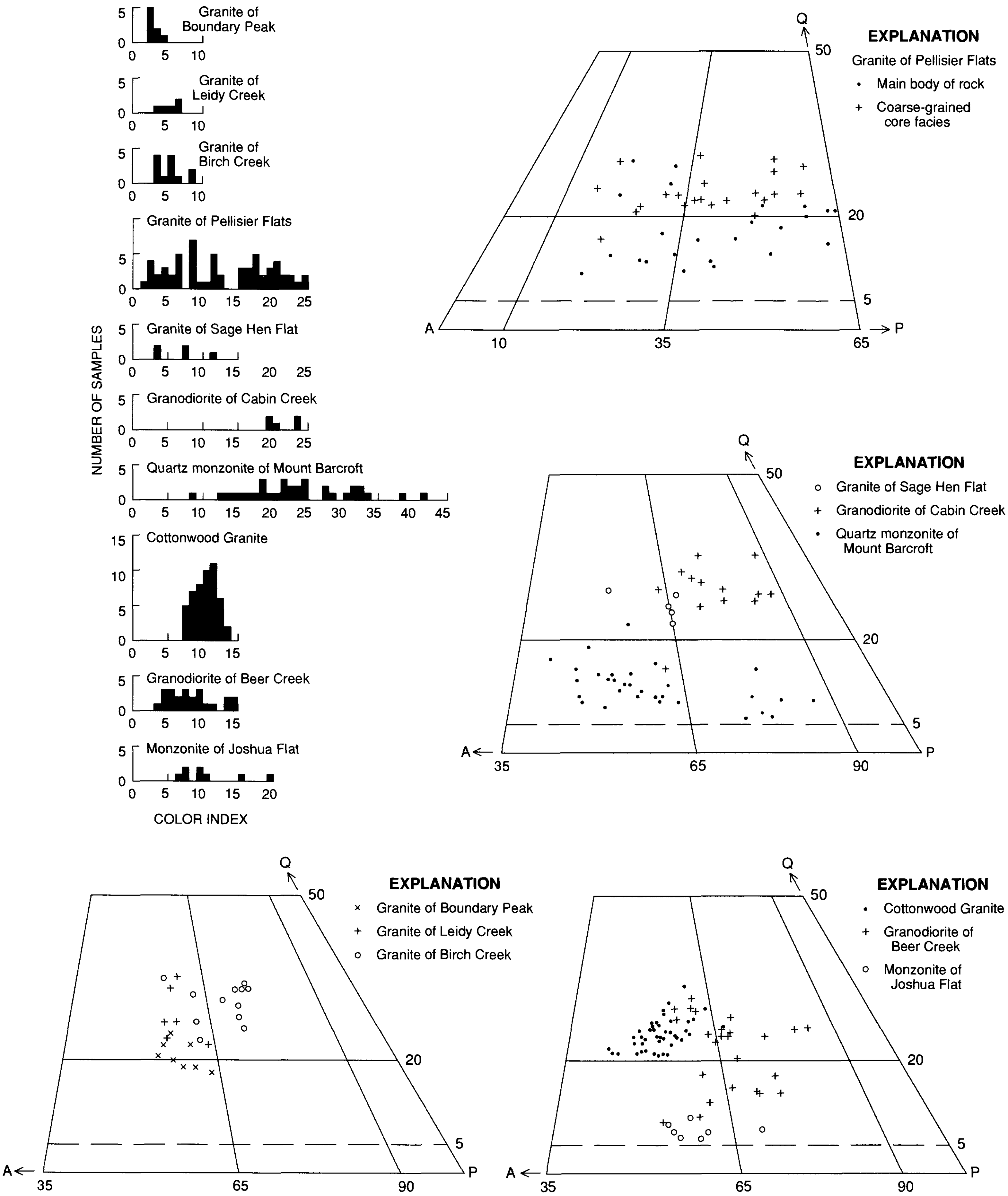


FIGURE 88.—Modes and color indices of the Soldier Pass Intrusive Suite and unassigned Jurassic and Cretaceous intrusive rocks of White and Inyo Mountains. See figure 6 for Q-A-P classification diagram.

however, Sylvester, Miller, and Nelson (1978) determined that many samples from east and south of the map area contain less than 5 percent quartz and thus are monzonite. The color index ranges from 7 to 20. Hornblende is the chief mafic mineral and locally is accompanied by small amounts of biotite. Titanite and opaque minerals are abundant accessory minerals.

The monzonite of Joshua Flat is intruded by the granodiorite of Beer Creek, and east of the map area it intrudes the monzonite of Eureka Valley and the diorite of Marble Canyon (Sylvester, Miller, and Nelson, 1978), which yielded a U-Pb age of 179 Ma. Sylvester, Miller, and Nelson (1978) reported concordant U-Pb ages of 178 and 173 Ma for different fractions of the same sample of the Joshua Flat; Stern and others (1981) and Gillespie (1979) reported concordant U-Pb ages of 167 and 159 Ma, respectively. Samples also yielded K-Ar ages ranging from 175 to 157 Ma on biotite and from 188 to 172 Ma on hornblende (McKee and Nash, 1967). The true age probably is about 170 Ma.

GRANODIORITE OF BEER CREEK (Jbe)

The granodiorite of Beer Creek (pl. 1), formerly part of the quartz monzonite of Beer Creek (Nelson, 1966a; Krauskopf, 1971), is medium gray and fine to medium grained. In outcrop and hand specimen, grain boundaries appear notably ragged. In thin section, the texture appears aplitic. The ratio of quartz to alkali feldspar is highly variable and causes considerable scatter in the plot of modes on a Q-A-P diagram (fig. 88). This scatter is related to geographic location. The rock along the west side of the intrusion, closer to country rocks, has the composition of quartz monzonite and quartz monzodiorite, whereas that farther east has the composition of granite and granodiorite.

In the western facies, both hornblende and biotite are present, much of the quartz is interstitial, and the color index ranges from 8 to 14. In the eastern facies, biotite generally is the only mafic mineral, quartz is evenly distributed as subequant grains, and the color index ranges from 4 to 10. Both facies contain abundant titanite and opaque minerals. The texture and composition of the western facies suggest that it is transitional between the eastern facies and the monzonite of Joshua Flat and that the granodiorite of Beer Creek and the monzonite of Joshua Flat are related.

The interstitial habit of quartz in the western facies indicates that quartz was just beginning to precipitate from the melt phase when crystallization of all other phases was nearing completion, whereas its greater abundance and uniform distribution in subequant grains in the eastern facies indicate that the crystallization of quartz was more advanced there. These considerations suggest that the rock in the western facies crystallized over a higher temperature range than that of the eastern facies.

The granodiorite of Beer Creek intrudes the monzonite of Joshua Flat and is intruded by the Cottonwood Granite. Gillespie (1979) determined a discordant U-Pb zircon age of 161 Ma for the granodiorite of Beer Creek. Several K-Ar ages on hornblende ranging from 180 to 162 Ma and on biotite from 174 to 155 Ma have also been reported (McKee and Nash, 1967; Evernden and Kistler, 1970; Crowder and others, 1973), but some of these determinations undoubtedly were made using samples of the Cottonwood Granite instead of the granodiorite of Beer Creek.

COTTONWOOD GRANITE (Jc)

The Cottonwood Granite (pl. 1) originally was called the Cottonwood Porphyritic Adamellite by Emerson (1966). The name is herein changed to the Cottonwood Granite to reflect the nomenclature of the IUGS classification (fig. 6; Streckeisen, 1973). Its name probably was derived from the upper reaches of Cottonwood Creek north of lat 37° 30' N., which is designated the type locality. Typical Cottonwood Granite is light gray, medium to coarse grained, and has a seriate to megacrystic texture. Alkali feldspar grains range in size from small grains with irregular shapes 3 to 5 mm across to tabular megacrysts as much as 1 cm thick and 3 cm in the longest dimension. In some places, oriented megacrysts impart a foliation to the rock, but foliation generally is weak or absent. Mafic and (or) accessory minerals occur in clusters of small grains and include biotite, hornblende, sphene, apatite, zircon, and opaque minerals. The color index ranges from 8 to 14 (fig. 88).

The Cottonwood Granite intrudes the granodiorite of Beer Creek and is intruded by the Cretaceous granites of McAfee and Indian Garden Creeks. Stern and others (1981) determined a concordant age of 168 Ma and a discordant U-Pb age of 172 Ma on zircon from the Cottonwood Granite (incorrectly reported to be from the granodiorite of Beer Creek).

UNASSIGNED JURASSIC GRANITIC ROCKS OF THE WHITE MOUNTAINS

This group of intrusions includes the quartz monzonite of Mount Barcroft, the granodiorite of Cabin Creek, and the granite of Sage Hen Flat. They are grouped together because isotopic ages indicate that they are all Jurassic but probably younger than the intrusions assigned to the Soldier Pass Intrusive Suite. The quartz monzonite of Mount Barcroft and the granodiorite of Cabin Creek occupy different parts of a fault zone that originally separated metamorphosed volcanic and associated sedimentary rocks of Mesozoic age from sedimentary rocks of Late Proterozoic and early Paleozoic age.

QUARTZ MONZONITE OF MOUNT BARCROFT (Jmb)

The quartz monzonite of Mount Barcroft, equivalent to the granodiorite of Mount Barcroft (Krauskopf, 1971; Crowder and Sheridan, 1972; Crowder and Ross, 1973) and to the Barcroft Granodiorite (Emerson, 1966), occurs in two masses that are separated by the younger granite of McAfee Creek (pl. 1). Its dark-gray color is caused by its high percentage of mafic minerals, as reflected in its color index, which ranges from 9 to 42, and the bluish-gray color of the quartz and the feldspars. The rock is medium grained and generally equigranular but locally contains poikilitic alkali feldspar megacrysts. Hornblende and biotite commonly occur in ragged clusters together with titanite, apatite, and opaque minerals, giving the rock a splotchy appearance. Scattered euhedral crystals of biotite and hornblende are present locally.

The modes show considerable spread on a Q-A-P diagram (fig. 88) and extend across the quartz monzonite and quartz monzodiorite fields. Plagioclase is zoned, mostly in the andesine range, but cores are as calcic as An_{50}. Greenish biotite occurs in aggregates with chlorite and epidote. The presence of relict hornblende indicates that at least parts of these aggregates are secondary and were formed from hornblende. Crowder and Ross (1973) reported only sparse pyroxene in the western part of the larger mass, but Emerson (1966) reported augite in about 20 percent of the samples that he examined from the eastern mass and from the eastern part of the western mass.

The quartz monzonite of Mount Barcroft is intruded only by the Cretaceous granite of McAfee Creek. Gillespie (1979) and Stern and others (1981) determined U-Pb zircon ages of 165 and 161 Ma, respectively. Similar isotopic ages and rock compositions suggest that the quartz monzonite of Mount Barcroft and the Palisade Crest Intrusive Suite (30 km farther south) may be related.

GRANODIORITE OF CABIN CREEK (Jca)

The granodiorite of Cabin Creek (equivalent to the Cabin Granodiorite of Emerson, 1966) occurs in two small masses that originally were continuous but are now separated by Cretaceous granite (pl. 1). Typical rock is medium-gray, medium-grained biotite-hornblende granodiorite, but locally the rock is megacrystic. As in the quartz monzonite of Mount Barcroft, which it resembles, the quartz and feldspars are bluish gray. The color index is relatively high, ranging from 19 to 24, and plagioclase generally is zoned in the range of An_{39-33} (Emerson, 1966). These features suggest a genetic relation between the Mount Barcroft and the Cabin Creek. However, the distinctly higher quartz content of the Cabin Creek (fig. 88) deters interpreting them as having been parts of the same intrusion before the emplacement of the granite of McAfee Creek, which now separates them. It is also difficult to conceive of their being related to each other in a fractionation sequence.

The granodiorite of Cabin Creek is intruded by the Cretaceous granites of McAfee Creek, Leidy Creek, and Boundary Peak. Crowder and others (1973) determined K-Ar ages of 153 Ma on hornblende and of 88 Ma on biotite.

GRANITE OF SAGE HEN FLAT (Jsf)

The granite of Sage Hen Flat (pl. 1), called the Sage Hen Adamellite by Emerson (1966), forms a small zoned pluton that is entirely surrounded by Late Proterozoic and Cambrian strata. The rock that forms the margins of the pluton is fine grained and has an average color index of about 11, whereas the rock in the interior is medium grained and has a color index of about 5 (Emerson, 1966). Both hornblende and biotite are present, but biotite is more abundant. Modes plot close to the boundary between the granodiorite and granite fields on a Q-A-P diagram (fig. 88). Gillespie (1979) reported a U-Pb zircon age of 144 Ma, and Crowder and others (1973) reported K-Ar ages of 141 Ma on hornblende and 133 Ma on biotite.

UNASSIGNED CRETACEOUS GRANITIC ROCKS OF THE WHITE AND INYO MOUNTAINS

A cluster of intrusions of biotite granite crop out in the northern White Mountains, and three isolated plutons farther south (in the southern White Mountains and northern Inyo Mountains) are composed of megacrystic granite and granodiorite. The intrusions of biotite granite in the northern White Mountains include the granites of Pellisier Flats, Boundary Peak, Marble Creek, east of Dyer, Leidy Creek, McAfee Creek, and Indian Garden Creek. The intrusions of megacrystic granite and granodiorite farther south include the granites of Papoose Flat, Birch Creek, and Redding Canyon. Most of these granitic rocks are separated from one another by metamorphic rocks or by alluvium, and the granites of McAfee and Marble Creeks are in fault contact. The granite of Boundary Peak intrudes the granite of Pellisier Flats, but the relative ages of the other Cretaceous granites have not been established.

Contacts of several of the granites dip outward at moderate to gentle angles, especially on their west and southwest sides, indicating that only the upper parts of these intrusions are exposed. These outward dips contrast with the steeply dipping or vertical contacts of most intrusions in the Sierra Nevada and suggest that the White Mountains granitic rocks have been less deeply eroded since they were emplaced than most of the granitoids in the Sierra Nevada.

Barton (1987) showed that the megacrystic granite of Birch Creek is peraluminous, and Emerson (1966) and Crowder and Ross (1973) published chemical analyses that show that the granites of Boundary Peak and of Pellisier Flats also are moderately to strongly peraluminous. Kistler (oral commun., 1988) determined initial $^{87}Sr/^{86}Sr$ for the megacrystic granite of Birch Creek to be 0.71120 and for the granite of Boundary Peak to be 0.70769. Strongly peraluminous compositions and initial $^{87}Sr/^{86}Sr$ greater than 0.710 strongly indicate predominantly crustal sources. Initial $^{87}Sr/^{86}Sr$ has not been determined for the other Cretaceous granites of the White Mountains, but they are compositionally similar to granites farther east and northeast in the Basin and Range province, which, according to DePaolo and Farmer (1984), have initial neodymium and strontium isotopic ratios similar to the ratios for adjacent Precambrian rocks at the time the granites were emplaced.

GRANITE OF PELLISIER FLATS (Kpe)

The granite of Pellisier Flats underlies an area of about 125 km² in the northernmost White Mountains (pl. 1). Typically, it is an equigranular medium-gray rock with a color index that ranges from 2 to 25 (fig. 88; Crowder and Ross, 1973). Biotite is the sole mafic mineral in most samples but is accompanied by hornblende in a few samples. Accessory minerals include opaque minerals, titanite, apatite, and, less commonly, zircon. The mafic and accessory minerals form fuzzy aggregates and shreds and, together with numerous mafic inclusions, give the rock a ragged appearance. Despite the dark color, much of the rock contains abundant alkali feldspar, in places as grains large enough to produce a megacrystic texture. Modes are extremely variable, and on a Q-A-P diagram they plot across the granite, syenogranite, quartz monzonite, and quartz monzodiorite fields (fig. 88). Compared with most of the granites of the Sierra Nevada, the quartz content is low, ranging from 8 to 30 percent and averaging about 18 percent (Crowder and Ross, 1973). A coarser grained core facies in the northern part of the intrusion contains more quartz and fewer mafic minerals than the main body of rock and contains almost no hornblende.

On the west side the granite dips gently to moderately westward under metamorphic rocks, whereas on the east side the contact is variable and dips eastward in some places and westward in others. A zone of mixed metavolcanic and felsic granitic rocks, occupying an area of about 25 km² in the southern part of the intrusion, appears to be a fragmented and possibly foundered remnant of the roof rocks, later intruded by the granite of Pellisier Flats and permeated by solutions that also altered and albitized the southern part of the granite of Pellisier Flats (Crowder and Ross, 1973). The gentle dips of the west contact and the presence of the zone of fragmented roof rocks support the interpretation that present exposures in the southern part of the granite represent a high level in the intrusive body, whereas the core facies in the northern part represents the deepest levels exposed.

The texture of the granite of Pellisier Flats is uniformly xenomorphic, suggesting the rock was fluxed with volatiles during crystallization. Rock in the western margin is pervasively sheared; the shearing extends into adjacent metamorphic rocks but terminates at contacts with the granite of Boundary Peak, showing that the deformation occurred before the granite of Boundary Peak was emplaced. However, the age of the Pellisier Flats remains in doubt. If the D_1 foliations in the metavolcanic rocks along the west side of the White Mountains are older than the ~161-Ma quartz monzonite of Mount Barcroft—as reported by Hanson and others (1987)—and continue into the granite of Pellisier Flats, then the Pellisier Flats is of Jurassic rather than Cretaceous age. A K-Ar age of 157 Ma on biotite from a sample collected at the north end of the White Mountains (Evernden and Kistler, 1970), two K-Ar hornblende ages of 100 and 92 Ma (Crowder and others, 1973), and a discordant U-Pb age of 90 Ma (Stern and others, 1981) have been reported for Pellisier Flats.

GRANITES OF BOUNDARY PEAK (Kwb), MARBLE CREEK (Kwm), EAST OF DYER (Kwed), LEIDY CREEK (Kwl), MCAFEE CREEK (Kwa), AND INDIAN GARDEN CREEK (Kwig)

All these rocks are leucocratic biotite granites that have varying textures but only slightly different compositions (fig. 88). Peraluminous compositions and $^{87}Sr/^{86}Sr$ values greater than 0.71 for the granite of Boundary Peak indicate that, unlike the magmas that produced the Jurassic granitoids, the Cretaceous magmas had a crustal source similar to the Precambrian rocks exposed to the east and northeast. Probably most of or all the Cretaceous granites were emplaced during a brief timespan. Crowder and others (1973) reported probably reset K-Ar biotite ages for these granites ranging from 87 to 73 Ma, and Hanson and others (1987) reported a U-Pb age of 100 Ma for the granite of McAfee Creek.

The granite of Boundary Peak (pl. 1) is medium grained, equigranular, and massive. It is elongate toward the northwest and on the west and southwest sides dips gently beneath the granite of Pellisier Flats. On the northeast side, the relatively straight trace of the contact across rugged topography indicates that it dips steeply or is vertical.

Toward the southeast, along the northeast flank of the White Mountains and beyond a belt of metamorphic rocks is the granite of Marble Creek (pl. 1). The position and shape of this intrusion suggest that it is an extension of the granite of Boundary Peak. However, the granite of Marble Creek is megacrystic and highly variable in texture and composition, unlike the granite of Boundary Peak. East of the granite of Marble Creek and separated from it only by alluvium is the very similar granite east of Dyer (pl. 1), which may be part of the same intrusion. West of the granite of Marble Creek and separated from it by metamorphic rocks is the granite of Leidy Creek (pl. 1); this fine- to medium-grained and not very extensive granite, called the Leidy Adamellite by Emerson (1966), contains 1 to 4 percent biotite and a trace of hornblende.

In fault contact with the south end of the granite of Marble Creek is the coarse-grained, mostly equigranular, felsic granite of McAfee Creek (pl. 1). The northeastern part of this granite, adjacent to the granite of Marble Creek, is weakly megacrystic. The 100-Ma granite of McAfee Creek intrudes several of the Jurassic granitoids. It is also in contact with the granite of Indian Garden Creek to the south, but the relative ages of these two rocks has not been determined.

The granite of Indian Garden Creek (pl. 1) is equigranular, fine- to medium-grained felsic rock (Emerson, 1966) with an average grain size of 1 to 2 mm. The granite clearly intrudes the Jurassic Cottonwood Granite. Crowder and others (1973) reported K-Ar biotite ages of 121 and 95 Ma for the granite of Indian Garden Creek. These ages are distinctly older than those of the other granites in the northern White Mountains except for the 100-Ma U-Pb age for the granite of Marble Creek.

MEGACRYSTIC GRANITES OF PAPOOSE FLAT, BIRCH CREEK, AND REDDING CANYON (Kpp)

The granites of Papoose Flat and of Redding Canyon (pl. 1) form two plutons that consist of coarse-grained, megacrystic rock that closely resembles the Cathedral Peak Granodiorite and Mono Creek Granite. The granite of Birch Creek (pl. 1) is also similar except that it is megacrystic only locally. Although herein called granite, on a Q-A-P diagram modes of the Birch Creek plot in both the granite and granodiorite fields (fig. 88). Much of the rock in the Birch Creek and the Papoose Flat has been deformed and the grain size reduced, but typical undeformed rock is medium to coarse grained. All the rocks contain biotite, and muscovite is present locally. The accessory minerals, apatite, zircon, and opaque minerals, occur in clusters with biotite and secondary epidote and chlorite (Sylvester and Nelson, 1966; Nelson and others, 1978; Sylvester, Oertal, and others, 1978).

Although the Papoose Flat pluton is much larger than the Birch Creek pluton, the two have many features in common. Both were forcibly emplaced in the south ends of major anticlines, and both expanded westward, displacing and disrupting the adjacent strata to the west. On the west side, the margins of both plutons have been ductilely deformed, the grain size reduced, and the rock strongly foliated and lineated. K-Ar biotite ages have been reported for both plutons: 82 and 78 Ma for the Birch Creek (McKee and Nash, 1967) and 80 Ma for the Papoose Flat (Evernden and Kistler, 1970).

The strata adjacent to the Papoose Flat pluton on the west have been stretched and thinned to as little as 10 percent of their original thickness. The granite of Birch Creek was emplaced in the near-vertical east limb of the anticline along a bedding plane within the Reed Dolomite; the pluton expanded toward the northwest, bowing the adjoining strata outward and forming faults and secondary northeast-trending folds, which were then bent into arcuate patterns convex toward the northwest. Both plutons are postulated to have been emplaced as viscous, almost completely crystalline masses (Sylvester and Nelson, 1966; Sylvester, Oertal, and others, 1978). Nelson and Sylvester (1971) suggested that during emplacement of the Birch Creek pluton carbon dioxide released by decarbonation of the Reed Dolomite lowered the solidus temperature of the magma and caused isothermal crystallization.

ULTRAMAFIC ROCKS (um)

Within the map area (pl. 1), ultramafic rocks are confined to the western metamorphic belt, where they form elongate masses along and parallel to the Melones fault zone. These rocks are almost completely serpentinized and consist of serpentine-group minerals, talc, and chlorite. Weathered surfaces generally are reddish, but fresh surfaces range from light to dark green or black. Typically, these rocks are schistose, but in a few places where the rock is massive, bastite pseudomorphs after orthopyroxene can be seen.

DIORITE AND GABBRO (dg)

Small bodies of dark, fine- to coarse-grained rocks are widely distributed within the map area (pl. 1) but are especially abundant on the two sides of the batholith, where they intrude or are spatially associated with metamorphic rocks. The interior of the batholith, occupied by voluminous Cretaceous granitoids, is almost devoid of masses of diorite and gabbro. Mayo (1941) referred to the mafic masses as "basic forerunners," and it appears to be true that most of them are older than the granitic rocks (especially Cretaceous granitic rocks) with which they are in contact. However, few contacts between mafic and granitic rocks were carefully studied during mapping to determine the relative ages of the mafic rocks and the enclosing granitoids. Moreover, relatively hot gabbroic and dioritic magmas can have mobilized previously solidified granitic rock and led to misinterpretations of age relations at contacts.

Studies of volcanic fields suggest that the mafic rocks may be early members of the granitoid suites and are of several different ages corresponding to the ages of the granitoid suites. However, no convincing evidence to support this suggestion has been found. An alternative interpretation is that of Frost and others' (1987), who, on the basis of U-Pb ages in the range of 154 to 150 Ma for three masses of diorite and gabbro in the eastern escarpment of the Sierra Nevada south of Big Pine, proposed that the mafic rocks of the eastern Sierra Nevada are remnants of the plutonic portion of a Late Jurassic magmatic arc that was dismembered during intrusion of the Cretaceous granitoids. They further suggested that the mafic rocks of the western Sierra Nevada may be part of the same arc. Additional support for these suggestions is provided by a U-Pb age of 169 Ma (Stern and others, 1981) for a mass of quartz diorite that lies along the west side of the Pine Creek septum and by contact relations farther north that indicate masses of mafic rock enclosed in the Late Triassic Wheeler Crest Granodiorite are younger than the Wheeler Crest Granodiorite, contrary to Bateman's (1965) original interpretation, and may be Jurassic. No isotopic ages have been determined for the mafic rocks from the western part of the batholith. Obviously, additional isotopic ages and study of contact relations in the field are needed to understand the emplacement history of the mafic rocks.

The compositions of the mafic rocks range from quartz diorite to gabbro, and the color index, grain size, and proportions of constituent minerals are highly variable. The darker rocks (color index 40–60) generally are hornblende gabbro and are composed chiefly of hornblende and plagioclase in which the normative-anorthite content is greater than 50 percent. Cores of monoclinic pyroxene or uralitic amphibole in the hornblende are common; orthorhombic pyroxene is less common. In some rocks, colorless to pale-green uralitic amphibole is rimmed by pale-green amphibole that probably formed at the time of uralitization of pyroxene by reaction with feldspar. Plagioclase grains tend to be euhedral to subhedral and generally are strongly zoned from bytownite cores to calcic oligoclase or sodic andesine rims, but reversals are common. Apatite, titanite, and opaque minerals are common accessories. Secondary minerals include epidote, chlorite, sericite, and sparse serpentine-group minerals.

Many of these diorites and gabbros have unusual textures. Layering is common, and some layered sequences exhibit small-scale unconformities. The layering involves gradation within layers from a preponderance of fine-grained mafic minerals in the apparent base to progressively coarser grained, felsic minerals toward the top. This layering is identical with that in schlieren found in the granitic rocks and undoubtedly has the same origin. Some hornblende gabbros contain prisms of hornblende 2 to 3 cm long. In other gabbros, large hornblende crystals poikilitically enclose small plagioclase grains. A few tabular bodies are composed of a mat of hornblende needles that lie in a plane but which are randomly oriented within the plane. The possibility that some of these textures are the result of recrystallization at the time of emplacement of younger, enclosing granitic rock has not been investigated.

With increasing amounts of biotite, quartz, and alkali feldspar, the hornblende gabbros grade to diorite, quartz diorite, and mafic tonalite. The color index of these rocks generally ranges from 20 to 40. Plagioclase compositions commonly range from An_{50-40} in the inner parts of grains to An_{30-20} in their margins, but mottled cores as calcic as An_{60} have been observed. Although the modes of few samples of the mafic rocks have been determined and field classification is generally dubious, most rocks that have been called gabbro contain less than 50 percent SiO_2 and rocks that have been called diorite or quartz diorite contain more than 50 percent.

MAFIC DIKES

Sporadically distributed swarms of mafic dikes cut many of the granitic rocks. With the exception of the Independence dike swarm (Moore and Hopson, 1961; Chen and Moore, 1979), the oldest and most extensive dike swarm, individual swarms are apparently confined to individual intrusions of tonalite or granodiorite. Mafic dikes are rarely present in felsic rocks and are absent in entire intrusive suites such as the Tuolumne Intrusive Suite. Mafic dikes may be common in one part of an intrusion but absent elsewhere. Generally, the dikes cut across foliation in the host rock and have sharp walls, indicating that they were emplaced either during very late stages of crystallization of the host or after crystallization was completed. However, a few "synplutonic" dikes appear to have been deformed with the enclosing granitic rock during a late stage of consolidation.

The dikes are generally a few centimeters to a few meters thick. Most dikes dip steeply or are vertical, but some dip gently, as along the San Joaquin River below Big Creek where they intrude the Bass Lake Tonalite of the Fine Gold Intrusive Suite. Close spatial and temporal association of the mafic-dike swarms with particular host granitoids suggests that the dikes represent late, somewhat-contaminated intrusions of mantle material of the kind that earlier mixed at depth with lower-crustal material to produce the parent magmas for the granitoids. However, dikes that cut the Bass Lake Tonalite have yielded $^{87}Sr/^{86}Sr$ values that plot well above the 114-Ma isochron for the Bass Lake Tonalite. These ratios are much too high for the dikes to represent mantle material (fig. 18); presumably the dikes have a source in the lower crust.

The dikes are composed of fine-grained diorite or diorite porphyry. Plagioclase, hornblende, and biotite in variable proportions are accompanied by accessory magnetite, apatite, and titanite. Many dikes are net veined with fine-grained felsic material, and in some dikes this material composes a large part of the dike. Crenulate contacts between felsic and mafic zones suggest that two magmas coexisted. In one outcrop, Bateman (1965) observed an intricate mixture of mafic and felsic material where a felsic dike intersects a mafic dike. Nevertheless, the source of the felsic material in most dikes remains problematical. Some and perhaps most of the felsic material in irregular veinlike networks could have been derived by selective melting of host-rock fragments that were incorporated in the mafic dike.

The association of most of the mafic-dike swarms with particular granitoids shows that they are of several different ages. However, the Independence dike swarm (Moore and Hopson, 1961; Chen and Moore, 1979), the oldest, most extensive, and best known swarm, intrudes a variety of pre-Cretaceous granitic and metamorphic rocks and is an exception to the close association of mafic-dike swarms with particular granitoid intrusions. U-Pb ages of 148 Ma for three silicic dikes that Chen and Moore (1979) interpreted to be part of the swarm show that the swarm was emplaced at about the same time as the small masses of diorite and gabbro scattered through the eastern Sierra Nevada for which Frost and others (1987) reported U-Pb ages of 154 to 150 Ma. The Independence dike swarm is about 4 km wide and extends from near Bishop in the eastern Sierra Nevada and White Mountains southward into the Mojave Desert. Dikes of this swarm intrude Triassic and Jurassic granitoids and are cut off by Cretaceous intrusions. Many of the dikes are strongly sheared, especially along their margins. The Lamarck Granodiorite (of the John Muir Intrusive Suite), which is one of the Cretaceous intrusions that truncates mafic dikes of the Independence dike swarm, is intruded by younger mafic dikes, which are cut off, in turn, by the Lake Edison Granodiorite of the same intrusive suite. Other major intrusions of Cretaceous age that were intruded by mafic dikes are the Mount Givens Granodiorite of the John Muir Intrusive Suite and the Bass Lake Tonalite of the Fine Gold Intrusive Suite.

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