DEPARTMENT OF INTERIOR

U.S. GEOLOGICAL SURVEY

Constitution and Genesis of the Central Part of the Sierra Nevada Batholith, California

by

Paul C. Bateman[1/]

This report supplements Open File Report 87-670, Pre-Tertiary bedrock geologic map of the Mariposa 1° by 2° quadrangle, California. It provides expanded descriptions of the features shown on the map and interpretations of their significance and origin. It does not include the map.

Open-File Report 88-382

[1/]Menlo Park, California

Note for New Publications of the U.S. Geological Survey

88-382

Open File Report . Constitution and genesis of the central part of the Sierra Nevada batholith, California, by Paul C. Bateman.

This publication supplements Open File Report 87-670, Pre-Tertiary bedrock geologic map of the Mariposa 1° by 2° quadrangle, California by providing descriptions and interpretations of the rocks and structures shown on the map. Summarizes the results of more than 35 years of study by members of the Geological Survey of the bedrock geology of the central part of the Sierra Nevada batholith between 37° and 38° N. Lat.

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ABSTRACT

The Sierra Nevada batholith comprises the plutonic rocks of Mesozoic age that underlie most of the Sierra Nevada and is a different entity than the Sierra Nevada, a magnificent mountain range. Scattered granitoid 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 speaking) parts of the Sierra Nevada batholith. Nevertheless, all of the plutonic rocks are related in origin.

The batholith lies along the west edge of the Paleozoic North American craton. It was emplaced into 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 the metasedimentary rocks and isotopic ages for metavolcanic rocks indicate that the remnants range in age from Early Cambrian to Early Cretaceous. Within the map area (Mariposa 1° by 2° quadrangle), bedding, cleavage, and the axial surfaces of folds trend about N. 35° W., on the average, parallel with 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 be melanges in part. 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 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, but the possibility that they were transported from distant places has not been disproved. All of the country rocks have been 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 plutonic rocks range in composition from gabbro to leucogranite, but tonalite, granodiorite, and granite are the most common rocks. Most are medium to coarse grained, but some small rock masses are fine-grained. Most have hypidiomorphic granular textures and are equigranular, but some with compositions close to the boundary between granite and granodiorite are megacrystic. 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 granite, all of 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 intruded. 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 to one another. The lithodemes assigned to an intrusive suite commonly, but not invariably, 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 northwesterly direction, parallel with the dominant structures in the country rocks, and some intrusive suites only a few tens of kilometers wide extend northwesterly across the map area and unknown distances beyond its north and south boundaries.

The intrusive suites fall into regular age patterns. The Late Triassic Scheelite Intrusive Suite occupies 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 the 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 they cross in the map area. The locus of plutonism migrated eastward during the Cretaceous at a rate of about 2.7 km/m.y. Nevertheless, Cretaceous plutonism was episodic, occurring within the map area chiefly at about 114, 103, 98, and 90-85 Ma.

Regular compositional changes occur across the batholith and are largely independent of the age of the rocks. In the western foothills, tonalite is the most common rock and is accompanied by biotite granodiorite, granite, and minor trondhjemite. Farther east, along the axis of the Sierra Nevada, hornblende-biotite granodiorite is the most common rock and is accompanied by lesser amounts of biotite granite. In the White and Inyo Mountains, the Jurassic rocks 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 White Mountains. Plagioclase decreases irregularly from a little less than 60 percent in the west to about 30 percent at the east base of the Sierra Nevada, then rises slightly in the White and Inyo Mountains. Total mafic minerals decrease very irregularly from about 14 percent in the west to 4 percent in 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 between 20 and 30 percent across most of the Sierra Nevada and in 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 64 in the west to 55 in the east, reflecting the variations in K_2O, Na_2O, and CaO, but is significantly less sensitive for distinguishing intrusive suites than K_2O/SiO_2 plots, which rise regularly eastward. Among the minor elements, U, Th, Rb, Be, Ta, Ba, and total rare earths increase eastward. The oxidation ratio [mol $(2Fe_2O_3x100)/(2\ Fe_2O_3+FeO)$] 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 it intrudes, some of which have younger isotopic ages than the quartz monzonite of Mount Barcroft. Isotopic ages indicate that metavolcanic rocks are of Late Triassic, Jurassic, and early Early Cretaceous age and that the plutonic rocks are of Late Triassic, Jurassic, late Early Cretaceous, and early Late Cretaceous age. Easily erodable silicic tuffs probably were erupted when the voluminous late Early and early Late Cretaceous intrusions were emplaced and have been eroded away. Decrease in the amount of volcanic detritus present in the Great Valley sequence west of the Sierra Nevada from peak amounts in the Late Jurassic and late Early Cretaceous corresponds in a general way with Late Jurassic and mid-Cretaceous plutonism in the Sierra Nevada and indicates attendent volcanism. However, isotopic ages suggest that peaks of volcanism and plutonism may have alternated rather than occurred concurrently, but additional isotopic dating can alter this apparent relation.

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 skarn 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, and 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, 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 regional compressional tectonic event. but some intrusions were deformed during regional deformations that occurred after they were emplaced. Much of the deformation in country rock remnants within the batholith was caused by forcible intrusion of the plutonic rocks rather than by compressive regional deformation. Displaced wall rocks marginal to rounded intrusions and protrusions clearly indicate ballooning during emplacement, and the presence of foliation and absence of lineation in linear intrusions requires that they also expanded as they were emplaced. Lineations are generally confined to rocks in which the magma from which they crystallized continued to move after crystallization had advanced far enough to sustain a shear stress or to rocks that were sheared after they had solidified.

The source magmas for the granitoids were generated in the lower crust as the 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 antectic magmas in the overlying crustal rocks or mixed and mingled with crustal materials of variable bulk compositions to produce parent magmas with a wide range of compositions and isotopic properties. Some parent magmas were isotopically homogeneous when they rose into the upper crust, but variation of Sr and Nd isotopes within intrusions and intrusive suites indicate that some parent magmas either 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 the exposed plutonic rocks rose upward.

Possible causes of the compositional and isotopic changes that occur eastward across the batholith probably are, (1) increasing thickness of the prebatholithic crust and, consequently, a larger crustal component in the parent magmas, (2) eastward increase in the sedimentary component of the crust, and (3) less depletion of the mantle component in such constituents as potassium, uranium, and thorium with distance from the subduction zone. All of 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 are assumed to grade downward to more mafic rocks in the lower crust. Nevertheless, intrusions with rounded outcrop patterns and tongue-like protrusions from linear bodies, which clearly expanded at relatively shallow depths, must have inward dipping contacts at relatively shallow depths. Mixing processes, which occurred chiefly in the lower crust, and crystal-liquid fractionation account for the diversity of rocks that make up intrusive suites. Crystal-liquid fractionation 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 acretion 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 and rock compositions strongly indicate that, with the probable exception of the Cretaceous granites of the White Mountains, variable amounts of mantle-derived material was incorporated in the magmas. Space for this added material was provided by the uplift and erosion of the roof rocks of the batholith and the upper parts of intrusions during and following intrusion, eruption of easily erodable ash-flow tuff and other volcanic rocks and of air-born ash that was carried eastward by the prevailing winds, and spreading of the walls of the batholith. Geobarometers indicate that 3.5 to 7 km of rock was eroded after the batholith was emplaced. The forcful 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 doubless 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, poisen oak, nettles, chinquapin, and other stinging and spiney plants 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 speaking) 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 and others, 1978). Nevertheless, all of 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 37° and 38° north latitude and 118° and 120° west longitude (fig. 1; pl. 1), and geologic maps of most of these quadrangles have been published at a scale of 1:62,500 (one inch equals about one mile). 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 a few 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

<u>Note</u>: Plate 1 is available as USGS Open File Report 87-670; it is not included with this open file report.

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.

Mapping of 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 many of the skarn-type 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, 1965a). 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 latitude 37° and 38° N. This assignment was designed to utilize 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. This 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 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 and must be acknowledged: 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 of 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 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 Great Basin. 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 continent and 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 (for example: 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 unsolved.

In the central Sierra Nevada, beds, cleavage, the axial surfaces of folds, and the long axes of country rock remnants trend about N. 35° W., on the average. parallel with 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. To the east, the "Sierran trend" terminates east of the Sierra Nevada along the border between California and Nevada, in a large, faulted oroflex (Stewart, 1967; Albers, 1967; Stewart and others, 1968).

The batholith lies along the west margin of the Paleozoic craton. Proterozoic strata have been identified only in the White and northern Inyo Mountains, but Paleozoic strata extend westward into country rock remnants of the eastern Sierra Nevada and are present in the east side of the western metamorphic belt. Mesozoic strata predominate on the western slope of the Sierra Nevada and are exposed in the lower plate of a thrust fault in the White Mountains.

The country rocks of the Sierra Nevada are grouped into sequences, most comprising generally coherent packages of strata, but some of which 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, are considered to have originated in the general area in which they occur, and correlative Paleozoic strata in roof remnants of the eastern Sierra Nevada could 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 (Dickinson, 1970; Hamilton, 1969; Schweickert and Cowan, 1975; Saleeby and others, 1978). However, the places of origin of these strata have not been established, and it is probable that most of them originated nearby in marginal basins and island arcs.

<u>Metamorphism</u>

In the past, steeply dipping beds, folds, cleavages, greenschist and lower amphibolite-facies metamorphism, and other aspects of compressive deformation have been 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 intrusion occurred episodically and overlaps the timespan of deformations invalidates this simple picture. Furthermore, the regional metamorphism may be, at least in part, the result of 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).

Uncertainty as to the areal extent of the Nevadan orogeny has also been increasing. The Middle Jurassic tonalite of Granite Creek in the western Sierra Nevada and Early Cretaceous metavolcanic rocks of the Ritter Range roof pendant and the Goddard septum both have cleavages and lineations parallel with northwest trends considered to be Nevadan. The structures in the tonalite of Granite Creek 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 they 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 Proterosoic 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 span in North America (Nelson, 1962). These strata are overlain by another 2,500 m of Middle Cambrian to Permian strata. In this thick section, unconformities have been reported between the Late Proterozoic Wyman Formation and the Reed Dolomite and between the Sunday Canyon Formation, considered by Miller (1976) to be Devonian, and the Mississippian Perdido Formation (Nelson, 1962, 1966a). 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 oolitic limestone, locally dolomitized. 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 above 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, which varies in thickness between about 200 and 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 archeocyathid limestone are present locally in the lower part of the Harkless. The overlying Saline Valley Formation is highly variable, both in 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 overlying Mule Spring Formation 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 minor interbeds of gray shale.

Monola Formation and Bananza 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 Fomation 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 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, unidivided--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 the tectonically thinned Poleta, Harkless, Saline Valley, and Mule Spring Formations.

Al Rose, Badger Flat, Barrel Spring, Johnson Spring, Palemtto, 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 minor limestone interbeds. Upward, thin-bedded shaly limestone is progressively more abundant.

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 roof pendant 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.

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 younging, and the strata of the eastern block dip steeply both east and west. Rinehart and Ross (1964) have 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) have 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, have concluded that the contiguous, superjacent Koip sequence has been tectonically thinned about 50 percent. They suggest that other sections of stratified rocks in the Sierra Nevada have been similarly thinned. Both interpretations may be correct--the strata may have been concurrently 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 existence of the 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 with 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 would very likely be 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 here are mainly from their studies. Because their studies are still incomplete, modifications can be expected.

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 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 and within the Paleozoic sedimentary strata adjacent to the Koip, 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, 1965a; 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 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 tectonic 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 cross-beds, 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 Tobisch and others (1977) (see also Fiske and Tobisch, 1978) have carried out strain studies that indicate the original thickness approached 11 km.

Tabular or broadly lens-shaped depositional units with essentially planar tops and bottoms, and an 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. Pectenoid 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 equalled the rate of subsidence throughout the entire period of deposition. Shallow subaqueous (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 (unpublished data of R.S. Fiske and O.T. Tobisch, written commun., 1985). These ages suggest that the eastern part of the section is Late Triassic to Early Jurassic and that the western part is Middle and Late Jurassic. This simple picture is confused by the earlier discovery of a pectenoid pelecypod of the genus Weyla, known only from the Early Jurassic, in the eastern part of the zone that yielded Middle and Late Jurassic ages (Rinehart and others, 1959). If the U-Pb ages are correct, the possibility that the pectenoid was redeposited or misidentified must be considered.

Kistler and Swanson (1981) have also divided the Koip sequence into younger and older successions separated by an unconformity, but their break would have occurred earlier than the horizon that separates the two groups of U-Pb ages. They argue that 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 westerly 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 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 on the older strata of the Koip are correct, the older part of the Koip may be cogenetic with the Scheelite Intrusive Suite, which has an isotopic age of ~210 Ma (Stern and others, 1981; Chen and Moore, 1982; Kistler, 1966b). Isotopic dating has not revealed any granitoids that were emplaced during the extended timespan when the younger parts of the Koip 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 minor felsic lava flows (Tobisch and others, 1986). Beds dip steeply west, 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 on rocks from the Koip. 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 occupy 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, stongly deformed phyllitic schist, lapilli tuff and tuff-breccia, sills, dikes, and lava flows. The strata in the west side of the Goddard septum are mostly water lain, whereas those in the east side are mostly subaerial deposits (Tobisch and others, 1986). All the strata dip steeply west and face west, and are structurally conformable with the Jurassic strata that occupy the core of the septum. The strata in the west side of the septum may stratigraphically overly the Jurassic strata in the core of the septum, but the strata in the east side are in fault contact with the Jurassic strata. Three U-Pb ages on the strata of the east side of the septum are 144, 141, and 140 Ma, and two ages on strata of the west side are 137 and 136 Ma (Tobisch and others, 1986).

Dana sequence

The Dana sequence occupies 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 syncline. Kistler and Swanson (1981) report a Rb-Sr whole-rock age of 118 ± 11 Ma based on an isochron drawn on 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 may be 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 ± 2.2 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 they call 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 with 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 the unassigned leucogranite of Graveyard Peak, also suggest a cogenetic relation with the Minarets sequence.

Metavolcanic rocks, undivided

The U-shaped mass of metavolcanic rocks 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 and (or) the dominantly pyroclastic strata of Early Cretaceous age.

In mapping the White Mountains, Crowder and Sheridan (1972) and Krauskopf (1971) 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 of 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. Pyroclastic textures and flattened pumice fragments are common among the felsic rocks. The dominantly metasedimentary succession contains lithologies as diverse as schist and phyllite, disrupted limestone lenses, metavolcanic siltstone and sandstone (in part calcareous) and coarse conglomerate. 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 of 137 Ma on a nearby sample of hypabyssal rock. Hanson and others (1987) also report a U-Pb age of 100 Ma on the granite of McAfee Creek, which intrudes the metavolcanic succession on the west. All of these ages are younger than published U-Pb ages of 161 and 165 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 (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 metavolcanic 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 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 Thelan, 1905; Christensen, 1963; Saleeby and others, 1978).

The Late Triassic and (or) Early Jurassic age designation assigned to the Kings sequence rests 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, 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 lithologies in the Yokohl Valley roof pendant, as described by Saleeby and others (1978), 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 Strawberry mine area and from the Potter Pass septum (fig. 2). The first fossil found at the Strawberry mine area was a bivalve, possibly *Inoceramus pseudomytiloides* of Early Jurassic age (Nokleberg, 1981). However, in 1982, R.S. Fiske and O.T. Tobisch collected an extensive fauna from the same general area, which has been only partially studied. Preliminary determination of about 25 percent of the collection suggests 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, 1981), and the May Lake septum (Bateman and others, 1983b). The strata in the Dinkey Creek roof pendant are folded and thrust faulted; nevertheless, Kistler and Bateman (1966) determined the stratigraphic succession to be, from oldest to youngest, (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, 1981). The May Lake septum (fig. 2) consists of light-gray quartzite, diverse calc-silicate rocks, and marble (Bateman and others, 1983b).

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. Although no contacts between these sequences have been observed within the map area, Saleeby and others (1978) report 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 could have been interlayered with the Kings sequence where both sequences were present in the same area. Saleeby and others (1978) picture 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, somewhat resembles quartzite in the Kings sequence, except that cross-bedding 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, obscure structures, and 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 lithologies 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 of 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 that lie west of the Melones fault zone, which Duffield and Sharp (1975) have shown to be part of a melange.

Schweickert and others (1977) proposed that the name Calaveras Complex be substituted for Calaveras Formation between latitudes 37°30' and 38°45' N. and for isolated outcrops as far south as 36° N. They excluded rocks west of the Melones fault zone, which had previously been included in the Calaveras, and did not discuss rocks currently assigned to the Calaveras between 39°30' and 40° N. They identified four units (from east to west--volcanic, argillite, chert, and quartzite), which they assumed, following Clark (1964), to range in age from Carboniferous or Permian in the west to Triassic and Jurassic in the east. However, the Calaveras Complex of Schweickert and others (1977) had a short life. Shortly after publication of the report by Schweickert and others (1977), Schweickert (1977) traced the Calaveras-Shoofly 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. More recently, Schweickert and others (1984) have deleted the two units that 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) has been reduced to a small part of its original extent, Bateman and others (1985) did not use "Calaveras" and instead used informal names for the four units that correspond approximately with those that Schweickert and others (1977) used. These 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 sparce 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 younging of units may be at least partly 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(?) arc-volcanic rocks. He suggested that the strata in the south end of the western metamorphic belt east of the Melones fault zone were deposited at the distal end of a westward-flowing drainage system. A marginal 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 on 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 probably Paleozoic (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-Shoofly 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 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 spheroids visible in thin section, which suggest ghosts of radiolaria, and $\delta^{18}O$ values that range from 22.2 to 24.1 in five samples (Bateman and others, 1985).

Fossils have been collected from two limestone beds within this unit. Samples collected from an extensive bed of limestone that crosses the Merced River in the central part of the unit were found to contain conodonts of Early Triassic (Griesbachian and Smithian) age (Bateman and others, 1985). This bed is interstratified with rhythmically bedded chert and phyllite and is not an olistolith. The other reported fossils, collected much earlier by H.W. Turner (1893a) 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 in the crest of the ridge north of Hite Cove, have been unsuccessful.

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

Phyllite of Briceburg--The phyllite of Briceburg includes phyllite, argillite, metagraywacke, sparse chert, and limestone lenses 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 Cement Quarry limestone. All of these blocks are probably olistoliths whose source may have been a limestone bed in the chert and phyllite of Hite Cove, probably the limestone 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. They probably are comagmatic with the greenstone of Bullion Mountain; if so, they indicate that 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 thicknesses of as much as a few hundred meters. 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 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 are probably 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 the western part, adjacent to the Melones fault zone. However, samples collected from the northeastern margin along the Merced River where the rocks consist chiefly of metamorphosed tuff, breccia, and possibly lava flows and exhibit compositional diversity have 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 eastern part of the unit is Early Jurassic. This age is comparable to an Early Jurassic or Triassic age inferred by Morgan and Stern, 1977) for the Penon 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 in a single unit on the map of the Mariposa 1° by 2° quadrangle (pl. 1), but the greenstone of Bullion Mountain retains its identity (Bateman and Busacca, 1982; Bateman and others, 1982). Quartzite like that in the quartzite 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 a little 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 strats west of the Melones Fault Zone

The strata west of the Melones fault zone consist of the Mariposa Formation adjacent to the Melones fault zone and a series of metasedimentary and metavolcanic units to the west and farther 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 of 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 these strata are successively younger toward the northeast, the Mariposa being the youngest unit.

According to this interpretation, the oldest unit, in the western side of the southern prong of the western metamorphic belt, is chlorite schist. This unit 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 probably were mafic volcanic rocks (probably tuffaceous), argillaceous sediment, silt, intermediate to silicic volcanic rocks and hypabyssal intrusions, and silt or fine sand, 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 locally conglomerate. It has yielded abundant fossils of late Oxfordian and early Kimmeridgian (Late Jurassic) age (Clark, 1964). None of the other units has yielded fossils, but I assume them to be Late Jurassic because they are thin and 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 largely abandoned, but a wholly satisfactory alternative interpretion of their stratigraphic and structural relations has not yet been developed.

All of the rocks are strongly deformed, and many have been deformed more than once. The structures are increasingly obscure and may be more complex westward across the Sierra Nevada. 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 folded and faulted complexly, but the presence of fossils and distinctive stratigraphic units makes 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 within the western part of the batholith and the western metamorphic belt, structures are extremely difficult to identify because of poor exposures, a lack of distinctive stratigraphic units, a 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 of the structures have been explained in terms of regional compression and crustal shortening, but the role of batholith emplacement was also important. Episodic emplacement ot the plutonic rocks, which expanded as they were emplaced, undoubtedly accounts for much of the deformation of the wall rocks of the batholith and of 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 the presence of competent and distinctive marker beds generally makes 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, thicken and thin irregularly. Thicknesses in the cores of folds may be several times those in limbs.

The largest folds 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 Springs 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 and the granodiorite of Beer Creek. The southern anticlinorium is in the northern part of the Inyo Mountains just south of Deep Springs 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. 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 Springs 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 competent Campito Formation has squeezed the incompetent shale member (Montenegro 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 that either they 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 units of the Middle Jurassic Soldier Pass Intrusive Suite (p. 160-161).

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. The fault most likely to underlie the area is 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) have 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 existence of this fault is uncertain. Nelson (1966a), who first mapped the area, recognized only normal faults and subsidiary landslides, and Dunne and Guliver (1978) have 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 Crowder and Ross (1973) have suggested that Proterozoic and early Paleozoic strata have been thrust westward over Mesozoic metavolcanic strata along the "Barcroft structural break." The metavolcanic rocks form an irregular 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 metavolcanics 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 metavolcanics 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 marginal 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 no more than 137 Ma and before the granite of McAfee Creek intruded the strata 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, eliminating about 1,500 m of strata, along the east side probably occurred during the same interval. The map pattern suggests that the thrusting that carried the Paleozoic and Proterozic 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 the shearing in the western part 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 metavolcanics, are intruded by the quartz monzonite of Mount Barcroft and are therefore older. If these D_1 foliations and lineation 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.

Across Owens Valley, small remnants of metavolcanic rocks are present southwest of Keough Hot Springs. These remnants are enclosed in granitoids, but their distribution suggests that before the granitoids were emplaced they may have been in contact with Paleozoic strata to the west in the Bishop Creek and Pine Creek roof pendants. Although other interpretations can be made, it is possible that the thrust underlying the White and northern Inyo Mountains, assuming it exists, passed over these metavolcanic remnants and beneath the Paleozoic strata of the roof pendants.

Structures in remnants of country rocks of the eastern Sierra Nevada

Although the structures in roof pendants and septa of the eastern and high Sierra Nevada are better understood than those in the western metamorphic belt and adjacent remnants of country rocks, many problems 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 the 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. Folds along northwest Nevadan trends in the late Paleozoic strata of the Mount Morrison roof pendant and the Pine Creek septum were formed before they were intruded 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. Angular unconformities between the Koip and Dana sequences and between the Koip and Minarets sequences, and the westward tilt of the west block of the Minarets sequence, indicate that the present steep dips resulted from repeated tilting, the last of which occurred after deposition of the Minarets sequence, about 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 expansion that accompanied emplacement of the batholith. Most of the tilting would have occurred after deposition of the Early Cretacous metavolcanic rocks and before eruption of the ~100 Ma Minarets caldera, but renewed tilting must have occurred later--after eruption of the Minarets sequence and before intrusion of the mid-Cretaceous (~98 Ma) granodiorite of Jackass Lakes--then was followed by minor faulting after 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 to form (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 metavolcanics 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 these 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

Most of the remnants of metamorphic rocks within the map area that have been assigned to the Kings sequence are small, and structural patterns have been determined only in the Dinkey Creek and Strawberry Mine area (Kistler and Bateman, 1966; Nokleberg, 1981). 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° N. 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 parallel 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 with first-fold axial surfaces. The first-folds are spatially associated 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 southeast-directed thrusting, which overturned earlier open folds.

The strata in the Strawberry mine area are chiefly calc-silicate hornfels and marble. Nokleberg (1981) assumed them to be of Early Jurassic age, but fossils collected more recently suggest a Late Jurassic or Early Cretaceous age (see p. 28-29). They have been intruded by metaigneous rocks and are surrounded by metavolcanic rocks of the mid-Cretaceous Minarets sequence. Nokleberg (1981) recognized four generations of folds, but the fourth is poorly represented. The first three fold systems are similar in their orientations to those in the Dinkey Creek roof pendant. The first folds affect 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 (1981) interpreted them to have been moderately appressed and to have 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. Nokleberg assigned the next younger generation of folds, which trend N. 25° W. to the middle Cretaceous because they are cut by the "early Late" Cretaceous (actually mid-Cretaceous) granodiorite of Jackass Lakes. He assigned a middle or Late Cretaceous age to the youngest folds, which trend 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. Figure 5*A* shows 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. Figure 5*B* shows successively younger strata thrust eastward over older strata, then modified by backfolding and west-directed folds and faults. This interpretation agrees with an interpretation by Moores and Day (1984) of relations in the northern Sierra Nevada; however, the early eastward overthrusting is not required to explain the relations within the map area. Figure 5*C* shows younger units thrust eastward beneath older units similar to the pattern in accretionary prisms. 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 phyllite and chert of Hite cove, which appear to constitute a relatively unbroken package.

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 the Melones may be a major zone of 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 to the west and a phyllite-greenschist belt to 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 Patterson (written commun., 1985) that indicate changes in strain, structures, lithologies, and metamorphic grade across the fault 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, a contact that includes discontinuous lenticular masses of serpentinite. Serpentinite lenses are also present on either side of this contact for distances of as much as 2 km, suggesting that the zone affected by faulting is at least several kilometers wide. Forcible intrusion of the Bass Lake Tonalite has bent the south end of the fault zone, together with a serpentinite lens that lies along the contact between the Mariposa Formation and the greenstone of Bullion Mountain, sharply west for about 3 km (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 bend toward the south and pass into the Central Valley between the Adobe Hill and Tick-Tack-Toe roof pendants. Bateman and others (1983) suggested that the leucotonalite of Ward Mountain (Ward Mountain Trondjhemite of this report) occupies the ancient trace of the fault zone and that renewed activity at depth after emplacement of the Bass Lake Tonalite may in some way 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, it may lie between the quartzite of Pilot Ridge and the phyllite and chert of Hite Cove, but it also could lie farther east. The principal evidence that the thrust coincides with the contact between the quartzite of Pilot Ridge and the phyllite and chert of Hite Cove is the Paleozoic isotopic ages on orthogneisses that are reported to intrude strata west and northwest of the map area 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 Scott 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 trends about N. 85° W., whereas east of the line it trends 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; emplacement of the batholith caused more localized deformation during the Mesozoic; and tectonic movements of local extent or lesser intensity occurred during the Late Cretaceous and Cenozoic. The most reliable means of determining the time when a 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 an 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 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 and Bateman, 1966; Kistler, 1966b; Russell and Nokleberg, 1977; Brook, 1977; Nokleberg and Kistler, 1980). However, this criterion must be used with extreme caution because the attitudes of fold axes and axial surfaces of the same age commonly 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

The presence of the Antler orogeny within the map area is suggested by (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 (Sonaman?) orogeny

An orogeny in the Early Triassic, which may be the Early Triassic and 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 probably were formed during this deformation but could have been formed during the Late Triassic deformation.

Late Triassic deformation

Kistler and Swanson (1981) proposed an unconformity between their younger and older successions of the Koip sequence. According to them, the older Koip 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 Koip. 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 in doubt because of disagreement between U-Pb and Rb-Sr ages on the Koip sequence. Nokleberg and Kistler (1980) attributed folding in Calaveras strata and in the strata of the Boyden Cave and 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 that were intruded earlier into 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 make up the base 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 beginning of 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) have summarized data that bear on the age of the Nevadan orogeny, including the isotopic ages of granitic rocks that were emplaced before and involved in the Nevadan orogeny and of rocks 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 on the tonalite of Granite Creek, which was deformed and lineated, presumably during the Nevadan orogeny, and 151 Ma on the contiguous undeformed granite of Woods Ridge. The absence of evidence of deformation during the time when the Jurassic and Early Cretaceous metavolcanic strata of the High Sierra were being deposited suggests that the Nevadan orogeny was limited to the western foothills.

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 were deposited. An accurate isotopic age for the Dana sequence, which also rests unconformably on the Koip, would narrow the timespan during which the westward tilting could have occurred. Tobisch and others (1986) suggested that the tilting occurred by rotation on listric faults and reflects regional extension rather than shortening during a compressional event.

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 old 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 in the granite of Shuteye Peak and 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?) deformations

Shear zones similar to those within the grandiorite 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 trends 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 are uplift and westward tilting of the Sierra Nevada, which continues to the present.

HIERARCHIAL ORGANIZATION OF GRANITIC UNITS

The Sierra Nevada batholith is composed of hundreds of separate granitic plutons that must be assigned positions within a hierarchial system for the larger problems of the batholith to be dealt with effectively. 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 here 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 accord 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, 1965a).

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. A few plutons within the map area are named after prominent features within them, but most plutons have not been named. If a pluton assigned to a lithodeme contains the type locality of the lithodeme, it is given the same name as the lithodeme. For example, the pluton of Ward Mountain Trondhjemite that contains Ward Mountain is called the Ward Mountain pluton, whereas the second pluton of this lithodeme is called the Experimental Range pluton (Bateman and Busacca, 1983).

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 margins of a pluton to grade inward to more leucocratic rock. However, some 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 cross-cutting relations of dikes, 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 the relative ages of a few were not determined because they 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 slopewash 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 to originally have been parts of a single body; nevertheless, they are now considered as separate plutons.

In this report, lithologic designations for the intrusive rock units are in accord with the classification (slightly modified) recommended by the International Union of Geological Sciences (fig. 6) (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. Because some units are small and unimportant or because only part of the unit falls within the area being mapped and its full dimension and average composition were 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; 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.

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 Stratigraphic Commission in 1983 as the equivalent among plutonic and high-grade metamorphic rocks of "group" among stratified rocks. The formal name of an intrusive suite consists of a geographic name followed by "Intrusive Suite" (capitalized)--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 intrusive suites involves considerable uncertainty, and reassignments can be expected in the future. 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 the 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 a 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 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 to one another to unambiguously demonstrate their consanguinity. Some of the units that compose these less 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 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 of them 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 the 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 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, entirely surrounded by plutonic rock, that is presumed to be a downward projection of roof rocks that before erosion overlaid the plutonic rocks. Roof pendants may be enclosed within a single intrusion or be bounded by two or more intrusions. Usually they are large and underlie several square kilometers, but size is not an essential requirement.

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. A septum may also be a roof pendant, but it is not necessarily a roof pendant. "Screen" is an equivalent term.

Inclusion: A mass of rock that is surrounded by rock of the same intrusion and is assumed to to have been covered by the the same 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 (1 mile) apart, but no attempt was made to follow a rigid pattern, which the terrane would not allow. 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 from the map area between 1953 and 1983 has been published (Bateman and others, 1984a), and only chemical analyses needed to support interpretations are given in this report.

Because of the limited capacity of the analytical laboratories of the U.S. Geological Survey 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 increased the capacity of the chemical laboratories and reduced costs. Some chemical analyses reported 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 given 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, all modes in reports 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 to the total mafic content as determined on stained slabs.

The number of point counts made on the stained slabs would be 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; gradual changes over large areas indicate systematic distribution patterns.

The modal abundances of quartz, alkali feldspar, and plagioclase are plotted on Q-A-P (quartz-alkali feldspar-plagioclase) diagrams on which the fields of the different granitic rocks are shown, according to the IUGS (International Union of Geological Sciences) classification (Streckeisen, 1973), but with two minor modifications (fig. 6). The modifications are as follows: (1) The upper limit for quartz in granitic rocks is lowered from 60 to 50 percent because not a single mode from the central Sierra Nevada plots between 60 and 50 percent, and (2) The monzogranite subfield of the granite field of the IUGS classification is shown simply as the granite field because most rocks commonly called granites fall in that field. Rocks that plot in the syenogranite subfield of the IUGS classsification continue to be referred to by that name. For most rock units, the color index (sum of volume percentages of mafic and accessory minerals) is shown by histograms, but where the volume percent of hornblende and biotite was determined separately, their abundance is 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 ages are needed to establish the ages of all of the rocks. The U-Pb ages are generally more reliable than K-Ar ages because the blocking temperature 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 anomolous 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 $\pm$ 2 percent (Stern and others, 1981). R.W. Kistler (written commun., 1984-1987) has determined a few Rb-Sr isochrons on both plutonic and metavolcanic rocks. The ages determined on plutonic rocks by this method appear to be generally reliable, but the ages determined on volcanic rocks have been variable. Redistribution of Rb and Sr by volatiles during or after consolidation may explain some anomalous Rb-Sr ages in 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 can be reduced from the age of crystallization because of slow cooling to the blocking temperature or because of reheating by younger intrusions. Some K-Ar ages on hornblende are inexplicably older than U-Pb ages on 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, the Bass Lake Tonalite, which is an I-type granitoid, belongs to the magnetite series, and it is possible that further study will 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 a sedimentary source, but the criteria that distinguish these two types of granitoids are complex and have been revised 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: I-type granitoids have Na_2O >2.2 percent, mol $Al_2O_3/(Na_2O+L_2O+CaO)<1.1$, C.I.P.W. normative diopside or <1.1 normative corrundum, a broad spectrum of compositions, and regular inter-element variations (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) redefined S-types narrowly to agree with the properties of S-type granites in the Lachlan Fold Belt of southeastern Australia where they were first identified. According to these restrictive criteria, S-type granitoids are very uncommon rocks, and none are present within the Mariposa 1° by 2° quadrangle (pl. 1).

The distinction between magnetite- and ilmenite-series granitoids is simpler than the distinction between I- and S-type granitoids. Magnetite-series granitoids contain more than 0.1 volume percent magnetite, and ilmenite series granitoids contain less than that amount. Ishihara (1977) interprets ilmenite-series granitoids to have originated in the middle to lower continental crust where they incorporated carbon-bearing materials that reduced Fe^3 to FeO and magnetite-series granitoids to have been generated in a deep level of the lower crust or upper mantle and where little or no carbonaceous material was present. This interpretation doesn't provide a satisfactory explanation for the low magnetite content of the Bass Lake Tonalite, which is less likely than most Sierran magmas to have large amounts of carbonaceous sediments among its source materials.

PHYSICAL FEATURES OF THE GRANITIC ROCKS

Minerals

The minerals of Sierran granitic rocks are typical of granitoids in the compositionsl 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 Trondjemite, but may be secondary. Cordierite, andalusite, sillimanite, and garnet have not been reported, except locally in the granite of Dinkey Dome, an epizonal pluton very nearly water saturated that 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. Under polarized light, the larger grains in many rocks can be seen to extinguish irregularly or consist of a mosaic of diversely oriented components. Irregular extinction and mosaics reflect late- or post-crystallization strain; quartz responds to strain more readily than other minerals and in many rocks is the only indicator of strain. Extinction generally is either undulatory or by sharply defined polygonal areas that are visible near extinction.

Alkali feldspar

Alkali feldspar is white, light gray, or pinkish in hand specimen and generally forms anhedral grains. In low-potassium rocks of the Fine Gold Intrusive Suite (quartz diorite, tonalite, biotite granodiorite, and trondhjemite) most of the alkali feldspar is interstitial, whereas in rocks with higher potasium content some of the alkali feldspar may form euhedral to subhedral megacrysts. In thin section, alkali feldspar generally has quadrille structure (grid twinning, gridiron structure, or grating structure), but in some rocks the quadrille structure is inconspicuous or absent. Most alkali feldspar is perthitic, and plagioclase grains contiguous to alkali feldspar are bordered with rims of 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 that would be required for the alkali feldspar to coexist in equilibrium with the plagioclase in the magma (Bateman and Nokleberg, 1978; Seck, 1971; Stormer, 1975). Similar results have been obtained (Noyes and others, 1983a, b) in the granodiorites of Red Lake and Eagle Peak. Clearly, the alkali feldspar in the Mount Givens Granodiorite and the granodiorites of Red Lake and Eagle Peak, and by inference in other granitic rocks as well, has reequilibrated (or partially reequilibrated) at subsolidus temperatures, and albite has been expelled to form perthitic intergrowths and albitic rims on plagioclase in accordance with the observations of 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 a centimeter 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 outcrop dimensions 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 is a common dimension in the outer part of the intrusion, and 1 by 1.5 cm is common in the inner part where megacrysts are scarse. 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 commonly selected growth zones that parallel the crystal faces 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 and 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 (Bateman and Chappell, 1979; Reesor, 1958; Bateman and Wones, 1972a, b; Wagener, 1965; Chappell and White, 1976). Lockwood (1975) determined that the abundance of megacrysts decreases inward in the southern part of the Mono Creek Granite and is not related to altitude (fig. 8). In the Tuolumne Intrusive Suite, these changes take place in the absence of significant changes in the total amount of akali feldspar in the rocks.

Figure 9A is a plot of all of the modes of megacryst-bearing granites and granodiorites within the map area (pl. 1). 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 areas of the Cathedral Peak Granodiorite, show much less scatter and plot mostly within the 10 percent contour for all the modes (fig. 9B).

Megacrysts typically contain more barium than groundmass alkali feldspar, and cores of megacrysts are more enriched in barium than margins (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. Evidence that they are magmatic is that they have been carried in dikes, are oriented parallel with 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 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 plagioclase compositions in megacrystic rocks range from An_{23-39} to An_{9-23}. An_{23} is both the minimum high value and the maximum low value, and consequently the only composition common to all the plagioclase crystals in rocks that contain megacrysts; An_{23} 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 enclosed in megacrysts would provide data needed to evaluate this speculation.

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 their growth. Mafic minerals and plagioclase probably were abundant in the marginal magma of the Tuolumne Intrusive Suite when alkali feldspar first began to crystallize, but if the marginal rocks were formed by the 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 interrrupted by discontinuities and (or) reversals. Thin oscillations are common. Zoning reflects changes in the temperature, pressure, composition, or dissolved-volatile content of the melt phase during crystallization. The anorthite content generally 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. Common ranges in granodiorite are from An_{45-30} to An_{25-22} and in granite from An_{20} to An_{10}. Marginal 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, but 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 clusters that include 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 trains of such crystals that terminate at mafic inclusions, suggest that the mafic inclusions supplied material for the formation of the euhedral hornblende crystals.

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

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

In studying the variations within the compositionally zoned bulge at the north end of the Mount Givens Granodiorite, Nokleberg made 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 margin 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 increase of Fe (FeO). The TiO_2 and MnO concentration in hornblende also increases 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 biotite. Alteration of hornblende to biotite 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 observation seems a viable explanation for the aggregates of minerals in mafic inclusions and small clots of anhedral minerals, but discrete subhedral hornblende crystals probably precipitated directly from the melt phase. Those that have pyroxene cores appear to have grown with only minor reaction with the pyroxene. Burnham (1979) states that although a 3 percent water content would allow hornblende to coexist stably in 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 kb water pressure hornblende forms below 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 flourine would significantly 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, presumably 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 distinguished in most rocks, but microprobe studies show that ilmenite is present in the Mount Givens Granodiorite and absent in the Tuolumne Intrusive Suite where titanite is abundant. Czamanske and Mahalik (1972) and Czamanske and Wones (1973) suggested several reactions in which an increase of $f_{O}2$ can cause increase in the ratio of magnetite to ilmenite. Titanite that occurs as discrete double wedges appears to be primary, but the titanite in 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 hypidiomorphic granular, but some are fine grained, and a few coarse grained. Undulatory extinction in quartz is minor and fine-grained granoblastic zones between grains are absent. 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. The clots may grade in size to typical mafic inclusions.

In medium-grained rocks, the accessory minerals form the smallest grains, hornblende and biotite the next smallest, and the felsic minerals the largest. 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 mm across, but in some rocks they are as much as 1 cm across. Plagioclase grains are usually 2 to 3 mm long and are a little smaller than quartz and alkali-feldspar grains. Porphyries in which the same minerals occur in a fine-grained groundmass and as small phenocrysts are scarce; generally they form small plutons associated with metavolcanic rocks. Some medium- to coarse-grained granites and granodiorites are megacrystic and contain large crystals of alkali feldspar. Miarolitic cavities are uncommon but are present in a few felsic rocks, especially fine-grained varieties that crystallized late in an intrusive suite.

Textures in deformed granitoids

The granitoids develop a gneissic foliation and commonly a lineation as well, as strain increases at elevated temperatures. Strain can result from flow during late stages of crystallization or from regional stress. Different minerals respond differently during ductile deformation. Quartz is the most sensitive mineral to strain and readily breaks down to form a microcrystalline mortar between the other minerals in the rock, expecially 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 resistent, and in moderately deformed rocks porphyroclasts commonly survive, although different parts of crystals may be optically discontinuous across sharp lines as the result of fracturing and rotation. Interstitial alkali feldspar appears to fail with plagioclase and forms a granoblastic mortar with quartz and plagioclase with moderate strain. However, equant grains and magacrysts are resistent 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 needle-like aggregates of quartz and feldspar, 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: biotite+plagioclase (An_{45})+quartz=hornblende+titanite+plagioclase (An_{35})+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 the result of differential flow velocities in the magma as it crystalized. 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 lenticular mafic inclusions. The mineral foliation generally parallels that shown by mafic inclusions but in places is less regular and varies nonsystematically a few degrees from the foliation shown by mafic inclusions. Magmatic lineation generally is difficult to identify, but in some granitoids oriented hornblende prisms define a lineation in the plane of mineral foliation. 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.

In general, foliation is strongest close to external intrusive contacts with older rocks, which it approximately parallels. 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 plutons, 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 pluton (fig. 11).

Balk (1937) has summarized and extended the pioneering work of Hans 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, and Mackin (1947) has identified three types of nonuniform flow that can produce foliation (fig. 12). Mackin terms these three types of flow as 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, magma introduced into the central part of a chamber moves outward toward the chamber walls with diminishing velocity. Because trailing edges and ends of crystals and mafic inclusions move faster than leading ends or edges, inclusions are flattened and crystals rotated toward parallelism with the chamber walls. This type of flow can produce a strong foliation but only a very weak lineation or no lineation. It explains almost all of the observed magmatic foliation in the granitoids and provides strong evidence that the magmas ballooned as they were intruded.

Acceleration flow occurs when magma moves through an orifice of diminished size, which requires increased velocity of flow. The increasing rate of flow in the direction of flow stretches mafic inclusions and orients 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 dike-like intrusions. However, it should produce a strong magmatic lineation, which has not been oberved in the Sierra Nevada.

In velocity gradient flow, drag along an interface causes the rate of flow parallel to an interface to decrease as the interface is approached. Initially, the interface is a sharp contact of crystal-laden magma with another rock, but with inward crystallization a transitional zone between flowing and stagnent solidifying magma moves away from and effectively replaces the contact. The increased rate of flow with distance outward 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, which has not been observed in undeformed granitoids.

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 was 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 was intruded from the northeast. The isotopic ages of these two intrusions differ by more than 10 million years, 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 contain only a very minor amount of granoblastic material between grains, and quartz exhibits no more than the usual amount of strain. Presumably the Bass Lake Tonalite was heated to a temperature above its solidus during intrusion of the Dinkey Creek Granodiorite.

Ductile foliation and lineation

Ductile foliation and lineation in the granitic rocks were caused in two ways: (1) by regional stress after the deformed rock was completely crystallized, or (2) as the result of continued movement of the magma into an advanced state state of crystallization.

Deformation attributed to regional stress

Evidence of deformation is more common in plutonic rocks of Triassic and Jurassic age than in rocks of Cretaceous age. Nevertheless, regional stresses have affected all or part of the Sierra Nevada repeatedly, and linear belts of shearing are present in some Cretaceous intrusions. The effects of deformation range from intergranular granoblastic material in weakly deformed rocks to almost complete recrystallization and lenticular layering in gneisses. The clearest evidence of the regional deformation of an older pluton is in the Jurassic tonalite of Granite Creek, which aparently was involved in the Nevadan orogeny. Pervasive northwest-trending ductile foliation and lineation in the tonalite is continuous with foliation and lineation in the adjacent metamorphic rocks.

A much more subtle type of deformation that probably resulted from regional stress is present in the ~87 m.y. old Tuolumne Intrusive Suite and in the Morgan Creek pluton of the Lake Edison Granodiorite of the approximately coeval John Muir Intrusive Suite. All units of the Tuolumne Intrusive Suite and the Morgan Creek pluton of the Lake Edison Granodiorite contain a faint, second northwest-trending foliation that crosses contacts and magmatic foliations, generally at high angles. The presence of this second foliation is indicated by the preferred orientation of small biotite flakes; and it is also indicated in the Morgan Creek pluton by closely spaced microfractures (Cray, 1981). The biotite flakes in the Tuolumne Intrusive Suite are indistinguishable from common yellow-brown biotite, showing that the second foliation was produced at relatively high, magmatic or near-magmatic temperatures. According to Cray (1981), the second foliation in the Morgan Creek pluton is also present in older granitoids west of the Morgan Creek pluton but not in younger granitoids.

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 appear to have originated during the early Late Cretaceous but at somewhat different times. Most cut rocks as young as ~97 Ma and are truncated by rocks as old as ~90 Ma. However, one younger shear zone in the ~90 Ma Lake Edison Granodiorite is cut off at the contact with the Mono Creek Granite, which contains a still younger shear zone a few kilometers to the north.

An extensive zone of shearing within the Goddard septum affected both metamorphic and plutonic rocks. Mafic dikes thought to be part of the ~148 Ma old 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 intruded before any shearing occurred or were intruded 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 of the shear zones extends southeast from Courtright Reservoir to the Cape Horn septum (Bateman and others, 1984b). 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 granite of Lost

Peak, a small mass that intrudes the Dinkey Creek Granodiorite, but not into the Mount Givens Granodiorite.

As the contact between the Dinkey Creek and Mount Givens Granodiorite 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 down-dip 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).

A variety of felsic dikes intrude the deformed Dinkey Creek Granodiorite where a septum separates the Dinkey Creek and Mount Givens. 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 undeformed coarser grained dikes are offshoots of the Mount Givens Granodiorite.

Inclusions of sheared Dinkey Creek in the undeformed Mount Givens Granodiorite and undeformed dikes of Mount Givens in sheared Dinkey Creek clearly show that the shearing occurred before emplacement of the Mount Givens Granodiorite was campleted amd suggest that it represents faulting that occurred along a preexisting contact of the Dinkey Creek Granodiorite with metamorphic rocks before the Mount Givens Granodiorite was emplaced. However, planor foliation in the adjacent Mount Givens Granodiorite and an absence of lineations allows the possibility that expansion of the Mount Givens magma caused the ductile deformation in the Dinkey Creek Granodiorite and that expansion cameto an end before the Mount Givens magma crystallized sufficiently to sustain a shear stress. Continuation of the ductile structures in the Dinkey Creek Granodiorite into a small body of granite (the granite of Lost Peak in figure 13) and the nearby granite of Short Hair Creek, which also intrudes the Dinkey Creek Granodiorite, shows that the deformation occurred after the Dinkey Creek Granodiorite was emplaced and had solidified and was not part of the mechanism of intrusion of the Dinkey Creek Granodiorite. The downdip elongation of inclusions and the 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.

Deformation assiciated with emplacement of intrusions

Although outcrop patterns (pl. 1), planar foliation, and absence of lineation indicate that intrusions ballooned as the were emplaced, only a few studies have been made of the expectable wall rock deformation. Some intrusions show no evidence of ductile deformation whereas others show increasing evidence of ductile deformation toward their margins. Intrusions that show no evidence of ductile deformation apparently stopped expanding before they had crystallized sufficiently to sustain shear stress, whereas those that show evidence of ductile deformation continued to expand after crystallization in the margins was well advanced. Intrusions such as the Mount Givens Granodiorite and the Tuolumne Intrusive Suite, which show little or no evidence of ductile deformation have, nevertheless, deformed their wall rocks and probably account for at least some of the cleavages and lineations within them. Apparently a magma too fluid to sustain a shear stress can, nevertheless, produce cleavage, lineation, and shears in adjacent solid rock. The structural effects of intrusion are readily apparent in and adjacent to intrusions that ballooned as they were emplaced and 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 granitoids.

Intrusions that were in late stages of crystallization when they were emplaced and have ductile fabrics include the megacrystic granites of Birch Creek and of Papoose Flat in the northern Inyo and White Mountains (Nelson and Sylvester, 1971; Sylvester and others, 1978b; Nelson and others, 1978) and the Ward Mountain Trondhjemite and granite of Hogan Mountain of the Fine Gold Intrusive Suite. All of these intrusions are composed of leucocratic rocks that crystallized in relatively low temperature ranges from viscous magma. Outward dipping external contacts of all but the granite of Hogan Mountains indicate that the exposures are of their upper parts. They may be localized in the flanks of the batholith because less rock has been removed by erosion and ambient temperatures were lower than in the central part of the batholith where contacts are uniformily steep, erosion deeper, and ambient temperatures probably higher. The structural effects in the wall rocks of these intrusions are more intense than in the wall rocks of intrusions that were hotter and more fluid when the were emplaced, and they produce conspicuous lineations in their wall rocks. They are described in more detail in the following pages.

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 pre-existing 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 Mount Hoffman pluton of the Cathedral Peak 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 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 and others, 1983) is centered on the Ward Mountain Trondhjemite, which forms two adjacent plutons, one just south of the other (Bateman and others, 1983a). North, east, and south of these plutons for distances ranging between 5 and 15 km, all of 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 plutons, 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 intrusion 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 and others, 1983a). 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 approximate a critical isotherm, possibly in the range of 300 to 400 °C where dislocation motion can occur in dry quartz but not in feldspar (Tullis and Yund, 1977).

Isotopic age determinations show that all of 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 emplacement of the granite of Birch Creek, and Sylvester and others (1978b) and Nelson 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 west margins of both plutons were deformed during intrusion, the grain size was reduced, and strong foliations and lineations were 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 with 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 and others (1978b) attributed the deformation to forcible emplacement of the deformed west end of the pluton when it was extremely viscous and almost completely crystalline. During 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.

The granite of Pellisier Flats also was deformed during emplacement, but a detailed study of the deformation has not been made (see p. 300-302). Like the Birch Creek and Papoose Flat plutons, the granite of Pellisier Flats was intruded into sedimentary rocks, and no older granitoids were deformed as a result of its emplacement.

Mafic inclusions

Dark, fine-grained inclusions with igneous textures are by far the most abundant material included 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 of the hornblende and biotite is in the form of euhedral crystals. In reports on the Sierra Nevada published since 1960, these inclusions have been referred to as mafic inclusions. They have also been designated basic segregations (Knopf and Thelan, 1905); autoliths (Holland, 1900; Pabst, 1928), basic concretions (Grubenmann, 1896), 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 decrease gradually from more than 2 per square meter in the margins where the host rock contains more than 15 percent of mafic minerals to zero in the interior where the host rock contains less than 6 percent of mafic minerals (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 a square meter 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 as seen in outcrop are greatest near external contacts of plutons and generally decrease with distance from the contact. Despite their apparent linearity in outcrops, the mafic inclusions are essentially 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 contacts.

The variably flattened shapes of the mafic inclusions in different parts of intrusions show that their shapes have changed in accordance with movements in the enclosing granitic magma. The absence of any indications of ductile deformation indicates that the inclusions were soft and contained interstitial melt when they attained their shapes; probably they were only a little less viscous than the granitic magma, and were themselves 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, as much as 15 percent quartz, little or no alkali feldspar, and relatively abundant opaque minerals and acicular apatite. Studies in progress by Bernard Barbarin (oral and written commun., 1985-1986) show that Fe/Mg in hornblende and biotite, and the range of the anorthite content in 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, $^{87}Sr/^{86}Sr$ of mafic inclusions in the Bass Lake Tonalite (fig. 17) plot on the isochron for the enclosing Bass Lake Tonalite, whereas $^{87}Sr/^{86}Sr$ of 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 minerals in the inclusions and the same minerals in the host granitoid and equilibrium with respect to $^{87}Sr/^{86}Sr$ of the inclusions and the host granitoid mask the earlier history of the inclusions, which must account for their bulk compositions.

Most of the constituents in the mafic minerals probably were derived from earlier crystallized anhydrous minerals that were present in the inclusions when they were first formed, as suggested by Burnham (1979). These minerals would include pyroxenes, calcic plagioclae, and possibly olivine. Sparse fragments of granitic rock and feldspar crystals present in some inclusions appear to have been incorporated mechanically during emplacement and solidification of the enclosing granitoid, and small amounts of interstitial quartz and alkali feldspar, especially where they are concentrated in the margins, probably were introduced by diffusion through the melt phase of the inclusion. 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 but possibly fed in part by material derived from the mafic inclusions. Granitoids in which almost all of the hornblende and biotite is in euhedral crystals 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 needle-like apatite crystals are the most likely survivors from the original crystallization of the mafic inclusions. Experiments by Wyllie and others (1962) on the system $CaO\text{-}CaFe_2\text{-}P_2O_5\text{-}H_2O\text{-}CO_2$ indicate that slow cooling and crystallization yields stubby crystals of apatite, whereas sudden cooling yields acicular crystals. The acicular apatite crystals present in some granitic rocks--for example, the Tinemaha Granodiorite--may have been derived from mafic inclusions. However, it is also possible that some unknown conditions may also favor the formation of acicular apatite crystals. The cores of the plagioclase in the mafic inclusions are less likely relics of original crystallization because their compositions are the same as in the adjacent granitic rocks, and many are too sodic to be relics. They probably attained their compositions during early crystallization of the granitic rock at depth under high pressure and temperature.

The fine grain size of the mafic inclusions and even finer grained margins of some inclusions is frequently cited as evidence that the inclusions were blobs of basalt 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 changing shapes of the inclusions in different parts of an intrusion indicates that during their residence in granitoid magma they were magma themselves, 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 like those in xenoliths carried from depth in feeders for Tertiary volcanics (p. 177-178), 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, potasium metasomatism has transformed hornblende crystals to aggregates of tiny biotite 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 marginally accreted minerals of the 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, (5) mafic dikes of mantle or lower crustal origin that were intruded into the crystallizing but still mobile granitoid magma and subsequently disrupted, and (6) refractory residue of magma generation.

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 and others (1983b) considered wide variations in the modal proportions and major-element compositions of the mafic inclusions in the granodiorite of Red Lake and Eagle Peak to be inconsistent with an origin from crystallized melt and consistent with their formation from entrained residual material (restite). Sawka (1985) considered high radiogenic strontium content and a greater REE concentrations in the mafic inclusions than in the host Tinemaha Granodiorite to be incompatable with either a primary magma source or a cumulate origin.

At this time the origin of the mafic inclusions in granitic rocks is still an enigma, but current evidence seems to me to favor a refractory residue from the formation of the granitic magmas, probably largely basalt, which was later modified by exchange with the enclosing granitoic magma. Basaltic material could have been introduced into the granitic magmas as magmatic blobs, as mafic dikes that disintegrated with movements in the magma, and as disrupted fragments of mafic lower crustal rocks.

Modes of granitoids and mafic inclusions plotted together in figure 19A form an arcuate field that shows some interesting relations. Mafic inclusions occupy the end of the field that extends toward mafic and accessory minerals, and granitoids occupy the end that extends toward quartz + alkali feldspar. Mafic dikes plot within the field of mafic inclusions. A plot of norms of granitoids (fig. 19B), mafic inclusions, and mafic dikes shows the same curved pattern as the modal plot. In the normative plot, the norms of country rocks within and east of the Sierra Nevada (table 3), graywackes of the Franciscan assemblage, and typical high-alumina basalts also are shown. The distribution of norms suggests that mixing of basalt and the country rocks could have produced the parent magmas of the granitoids, which then fractionated along along the curved trend.

Schlieren

Because "schliere" is the German word for streak, elongate mafic inclusions could be included under the term and probably have been in some studies. However, workers in the Sierra Nevada have followed the usage of Ernst Cloos (1936), who included only rock with layered structure in describing schlieren. Schlieren have many different forms, but all of them are composed of groups of composite layers that grade from dark fine-grained rock on one side to felsic coarser grained rock on the other (fig. 20). These composite layers are separated by sharp interfaces, and uncritical examination may suggest gravity layering, the dark fine-grained rock in each layer grading upward into felsic coarser-grained rock. However, the size grading is the reverse of that resulting from differential settling rates; moreover, many schlieren dip steeply or are vertical, and there is little to indicate that they have been rotated from initially horizontal or gentle dips. Overturned grading has not been observed.

Ernst Cloos (1936), following an earlier interpretation of 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 drums. Rotation of one of the drums produced shear stress that caused the grains to impact with one another and to disperse in accord 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 Barriere (1976) have further explored the mechanism of flow sorting. Shear stress reflects velocity differences (not velocity alone), and figure 21 is a simple illustration of the pattern of shear stress resulting from convection along an interface.

Nevertheless, gravity settling probably 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. This mechanism would be most effective in horizontal or gently dipping layers such as those shown in figure 19A or 19C; it would not be effective in the formation of steeply dipping or eliptical schlieren such as those shown in figure 19B. 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.

Schlieren are most common adjacent to contacts between different granitic rocks and near contacts of granitic rocks with country rocks, but schlieren also occur in dikes and in inclusions of older granitic rocks in younger. In plutons, the shear stress may result from convection, and in dikes from shearing along the dikes during crystallization of the magma. Although the flow sorting within inclusions could have originated earlier, gradational relations with the enclosing rock indicate that it originated after the inclusions were immersed and reheated in the enclosing magma.

In some plutons and intrusive suites, such as the Tuolumne Intrusive Suite, schlieren in both flanks dip outward from the core, and the grain size increases outward and upward in each layer. This arrangement suggests that episodic resurgence of magma in the core of the intrusion caused expansion and shearing in the partly solidified margins (fig. 22). Schlieren in "faerie rings" and in troughs probably result from the pipe-like resurgence from below of magma rich in cumulate material into crystallizing more felsic magma. In many places, the rings and troughs are in areas of abundant mafic inclusions and scattered alkali-feldspar megacrysts. The inclusions could have supplied some or all of the finer mafic material to the schlieren, but it is equally possible that they merely accompanied the finer mafic material.

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, but concentration of an aqueous fluid close to the external contacts of pluton during inward crystallization has not been explained.

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, surrounded by a series of comb layers identical with 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.

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 take place, joints of one strike interfinger those joints of another.

The joint patterns are regional in extent, and joints cross contacts between intrusions with little of 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 (1965a) 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 a centimeter 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 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 Eocene in age. Probably they formed during the Late Cretaceous as the Sierra Nevada was being uplifted, eroded, and cooled (see p. 103).

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 Great Basin is dominated by normal faulting. Microfaults trending generally northeasterly have sinistral movement, whereas those trending northerly 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 36.5° N. A pure-shear constant-volume study based on detailed work at 37°20' N. indicated a maximum extension of 2.3 percent in a N. 61° W. direction. This direction is approximately parallel to Cenozoic tectonic extension directions in the Great Basin (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.

Vertical or steeply dipping contacts of most plutons with metamorphic or older plutonic rocks and a general absence of miarolitic cavities (except in subvolcanic porphyries and a few felsic rocks) and of fractures attributable to intrusion, indicate depths generally greater than those at which brittle fracture would occur. However, other relations indicate that the batholith was not deeply eroded. They include 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, 1965), the tongue of Half Dome Granodiorite that splits the intrusive suite of Washburn Lake, and horizontal and gently dipping foliations in the Lamarck and Tinemaha Granodiorites and numerous small plutons of Bass Lake Tonalite in the western metamorphic belt, which indicate proximity to the tops of these intrusions. Gently outward dipping contacts of the Ward Mountain Trondhjemite in the west 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 east and west margins of the map area 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 kb and Robie and Hemingway (1984) at 4.0±05 kb, 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 hypersthene-bearing mineral assemblages in the southernmost Sierra Nevada, whose presence suggests 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 kb, 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 (fig. 26), Putnam and Alfors (1965) inferred that P_{H2O} during crystallization of the Late Jurassic Rocky Hill stock in the western foothills south of the map area was 1.5 kb. Using the same method, Noyes and others (1983a) deduced a pressure of about 1 kb during the crystallization of late-stage aplites intruded into the granodiorites of Eagle Peak and Red Lake. Sylvester and Nelson (1966) suggested that P_{H2O}=P_{total}=2 to 3 kb for the Birch Creek pluton in the White Mountains on the basis of the 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 kb 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 cordierite [Fe/(Fe+Mg) = 0.7] together with alkali feldspar, muscovite, andalusite, and quartz to indicate pressures of 1.5 to 2 kb in the country rocks at the Pine Creek tungsten mine near Bishop.

Physical constitution of Sierran magmas

Four lines of evidence, none of which is completely convincing by itself, indicate that the 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. Evidence for these conclusions is given in the following paragraphs. Possible exceptions are aplite and small bodies of equigranular leucogranite that crystallized in a narrow, low-temperature range and may have been intruded as melt.

(1) The interpretations offered for the origin of magmatic foliation and lineation (p. 104-107) and for schlieren (p. 122-124) 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.

(2) 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). This interpretation now seems dubious, but if so, the magma was never a complete melt.

(3) During the progressive melting of a volume of rock, it is highly probable that the volume would become unstable and move generally 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 suspension-like behavior. With this amount of melt a magma can move freely in response to gravity and tectonic forces.

(4) 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$\pm$15 °C, and 620$\pm$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 °C and 630 °C, and Sylvester and others (1978b) 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 garnet-pyroxene skarn from 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, for undoubtedly magmatic temperatures were considerably higher. Winkler (1979) stated that the temperature at the contact is somewhat greater 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 temprature of the wall rocks 150 °C.

The water content of a magma is critical to the temperature of the liquidus 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 kb confining pressure, but to decrease with increasing amounts of H_2O (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 first emplaced, and remained undersaturated during most of the period 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 kb total pressure (equivalent to a depth of about 7 km) hornblende is stable in magma of granodorite composition only with 4 percent or more of 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 that 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 kb total pressure and 2 weight percent water is about 1,000 °C (Naney, 1977), a temperature substantially above 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 wall 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 ~0.704 in the Fine Gold Intrusive Suite to >0.710 in some intrusions in the White Mountains indicate that most of the granitoid magmas contained substantial amounts crustal material. Conversely, they also indicate that most of the magmas contained material derived from the mantle and that extra space was required to accomodate 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 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 wall rocks 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 wall rocks was small and the magmas tended toward a spherical form with minimum surface area.

Most, and probably all, outward bulges of plutonic rocks at the surface are underlain at relatively shallow depths by older rocks (see figs. 31 and 32). In most places where granitic magmas have crowded the adjacent country rocks aside, felsic granitic rocks were involved; the lower temperatures and greater viscosities of the felsic magmas 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 have been described on p. 56-58. Here, dislocations of the stratified country rocks during the forcible emplacement of intrusions are described. 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 septum and Bishop Creek and Mount Morrison roof pendants have been disturbed by several intrusions (Rinehart and Ross, 1964; Bateman, 1965, 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 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 and cogenetic Mono Recesses Granite were emplaced.

In the southern part of the Mount Morrison roof pendant, metavolcanic strata follow the southern contact of the the northwesternmost pluton of Round Valley Peak Granodiorite (called the Lee Lake mass by Rinehart and Ross 1964), through and arc of 45°, 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 earler 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, deformation of the wallrocks adjacent to the granites of Papoose Flat, Birch Creek, and Pellisier Flats has been mentioned briefly in connection with the ductile deformation of marginal parts of these intrusions during their emplacement (p. 108). The sedimentary strata around the western side of the Papoose Flat pluton have been thinned to about 10 percent of their regional thickness (Sylvester and others, 1978), and the strata around the western 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

Stoped blocks of country rocks can make space for the intrusions by exchanging places with magma and either sinking to lower levels or rising to higher levels where they may have been eroded away. However, no evidence has been recognized for either the sinking or rising of stoped blocks. Visible stoped blocks within the Sierra Nevada generally are relatively small and confined to the margins of intrusions, and 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 a large block may have subsided 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 question of how space was made for the tonalite is uncertain.

Prior to erosion, additional evidence of stoping probably would have been evident at higher levels of exposure where brittle fracture is more common. Pitcher and Cobbing, (1985) cite flat roofs, steep sidewalls, and "cut out" outcrop patterns 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

Absence of evidence that any intrusion was emplaced during regional tectonism, although some intrusions have obviously deformed themselves or were subsequently deformed, implies that the intrusions were emplaced during inactive or extensional regimes and not during tectonic compressive regimes.

However, the amount of extension appears to have been limited. The dissimilar compositions of the Jurassic granitoids on the two sides of the batholith precludes them having been emplaced in adjacent positions and spread apart the distance between them during intrusion of the Cretaceous granitoids. Furthermore, the absence of granitoids associated with the Independence dike swarm of mafic dikes, which extends from the latitude of Bishop southerward to beyond the southern 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 sparse evidence of deformation in the wallrocks attributable to the intrusions suggest that their wall rocks 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 greatest. The most obvious area that swelled 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 linear and parallel to other intrusions of the John Muir Intrusive Suite. If these bulges are removed, 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. However, it seems equally likely that the faults steepen with increasing depth. Such faults can develop during regional arching, with little or no extension, on the flanks of a downdropped keystone block. The arching could have been caused by the tumescent rise of linear belts of magma.

Erosion and volcanism

Erosion that accompanied and followed uplift during the emplacement of the batholith and expulsion of material through volcanism provided substantial amounts of space for the batholith. Geobarometers indicate that about 3.5 to about 7 kilometers of rock was 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 relief in the region of the Sierra Nevada was very low by Eocene time (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 air-born tuff, also provided space. Ash-flow tuff would have been quickly eroded and carried westward, whereas prevailing winds would have carried the air-born tuff eastward.

Sequence of crystallization of minerals

The order in which minerals crystallized has been partially 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 interstitial 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 intruded or to have precipitated shortly thereafter, whereas the interstitial grains crystallized from interstitial melt. Inward from the margins, the first subequant grains to appear 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 rounded 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 these magmas were ever heated to liquidus temperatures, and the sequence of crystallization of subequant grains in rocks closest to the margins of the intrusions cannot be established by field or petrographic observations. However, laboratory studies indicate that plagioclase would have formed at the liquidus and been followed by clinopyroxene, which occurs as cores in hornblende (Naney, 1977). Hornblende and biotite would have begun to crystallize next, but their relative order is uncertain. In both the Tuolumne Intrusive Suite and the Mount Givens Granodiorite, rapid inward decrease in the amount of hornblende and slow decrease in the amount of biotite reflect progressive depletion of MgO, CaO, and Fe in the melt phase of the magma (see figs. 33, 34). The constant abundance of both magnetite and 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 almost all the small amount of alkali feldspar in tonalite, trondhjemite, and most bodies of granodiorite is interstitial.

In contrast with 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 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 is consistent with the relations predicted by the salic tetrahedron (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 (1950), Schairer and Bowen (1947), Tuttle and Bowen (1958)' Yoder and others, (1957), Franco and Schairer (1951), 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 kb, but a few experiments were conducted with P_{H2O} as low as 0.5 kb and as high as 10 kb.

The most realistic conditions for the Sierra Nevada batholith would be melt undersaturated with water at confining pressures of about 2 kb (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 2 kb P_{H2O}. Changes in confining pressure cause temperature differences and shifts in the positions of the two internal phase-boundary surfaces. With increasing P_{H2O}, the temperatures at which minerals of given compositions crystallize 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.

In a melt whose composition falls in the quartz liquidus-phase volume, plagioclase is the first mineral to crystallize with falling temperature, whereas alkali feldspar is the first mineral to crystallize in melt whose composition falls in the alkali feldspar volume, and plagioclase is the first mineral to cryalize in a melt whose composition falls in the plagioclase volume. The norms of all Sierran granitoids fall within or on the internal phase-boundary surfaces of the plagioclase liquidus-phase volume. 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.

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 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 rocks of the White Mountains are shown on a diagram (fig. 28) in which An 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 intrusive suites are discussed below: the calcic Fine Gold Intrusive Suite in the western foothills, the calc-alkalic Tuolumne Intrusive Suite in the high Sierra Nevada, and the alkali-calcic Soldier Pass Intrusive Suite in the White and northern Inyo Mountains.

<u>Low-Or Fine Gold Intrusive Suite</u>

A granitoid parent magma such as that which crystallized to form the low-Or 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 amounts of An crystallizes, causing the melt to follow a curved path within the tetrahedron until it intersects the quartz-saturation surface. If the melt contains no K_2O, crystallization of plagioclase and quartz in constant proportions and at a constant temperature continues until crystallization ceases. 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 when quartz, plagioclase, and alkali feldspar crystallize together as the melt follows the cotectic toward the temperature minimum until crystallization ceases. 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 across the quartz-saturation surface; only a small amount of crystallization occurs after it reaches the cotectic. Consequently, the small amount of alkali feldspar is interstitial to the other minerals.

Medium-Or Tuolumne Intrusive Suite

A melt with medium amounts of Or such as the Tuolumne or John Muir 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. 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 between An_{35} and An_{32} and that alkali feldspar began to crystallize only slightly later when the composition of the crystallizing plagioclase was between An_{32} and An_{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-Or Soldier Pass Intrusive Suite

In an alkali-calcic suite such as the Soldier Pass Intrusive Suite, as on other 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. 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, it will move, as temperature decreases, 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, if crystals are separated or accumulated, the complete range of compositions as found in the Sierra Nevada batholith can be produced. In fact, the general agreement of the succession of rock compositions found in the different intrusive suites with those expected to result from partial separation of crystals from the melt during crystallization is one of the strongest arguments indicating that fractional crystallization was one of the principal mechanisms that produced the variety of rocks present within suites. 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 (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 the 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. Neverthless, 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 crystallization 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. From this concentration, modes plot 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 crystallized chiefly on the quartz-saturation surface in the salic tetrahedron. 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 higher-than-average color indices and higher maximum and minimum An compositions in plagioclase. The large amounts of 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. The modes that extend into the granodiorite and granite fields represent rocks having lower-than-average color indices and higher contents of alkali feldspar and represent substantial crystallization along 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 $^{87}Sr/^{86}Sr$ (fig. 37) in samples collected near the margins of suite, and which plot close to the Q-P sideline, in the tonalite, quartz diorite, and adjacent parts of the quartz monzodiorite and granodiorite fields in figure 29, indicate these samples owe their compositional variations, at least in part, to the mixing of source materials (see p. 83--84). The trend in figure 29 indicates that plagioclase began to crystallize first among the felsic minerals and was followed closely by quartz, and shortly thereafter by alkali feldspar. A linear trend directly toward P would indicate either mixing of end-member magmas or simultaneous crystallization of quartz and alkali feldspar, following crystallization of plagioclase. The abundance of modes in the granodiorite and adjacent part of the granite fields indicates that much of the crystallization was on the cotectic.

Soldier Pass Intrusive Suite

The modal plot for the Soldier Pass Intrusive 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, but most are in the quartz monzondiorite and granodiorite 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 mineral assemblages that crystallized on the two-feldspar surface of the salic tetrahedron, whereas modes that plot in the quartz monzodiorite, quartz monzonite, granodiorite and granite fields represent later crystallization along the cotectic.

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 dominent 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 anatected magmas generated in the crust. However, if it ponds at the base of the crust and acts only as a carrier of heat, it, nevertheless, 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 level of exposures in the Sierra Nevada are scarce, but mixtures of more mafic and 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. The most convincing evidence of magma mixing is found in radioactive isotopic ratios because they can be mixed but not unmixed. 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 of initial $^{87}Sr/^{86}Sr$ within an intrusion indicate incomplete homogenization or continued mixing. For example, fairly regular inward increase of initial $^{87}Sr/^{86}Sr$ in 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 of 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 occur in the chronology 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 from the central Sierra Nevada were of 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 ranged from 82.4 to 95.2 Ma, decreasing eastward in accord 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 the result of reheating during the west-to-east emplacement of intrusions.

Somewhat later, the K-Ar ages of hornblende as well as biotite were determined on 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 from the central part of the traverse (from the Mount Givens, Lamarck, and Round Valley Peak Granodiorites) are concordant in the range of 80 to 90 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 from the central segment were interpreted to be crystallization ages and the discordant ages to be reduced from older crystallization ages as the 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 m.y. and the ages of most of the other rocks to be considerably older. Subsequently, the 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; Bateman and Lockwood, 1976; Bateman and Sawka, 1981; Kistler, 1974) were published.

Evernden and Kistler (1970) published a summary of all the K-Ar ages that had been determined in the Sierra Nevada, which included many new ages on samples chiefly from north and south of the map area. These K-Ar ages show that a dense belt of Cretaceous intrusions trending about N.20° W. crosses a belt of scattered Jurassic plutons trending about N.40° W. within the map area (fig. 30) (Kistler and others, 1971). Evernden and Kistler (1970) also proposed five intrusive epochs that occurred periodically over a span that extended from 81 to 215 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.

The first U-Pb isotopic ages were published in 1978 and 1979 of granitoids of the White Mountains (Sylvester and others, 1978a; Gillespie, 1979). In 1980, these ages were supplemented by U-Pb ages on granitoids in the western foothills south of the map area (Saleeby and Sharp, 1980). Stern and others (1981) published 62 U-Pb ages on 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 from the east side of the map area (pl. 1). Most recently, Dodge and Calk (1986) published a few U-Pb ages on 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. All the U-Pb ages are summarized graphically in figure 31.

To date, relatively few U-Pb ages have been determined on the metavolcanic rocks, but a few have have been published by Fiske and Tobisch (1977) and Tobisch and others (1986) and by Hanson and others (1987). A single U-Pb age on a sample of rhyolite tuff from the White Mountains is given in table 1 (p. 28).

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. Ages determined in different laboratories and on different rock units show good agreement in indicating the intervals of plutonism and suggest that the distribution of ages within the map area is representative of the distribution throughout the 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 postulated five intrusive epochs of plutonism, on the periodicity of the epochs, and consequently on interpretations premised on the reality of the epochs (Kistler and others, 1971; Shaw and others, 1971). The U-Pb ages of samples collected from within the map area indicate markedly increased plutonism between approximately 214 and 201 Ma (latest Triassic and earliest Jurassic), 172 and 148 Ma (late Middle and Late Jurassic), and 119 and 86 Ma (late Early and early Late Cretaceous). The isotopic ages on rocks collected south of the map area agree reasonably well with these periods of accelerated plutonism; thus, they may 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 epochs. 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 fall close to this boundary, ages that fall between 97 and 101 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 $^{206}Pb/^{238}U$ ages unless otherwise indicated. K-Ar ages, especially hornblende ages, are used where U-Pb ages are dubious or lacking. All of 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 utilized.

The isotopic ages of the plutonic rocks show the following distribution patterns within the map area (fig. 32): (1) Triassic granitoids occur in the eastern 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 not into the White and Inyo Mountains. (2) Jurassic plutonic rocks are present in the eastern escarpment of the Sierra 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 crossing, on a regional scale, 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 on these rocks range from 123 to 217 Ma, but all determinations of less than 190 Ma were on rocks that also yielded ages older than 190 Ma on other samples or on 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 on the Tungsten Hills Granite (198 and 202 Ma) are significantly younger (Stern and others, 1981; Chen and Moore, 1982). Nevertheless, the Tungsten Hills Granite is included in the suite because of its close association with the Wheeler Crest Granodiorite and because the scatter of isotopic ages on both the Wheeler Crest Granodiorite and the granite of Lee Vining Canyon include ages younger than 200 Ma.

Jurassic intrusive rocks

The Jurassic intrusive rocks on the western side of the batholith include the tonalite of Granite Creek, with a U-Pb age of ~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 fall into two age groups. Two small plutons west of the map area having U-Pb ages of 182 and 190 Ma lie west of the Melones fault zone and intrude the Penon Blanco Volcanics to which they may be related. The other two plutons range in age from 140 to 151 Ma, and belong to a group of intrusions in the western foothills that includes the granite of Woods Ridge, which were intruded following the Nevadan orogeny. U-Pb ages on rocks south of the map area and K-Ar hornblende ages on rocks north of the map area indicate that most of these foothills plutons were intruded between about 154 and 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 on the eastern side of the batholith include the Palisade Crest Intrusive Suite with Rb-Sr isotopic ages of about 170 Ma, which is exposed chiefly in the eastern escarpment of the Sierra Nevada but which extends across Owens Valley into the Inyo Mountains; 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 with U-Pb ages of 161 and 165 Ma; the granite of Sage Hen Flat with a U-Pb age of 144 Ma; the leucogranite of Casa Diablo Mountain with a U-Pb age of 161 Ma; and quartz diorite from the Pine Creek tungsten mine 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.

Farther west, in the Goddard septum, Chen and Moore (1982) reported a U-Pb age of 157 Ma on a mass of sheared, fine-grained quartz syenite that 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, about 85 Ma. The oldest isotopically dated intrusions are those that were intruded in the western foothills beginning in latest Jurassic time following the Nevadan orogeny amd discussed in the previous section. These intrusions are scattered and relatively small. The Cretaceous magmatism that accounts for the great bulk of the batholith began with intrusion of the Fine Gold Intrusive Suite in the west side 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 span of several million years. It was followed in eastward succession´ by the Shaver Intrusive Suite and the coeval intrusive suite of Yosemite Valley (both about 103 Ma); the approximately coeval intrusive suites of Buena Vista Crest, Merced Peak, and Washburn Lake (all about 98 Ma); and the approximately coeval Tuolumne and John Muir Intrusive Suites (both about 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 younging.

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 on biotite of 161 Ma and on hornblende of 92 and 100 Ma. The ages of the other Cretaceous granites, indicated only by K-Ar determinations on biotite, range from 74 to 87 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 across the map area, based chiefly on the U-Pb ages published by Stern and others (1981), and the other farther south between lats 36° and 37° N., based on their own U-Pb ages. 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 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 in conjunction with intrusive relations with other suites suggest that the main phase of plutonism associated with a single magmatic epoch 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 of the concordant ages except one age of 105 Ma fall between 116 and 112 Ma. Seven U-Pb ages from the John Muir Intrusive Suite range from 93 to 88 Ma, and three from 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 of 19 zircon and 2 titanite size fractions of 11 samples from a nested and compositionally zoned suite of rocks, called the Sequoia Intrusive Suite by 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 m.y. They conclude that the entire suite was emplaced in a relatively short span (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 of 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 partial or complete 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, 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, especially ones that show compositional and textural 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. 33). 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 intruded as a single mass and disrupted during emplacement 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 span at the north end of the western contact. Compositional changes occur both gradually within units and abruptly at contacts. The different units were emplaced as magmatic surges (fig. 34). In places where a younger surge rose alongside the next older surge without intruding it deeply, compositional and textural changes are almost entirely within the units, but where a surge cut across or penetrated 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. 33) and concluded that the compositional zoning is the result of crystal fractionation. They suggested that crystals with which the magma was saturated acreted 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) have 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 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 studied by Bateman and Chappell (1979) follows California State Highway 120, except for a 5-km segment at the west end, which runs across country to the western external contact of the suite north of May Lake (fig. 33). Where this traverse crosses contacts, compositional and textural differences on the two sides are small, and changes take place gradually across units rather than abruptly at contacts. A major exception is the contact of the Johnson Granite Porphyry with the Cathedral Peak Granodiorite, which is abrupt.

The following discussion is largely of 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 $^{87}Sr/^{86}Sr$ along this traverse are shown graphically in figures 35, 36, and 37.

Quartz and alkali feldspar, sparse in the marginal rocks, increase in abundance and grain size inward well into the equigranular facies of the Half Dome, then remain constant across the megacrystic facies of the Half Dome and the Cathedral Peak Granodiorites (fig. 35). Matching these variations are corresponding 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. Farther inward, hornblende continues to decrease at a diminishing rate and is absent in the inner part of the Cathedral Peak Granodiorite and 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, then more gradually farther inward.

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. 36). 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 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 gradually inward from the margins to the core of the suite (fig. 35). These progressive changes indicate inward decreasing temperatures of crystallization or increase of H_2O, or both. Overall, plagioclase in this suite contains distinctly less anorthite than other Sierran intrusive suites having the same general compositional range, and suggests a relatively high content of H_2O in the magma, though less than the amount required for saturation until the Johnson Granite Porphyry was emplaced. The occurence 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 indicates that the magma was saturated in H_2O when the Johnson Granite Porphyry was emplaced.

Variations of initial $^{87}Sr/^{86}Sr$ along traverse A-B (fig. 33) are shown graphically in figure 37 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 intial $^{87}Sr/^{86}Sr$ is 0.0009, from 0.7057 to 0.7066. Although the range of replicate analses of individual samples is as much as 0.0003, the plot in figure 37 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.

As strontium isotopes can mix to give an average value, but cannot unmix during crystal fractionation, the progressive 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 require no further mixing until the Johnson Granite Porphyry was emplaced, where a small but sharp increase of $^{87}Sr/^{86}Sr$ occurs.

The close correspondence of variations of SiO_2 with variations of initial $^{87}Sr/^{86}Sr$ suggests that mixing of source materials can explain all of the compositional variations within the Tuolumne Intrusive Suite, and that no other mechanism of differentiation is required. However, systematic varions of other oxides in the inner part of the suite where SiO_2 and initial $^{87}Sr/^{86}Sr$ are constant indicate that another 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. 36). These changes reflect decreasing amounts of hornblende and biotite inward in the suite and decreasing anorthite in plagioclase. They are the result of decreasing temperatures of crystallization and (or) increasing H_2O inward, which control the solubility of these constituents in the melt phase of the magma. Plagioclase crystals contain more CaO and less Na_2O, and 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 in the interior of the magma and Na_2O enriched.

The explanation that I presently prefer for the compositional and isotopic patterns in the Tuolumne Intrusive Suite is that 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 outer, compositionally and isotopically strongly zoned outer part of the Tuolumne Intrusive Suite. Discontinuous, somewhat 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 the 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 initial $^{87}Sr/^{86}Sr$. Although all of the compositional variations in the outer part of the suite could possibly be attributed to magma mixing and mingling, crystal-liquid fractionation should operate wherever a thermal gradient exists in a granitic magma and it is unlikely that it was confined to the inner part of the suite. It probably played a role in producing the compositional changes in the outer part of the suite as well as in the inner part.

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 in composition 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 the possibility that the variations are chiefly vertical. Units of the Shaver Intrusive Suite that intrude the Dinkey Creek include the megacrystic granodiorite of McKinley Grove, several younger masses of equigranular granite, and the granodiorite of Whisky Ridge (fig. 38).

Isopleths drawn at 2.70 and 2.75 on specific-gravity determinations of samples collected from the Dinkey Creek Granodiorite reflect variations in bulk rock composition (fig. 38). 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. 38) shows that the tongue was forcibly protruded, and it seems reasonable to infer a shallow bottom. Section A-A' (fig. 38), 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 is compositionally and texturally similar to the megacrystic facies of the Dinkey Creek Granodiorite and has similar specific gravities of less than 2.70. Nevertheless, the granodiorite of McKinley Grove has sharp contacts with the Dinkey Creek Granodiorite and was intruded after the magmatic foliation in the Dinkey Creek Granodiorite was established. It was then 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. 38) imply that the magma was stratified according to density before the tongue of Dinkey Creek Granodiorite was protruded toward the southwest, folded, and steep foliations imposed. However, the granodiorite of McKinley Grove was intruded later, after magmatic foliation had been imposed on the Dinkey Creek Granodiorite. It is unlikely that unmelted country rocks remained at depth after the Dinkey Creek magma was generated 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 salic magma was concentrated as the result of fractionation processes in the core of the cooling and crystallizing Dinkey Creek magma as well as at the top.

<u>Mount Givens Granodiorite</u>

The Mount Givens Granodiorite consists of a large elongate pluton that terminates at its northwest end in a bulbous head. Most of the rock is equigranular granodiorite, but the core and concentric zones within the bulbous head are megacrystic (fig. 39). 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 in the Tuolumne Intrusive Suite, and 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 the Tuolumne Intrusive Suite and the presence of ilmenite, absent in the Tuolumne Intrusive Suite, indicate a lower H_2O content in the Mount Givens magma. The occurrence of most of the hornblende and biotite as euhedral crystals in the Half Dome and Cathdral Peak Granodiorites of the Tuolumne Intrusive Suite and a dearth of such crystals in the Mount Givens Granodiorite may also reflect the lower H_2O contant of the Mount Givens magma.

The spatial relations of the different facies of the Mount Givens have been interpreted in various ways in earlier publications (Huber, 1968; Bateman and others, 1970, 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 the enclosed porphyritic core rock, represents a resurgent intrusion from below. 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) has attributed to extraordinarily effective sidewall crystallization made possible by slow cooling because of the presence of surrounding heated rock.

Obviously, the bulbous head of the pluton expanded at the exposed level 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 original horizontal or gently dipping layering in which megacrystic magma overlaid denser equigranular magma as in the Dinkey Creek Granodiorite of the Shaver Intrusive Suite (conjectural section A-B'in fig. 39).

Bateman and Nokleberg (1978) examined a series of samples that were collected along a traverse that extends from the south margin of the pluton to the core of the bulge (fig. 40). Along this traverse, the average grain size increases inward, largely reflecting the increasing size and abundance of quartz and alkali-feldspar grains; grains of mafic minerals actually become smaller inward. Inward decrease in the grain size of the mafic minerals is consistent with a lower H_2O content in the Mount Givens than in the Tuolumne magma, where the mafic minerals increase inward in size into the Half Dome Granodiorite before decreasing in size. The inward decrease in the size of the mafic minerals shows that the fine grain size in the margins is not the result of chilling. This observation may apply to other intrusions with fine-grained margins, which have been attributed to chilling.

The general pattern of abundance of minerals along the traverse investigated by Bateman and Nokleberg (1978) is for quartz and alkali feldspar to increase inward, for plagioclase to be little changed, and for hornblende and biotite to decrease (fig. 41). These changes are interrupted at the intrusive contact of the horseshoe-shaped mass of granodiorite, and are partly repeated farther inward. The degree of compositional change is greatest in the outer 1 km, as it is 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 as Bateman and Chappell (1979) 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, but cannot be evaluated because of a lack of isotopic data. However, similar compositional zoning in the Palisade Crest Intrusive Suite occurs in the absence of isotopic change.

Palisade Crest Intrusive Suite

Sawka (1985; and written and oral communs., 1982-1985) investigated compositional variations within the northwestern part of the Palisade Intrusive Suite in the eastern escarpment of the Sierra Nevada southwest of Big Pine. The area 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. 42). 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 are the result of some other mechanism 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 (6,800 ft) in the southeast corner to 4,285 m (14,058 ft) in 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 intruded, but the presence of extremely flattened and attenuated mafic inclusions close to this contact in both intrusions suggests that the marginal part of the Tinemaha may have been remobilized when the Inconsolable Quartz Monzodiorite was intruded. The foliation in both rocks 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 (10,800 ft), which extends inward from the west margin of the Tinemaha Granodiorite, across the zone of steeply dipping foliations and into the zone of horizontal foliations, then across the granodiorite of Mc Murry Meadows (traverse A-A' in fig. 42). He also collected samples upward from this traverse thru a vertical distance of about 900 m to an altitude of 4,200 m (13,665 ft), chiefly along traverses B-B' and C-C' in fig. 42). The samples collected along the equal-altitude traverse show that inward from the western 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. 43). Samples collected adjacent to the Inconsolable Quartz Monzodiorite, within the zone of steeply dipping foliations, and to a lesser extent from the other end of the traverse 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. Clinopyroxene is present in the west margin but is scarce in the interior of the intrusion.

Samples collected within the zone of horizontal or gently dipping foliations between 3,300 m (the equal-altitude traverse) and 4,200 m show the following downward variations along traverses B-B' and C-C' (fig. 44): (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 they were gained by diffusion from the interior of the magma chamber caused the magma adjacent to the boundary 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. Figure 45 shows the observed and inferred distribution of rock compositions within the lobe. This model has been demonstrated in laboratory experiments using saline solutions (Turner, 1980; McBirney, 1980).

Granodiorite of McMurry Meadows

The rock in the margins of the granodiorite of McMurry Meadows has almost the same composition as the rock in the west margin of the Tinemaha Granodiorite and, like the Tinemaha Granodiorite, is more felsic inward (fig. 43). However, in contrast with 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 marginal 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. Augite cores are present in some euhedral to subhedral hornblende crystals, especially within the marginal rock.

Sawka (1985) postulated that when the granodiorite of McMurry Meadows was intruded, 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 constituents from the boundary layer by crystallization and marginal accretion and the renewal of constituents by diffusion from the interior magma. Slow but efficient precipitation and accretion of crystals at the sidewalls was almost equalled by diffusion of material from the core magma. Consequently, the compositional change and density decrease within the boundary magma was small, and it did not rise as buoyantly as the boundary magma within the Tinemaha Granodiorite (fig. 45). Thus, the minerals that acreted at the margins represent one peak in a bimodal pattern, and the depleted magma the other peak.

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 peridotites, pyroxenites, 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 papers, Moore (1959), Wollenburg and Smith (1968; 1970), Bateman and Dodge, (1970), Dodge (1972), Bateman (1979), and Dodge and others (1982) have 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 increases gradually 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 minerals on line A-B in figure 32, which extends from the southwest corner of the map through Bishop to the east boundary (fig. 46). The profile 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 about 60 percent in the west to less than 40 percent in the east. Alkali feldspar remains constant across the Fine Gold Intrusive Suite, increases markedly to the east base of the Sierra Nevada, and rises slightly in the White Mountains. Plagioclase decreases irregularly from a little less than 60 percent in the west to about 35 percent in the east, then increases slightly 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 range, and increases to about 10 percent in the White Mountains. Quartz varies irregularly between 20 and 30, percent except in the Tungsten Hills near Bishop where it increases to about 35 percent.

Variations of major oxides

The variations of major-oxide content reflect the regional modal variations. In figure 47, the percentage contents of major oxides of most chemically analyzed samples within the map area have been projected onto line A-B of figure 32 parallel with the long axes of the major intrusions. However, samples in the northwest part of the map area, which are distant from the line of profile, were not used. Because chemically analyzed samples are far fewer than modally analyzed samples and were projected from much greater distances, the plots and moving-average profiles are much less regular than for the modes. Irregularities are especially noticeable in the eastern part of the profile, 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 profile, and the plots of SiO_2 and Na_2O show the least differences. 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 profile, remains constant as far as the eastern base of the Sierra Nevada, and then rises to about 4.5 percent in the White Mountains. The moving-average curve for SiO_2 ranges nonsystematically between 61 and 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 profile.

In figure 49, K_2O is plotted against SiO_2 for several representative suites. The plots clearly show that SiO_2 varies little across this part of the batholith whereas K_2O is progressively higher eastward. The plots also show that the eastward increase of K_2O takes place in steps between the different suites and is not continuous as shown in the profiles along line A-B. 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 is from 57 to 74 percent, but the range is narrower in the Fine Gold Intrusive Suite (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.

The moving-average curve for Na_2O (fig. 47) varies between 3.0 and 3.8 percent except along a short span in the western part of the profile 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) lead to the expectation 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 profile, 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 profile but varies widely in the eastern part of the profile.

The eastward decrease in the content of mafic minerals (fig. 46) indicates that collectively Fe_2O_3, FeO, and MgO should decrease eastward, and the plots substantiate this expectation (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 profile, then varies erratically between 1.0 and 3.0 percent in the eastern part of the profile. The moving-average curve for MgO similarly shows a decrease from almost 3 percent in the west to less than 1.0 percent in the central part of the profile and varies between 1.0 and 2.0 percent in the eastern part. Dodge (1972) showed that the oxidation ratio [mol $(2Fe_2O_3 \times 100)/(2\ Fe_2O_3 + FeO)$] ranges from 7 to 65 and increases systematically eastward across the batholith. He attributed this variation to an eastward increase of oxygen fugacity caused by an increase in the water content of the source materials. However, the eastward increase in the oxidation ratio reflects eastward decrease of FeO rather than increase of Fe_2O_3, which is fairly constant.

Variations in the alkali-lime (Peacock) index

Determination of the alkali-lime (Peacock) index (percentage SiO_2 at equal amounts of CaO and $Na_2O + K_2O$) 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 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 of minor elements

Uranium, thorium, rubidium, beryllium, tantalum, barium, and total rare earths have been shown to increase eastward across the Sierra Nevada (Wollenburg and Smith, 1968; Dodge and others, 1970, Dodge, 1972; Dodge and others, 1982). In support of heat-flow studies being carried on 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, then remain approximately constant eastward into the White Mountains. Uranium increases eastward from an average of 1.5 ppm 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 Hills Granite and Tinemaha Granodiorite. Thorium similarly increases from 4.3 ppm in the Ward Mountain Trondhjemite to 21.5 in the Mount Givens and Tinemaha Granodiorites. The effect on radioactive heat generation of the decreasing amounts of uranium, thorium, and potassium eastward within the granitic rocks is shown in figure 57.

Dodge and others (1970), using semiquantitative spectrographic determinations, showed that rubidium increases eastward across the Sierra Nevada from an average of about 50 ppm in the west to about 150 ppm in the east (fig. 51). Subsequently Dodge (1972) plotted semiquantative data for other minor elements and found that beryllium increases eastward from an average of about 1 ppm in the west to about 3 ppm in the east. Plots of other elements are not definitive but suggest that boron, barium, copper, lanthanum, niobium, lead, strontium, and zirconium may increase eastward. Since these determinations were made, neutron activation analyses have confirmed eastward increase of barium, tantalum, and light rare earths (Dodge and others, 1982) (fig. 51).

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

Isotopic variations

Strontium, neodymium, and oxygen isotopic ratios for rocks of the batholith have been studied. Of these ratios, far more data are available for strontium than for neodymium or oxygen.

Strontium

Study of strontium isotopes within the map area has been carried on almost exclusively by R.W. Kistler and his associates. Initial $^{87}Sr/^{86}Sr$ (ratio at the time of crystallization) of samples of granitoids increase generally eastward from less than 0.704 in the western foothills to 0.707 in the high Sierra Nevada 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 miogeosynclinal Paleozoic crust, that 0.706 and 0.704 are the eastern and western limits, respectively, of eugeosynclinal 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 increasing proportion eastward of crustal to mantle material in the parent granitoid magmas as the crust thickens and 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 these 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 the White Mountains (R.W. Kistler, written communication, 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. 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 deduced from isochrons.

Although changes in both the composition and isotopic characterics of the crust and the proportions of crustal and mantle components affected the isotopic and compositional characteristic 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, slighly less than the lowest initial strontiun 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 0.727 (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 of 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 of Paleozoic and Precambrian rocks in and east of the Sierra Nevada) is assumed for the 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 $^{143}Nd/^{144}Nd$ for many of the same samples as were used by Kistler and Peterman (1973) and converted both the neodymium ratios and initial $^{87}Sr/^{86}Sr$ to $\epsilon_{Nd}(T)$ and $\epsilon_{Sr}(T)$, which are ratios of these values to the ratios in a model chondritic reservoir at the time of emplacement of the intrusion, minus 1. Thus the ϵ value is 0 when the measured ratio is equal to the chondritic ratio. When plotted against each other, these values yield a hyperbolic curve. The $\epsilon_{Nd}(T)$ decreases eastward across the batholith from +6.5 to -7.6, whereas $\epsilon_{Sr}(T)$ increases from -16 to +54. Those $\epsilon_{Nd}(T)$ values greater than 0 fall close to the range of $\epsilon_{Nd}(T)$ values for island arcs, and DePaolo (1981) interpreted rocks with these values to be derived largely from depleted mantle material. Zero $\epsilon_{Nd}(T)$ coincides closely with the 0.704 $^{87}Sr/^{86}Sr$ contour of Kistler and Peterman (1973). DePaolo (1980) postulated that rocks with negative $\epsilon_{Nd}(T)$ contain crustal material in amounts roughly proportional to the negative increase in $\epsilon_{Nd}(T)$ and the positive increase in $\epsilon_{Sr}(T)$.

Determinations by Domenick and others (1983) of isotopic ratios for upper mantle and lower crustal xenoliths in 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 may take place in depth. The $\epsilon_{Sr}(T)$ and $\epsilon_{Nd}(T)$ values of the xenolith samples cover the full range of values for granitoid samples collected across the batholith and fall on the same hyperbolic curve of a plot for $\epsilon_{Sr}(T)$ against $\epsilon_{Nd}(T)$. The xenoliths from the trachyandesite have $\epsilon_{Sr}(T)$ values that range from -10.2 to +382.7 and $\epsilon_{Nd}(T)$ values that range from +4.8 to -17.4. The xenolith from the basanitoid lava flow extends the range of $\epsilon_{Sr}(T)$ values to -24.2 and of $\epsilon_{Nd}(T)$ values to +8.3. The value for the Dinkey Creek Granodiorite falls within the range of values for the xenoliths--$\epsilon_{Sr}(T)$ is +44.5 and $\epsilon_{Nd}(T)$ is -7.7. Interpretation of the probable depth at which the samples attained their present mineral compositions suggests that $\epsilon_{Sr}(T)$ may decrease and $\epsilon_{Nd}(T)$ increase downward as they do from east to west at the surface across the batholith.

Oxygen

Masi and others (1981) report that the range of $\delta^{18}O$ for granitoids of the Sierra Nevada is from 5.5 to 12.3 and that 80 percent of these values lie between 7.0 and 10.00. The value $\delta^{18}O$ is $10^3(^{18}O/^{16}O$ in a sample - $^{18}O/^{16}O$ in standard mean ocean water$)/(^{18}O/^{16}O$ in standard mean ocean water). The distribution of $\delta^{18}O$ is quite irregular, but within the map area (pl. 1) $\delta^{18}O$ appears to decrease slightly toward the east. Masi and others (1981) conclude that most values are primary and related to the compositions of the source materials of 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 other metavolcanic rocks. However, except for a few small felsic intrusions such as the Johnson Granite Porphyry, none of the plutonic rocks shows textural evidence of having erupted. In regions less deeply eroded than the Sierra Nevada, such as the San Juan Mountains of Colorado (Lipman, 1975) or 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 erodable, 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 between about 210 Ma (Late Triassic) and about 85 Ma (early Late Cretaceous), but the distribution of ages suggests that the peaks of volcanic activity may have alternated with peaks of plutonic activity (fig. 55). Alternating peak activity of plutonism and volcanism could reflect changing state of stress--volcanism dominating during times of extension when magmas flowed freely to the surface and plutonism dominating during times of little or no extension, or even compression, when mantle-derived mafic magmas ponded and mixed 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. Easily erodable 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 time span of plutonism.

Dickinson and Rich (1972) investigated the modal abundances of quartz and feldspar grains, 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 eruptives, peak abundances of lithic fragments in the Late Jurassic (Tithonian) and early Late Cretaceous (Cenomanian) correspond with 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 porcellaneous marine shale, especially the Mowry Shale, is enriched in silica as the result of ash fall. Bentonite beds have been dated isotopically at 79, 105, and 110 Ma, and the Mowry shale at 112 to 110 Ma; these ages are in the general range of magmatic activity in the Sierra Nevada though not corresponding exactly to the magmatic epochs 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 volcanic and plutonic rocks. Disagreement between U-Pb and Rb-Sr ages on 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 on the older succession of the Koip sequence range from 214 to 186 Ma, whereas the Rb-Sr whole-rock age for the same rocks is 237 ± 11 m.y. using only samples collected from within the map area, or 224 ± 14 m.y if additional samples from north of the map area are added (Kistler and Swanson, 1981). Another problem is that the U-Pb ages on volcanic rocks do not always agree with the order indicated by the positions of the dated samples 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 a 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 late manifestation of the same magmatic event that earlier 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 on the Scheelite Intrusive Suite are older than those on the older Koip, Kistler (1966a) interpreted a contact between the ~210 Ma granite of Lee Vining Canyon and the Koip sequence to be intrusive, and Bateman (1965) has interpreted a contact between the ~200 m.y. old Tungsten Hills Granite and metavolcanic rocks in the southern part of the Pine Creek septum also to be intrusive. The relations of these two groups of rocks to each other merits further examination.

SHIFTS IN THE LOCI OF MAGMATISM

In figure 54, the U-Pb ages are arranged in columns to show the distribution of rocks and their U-Pb isotopic 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 of 114 and 85 m.y. ago. 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 cause of the shifts in the locus of magmatism are changes 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 spans 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 unit here called 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 on the eastern part of the greenstone of Bullion Mountain. Early Jurassic U-Pb ages on two small plutons west of the map area, which intrude the Penon 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 assumed by Schweickert and Cowan (1975), but if it was confined to brief intervals in the Early and Late Jurassic, only one locus that shifted position from time to time may have existed. Twin loci are most likely to have existed when the ~163 Ma tonalite of granite Creek (Stern and others, 1981) in the western foothills and approximately coeval quartz monzonite of Mount Barcroft in the White Mountains were emplaced, when the younger succession of the Koip and Late Jurassic volcanics of the western foothills were erupting, and when the dominantly pyroclastic strata of Early Cretaceous age were erupted in the High Sierra and plutons were intruded in the western foothill. Aberrant isotopic ages on the Palisade Crest, Scheelite, and Soldier Pass Intrusive Suites obscure age relations and erronously suggest overlapping magmatism during the Late and late Middle Jurassic, in the interval of 177 to 155 Ma (fig. 54).

SUBSURFACE STRUCTURE

Interest in the deep structure beneath the Sierra Nevada began in 1936 when Lawson published "The Sierra Nevada in the Light of Isostasy." In this paper, Lawson used average crustal and upper-mantle densities in conjunction with principles of isotasy to estimate a crustal thickness of about 68 km in the vicinity of Mount Whitney. In a comment on Lawson's paper, Byerly (1938) confirmed the presence of the Sierran root from delay in the arrival times of earthquake waves at seismograph stations east of the Sierra Nevada, which originated 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 follows the axis of the batholith.

Seismic data

Seismic studies provide the principal evidence for the deep structure of the Sierra Nevada and 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 Great Basin.

The locations 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. The fence diagram (fig. 56) is constructed from seismic refraction profiles published by Eaton (1963, 1966) and Bateman and Eaton (1967). It is diagrammatic, and details may be incorrect, but it is the best approximation available 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 in 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 transverse profiles, one of which coincides with section B-B'.

The fence diagram shows that the Moho is depressed beneath the High eastern 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 on sections B B' and C-C' (fig. 55) show that in the northern half of the Sierra Nevada the root trends about N. 30° W. and that the axis closely follows the eastern escarpment of the Sierra Nevada. In this construction, the western 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) have 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 35-40 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 (1963,1966; Bateman and Eaton, 1967) interpretations are broadly correct. However, they concluded that the root is more stongly asymmetric than was indicated by Eaton's data and that the eastern flank rises steeply immediately east of the Sierra Nevada. This modification of the Moho is shown 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 above 12 km depth to 6.4 km/s at about 24 km, 6.9 km/s in the base of the crust, and 7.9 km/s in the upper mantle (Bateman and Eaton, 1967). An 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, with densities of 3.0 to 3.2, are appropriate for P-wave velocities of 7.0 km/s in the upper part of the lower crust and that pyroxenite and eclogite, with densities of 3.2 to 3.4, are appropriate for P-wave velocities of 7.5 to 7.9 km/s in the base of the crust.

Gravity data

Gravity data also show that the base of the crust beneath the Sierra Nevada has the form of an asymmetic trough. However, on the basis of gravity data, Oliver (1977) suggested that the axis of the trough lies 30 km west of the position shown by Bateman and Eaton (1967) on the basis of seismic data. The eastern 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 the seismic profiles A-A' and B-B' (fig. 55) coincides with the western boundary of the seismic root at 45 km. The break in slope of the gravity contours continues southward about half way between the eastern and western sides of the Sierra Navada, and presumably defines the western side of the root in the southern Sierra Nevada.

Both Bouguer and isostatic gravity anomalies for an assumed constant density of 2.67 decrease generally eastward to Owens Valley, then rise abruptly in the White Mountains (Oliver and Robbins, 1973; Oliver, 1984, unpublished gravity map of the Mariposa 1°x2° quadrangle) (fig. 57). Strong negative Bouguer anomalies coincide with thick accumulations of sediment--Owens Valley, Long Valley, the Mono Lake basin, and the Great Valley of California.

From west to east along profile A-B (fig. 32) Bouguer values increase from -85 milligals at the west end to -80 milligals 15 km to the east, decrease to about -230 milligals in the Owens Valley region, then rise to about -200 milligals in the White Mountains (fig. 58). Isostatic anomalies show the same general pattern in subdued form; they increase from about -40 milligals at the west end of the profile to -20 milligals 15 km to the east and to -10 milligals 40 km to the east, decrease gradually to -40 milligals in Owens Valley, then increase abruptly to -10 milligals in the White Mountains.

The high values at the west end of the Bouguer profile coincide with the first exposures eastward of metamorphic and granitic rocks, whereas the high values in the west end 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 decreasing surface densities eastward, 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 residual magnetic intensity profile shown in figure 57 trends 22° more northerly than the profile for gravity data (profile

A-B on figure 32) and crosses it about 10 km south of Huntington Lake (A in fig. 57). The west end of the magnetic 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 broad low just east of the twin highs. The granitoids that correspond with this broad low are chiefly tonalite, 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 hfu (heat-flow unit) 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 (2-7+ hfu) farther east across the Basin and Range province (Lachenbruch, 1968; Roy and others, 1972; Henyey and Lee, 1976).

All of 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 granitic rocks and lower than those obtained in the ocean basins. A plot of heat-flow measurements against heat-generation measurements made at the drill sites yields an approximately straight line that intersects zero heat generation at 0.4 hfu. Roy and others (1972) and Lachenbruch (1968) interpreted this amount of heat to represent the mantle contribution beneath the Sierra Nevada, the excess above 0.4 hfu coming from radioactivity in the crust. This inferred mantle contribution is low compared to other regions and has been interpreted by Henyey and Lee (1976) to indicate the presence, now or in the geologically recent past, of a cold subducted slab at depth, which shields the batholith from heat rising from the mantle. 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 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 western and eastern margins of the batholith. Three other studied localities, 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.

The Jackson Butte and nearby Golden Gate Hll 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, 1971) 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 peridotites 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, 1986); the Big Creek andesite neck contains abundant eclogite and garnet granulite, less abundant peridotite and feldspathic granulite, and sparse peridotite (Dodge and others 1986); 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 (in press) have suggested that plagioclase-bearing peridotite xenoliths such as those at Blue Knob originated as the result of diapiric rise of Cr-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 non-feldspathic peridotites, pyroxenites, and gabbroids. Wilshire and others (in press) suggest that the mingling at Blue Knob occurred at the time the enclosing alkali basalt was erupted, about 3.5 m.y. years ago, but it seems equally possible that the mingling occurred during the Mesozoic.

The andesite pipe near Big Creek, which intrudes a megacrystic facies of the ~104 m.y. old Dinkey Creek Granodiorite, contains peridotites, both olivine-bearing and olivine-free pyroxenites, eclogites, several varieties of quartz-free metamorphic rocks, and partially fused gabbroids and granitoids (Domenick and others (1983; Dodge and Bateman, in press). Inclusions having sedimentary protoliths probably represent the refractory residue of crustal material that remained in the lower crust after the parent magma for the Dinkey Creek Granodiorite was generated. The abundant eclogite and sparse grospydite (carbonate-bearing grossular garnet-pyroxene rock) are especially significant because their presence suggests that ocean floor has been subducted beneath the root of the batholith. The 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.

$^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ range widely in the different xenoliths from the Big Creek locality and approximately duplicate the spread of values in the granitic rocks across the batholith. Dominick and others (1983) postulate 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 suggest that the granitic magmas were generated within the different depth zones and that their isotopic diversity reflects the compositions of these zones rather than simple mixing of different proportions of crustal and mantle material as has been suggested by DePaolo (1980).

Olivine-free orthopyroxenites, websterites, and clinopyroxenites are the dominant rock types contained in the toe of trachyandesite flow remnant at Chinese Peak, but orthogranulites (metagabbro) and paragranulites are abundant, and Fe-rich harzburgites and lherzolites are also present. Dodge and others (1986) interpret 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 at about the same time as the nearby ~104 m.y. old Dinkey Creek Granodiorite, although they do not precisely indicate that age. Rare garnet-bearing ultramafic xenoliths yield pressures of 13 kb, corresponding to a depth of about 43 km, but an absence of garnet or garnet pseudomorphs in most of the ultramafic xenoliths suggests lesser equilibration pressures. Equilibration pressures between 5 and 9 kb, equivalent to depths between 17 and 30 km, are more likely. Dodge and others (1986) interpret initial $^{87}Sr/^{86}Sr$ of the granitoids, which is intermediate between the ratios of the paragranulites and mafic and ultramafic xenoliths and those of the granitoids, to be the product of mixing of basaltic magma (represented by the mafic and ultramafics xenoliths) and crustal materials (represented by the paragranulites), followed by crystal fractionation.

Water compositions in springs and wells

Barnes and others (1981) determined the chemical composition of waters in soda springs in the Sierra Nevada, five of which are within the map area, and Mack and Ferrell (1979) determined the compositions 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 are in alluvium marginal to Mono Lake.

Mack and Ferrell (1979) have determined that 31 wells and springs in the western foothills have abnormally high concentrations (averaging 1,300 mg/L) of sodium chloride and contrast markedly with the good quality bicarbonate-rich water in several thousand other wells drilled in granitic rocks. Eighteen of these springs and wells are within the map area; the others are more than 30 km to the south. Fifteen of the wells are in the vicinity of Oakhurst; the others are at the Rancheria Fire Station, nearby Arnold Spring, and Shaver Lake Point. All of the wells within the map area are in the Bass Lake Tonalite, except for the well at Shaver Lake Point, which is in the Dinkey Creek Granodiorite. 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 lineament that trends N. 30° W. They suggested that the lineament marks the trace of the foothill suture (Saleeby, 1977) in metamorphic rocks, which within the map area would underlie the granitoids.

The 15 wells near Oakhurst lie within a topographic basin drained by the Fresno River, which is partly bounded on the southwest, south, and southeast sides by the Oakhurst roof pendant; Arnold Spring and the well at the Rancheria Fire Station are less than 2 km north of the Mountain View Peak roof pendant and downslope from a group of metamorphic remnants. These spatial relations suggest the possibility that conate water contained in the exposed metamorphic rocks flowed along joints and mixed with meteoric water to charge the springs. Metamorphic rocks have not been identified adjacent to the well at Shaver Lake Point, but sodium, chlorine, and dissolved solids are present in small amounts in this well.

Composition, age, and origin of the Sierran root

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

Christensen (1966) concluded that since 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 with these interpretations of a Cenozoic age, Bateman and Wahrhaftig (1966), Bateman and Eaton (1967), Schweickert and Cowen (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) have 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-day thick crust. 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 extension of 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 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 (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 P-wave velocities, but normal upper-mantle velocities of about 8.2 km/s. They suggested that a subducted slab of cold lithosphere beneath the Sierra Nevada, as postulated by Henyey and Lee (1976), shielded the Sierra Nevada from high regional asthenospheric heat flux and caused the present low heat flow.

Seismic measurements and 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 with a surface seismic-density layer (v_p = 6.0 km/s; ρ = 2.67 g/cm^3). Studies in the southernmost Sierra Nevada, where deeper crustal levels are reported to be exposed (Ross, 1983, 1985), indicate that a largely meta-igneous assemblage of hornblende-rich gneissic amphibolite- to granulite-grade rocks underlie this layer. These meta-igneous rocks form a roughly 10-km-thick lens (v_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 (v_p = 6.9 km/sec; 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 55 km may mark downward transition to an easterly-dipping, down-dragged slab of ocean-floor basalt that was transformed to eclogite (v_p = 7.9 km/s; ρ = 3.23 g/cm^3) (Henyey and Lee, 1976). Westward at shallower depths, the eclogite should grade to 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.

Basaltic volcanism shows that mantle-derived magma was introduced into the crust during the Cenozoic, but very much larger volumes must have been introduced during the Mesozoic to bring heat for the generation of the granitic magmas. Recognition of a mafic-ultramafic complex of Mesozoic age at depth beneath the Chinese Peak trachyandesite flow (Dodge and others, 1986) and a spread of isotopic ratios in 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 is related to the late Cenozoic extension and deformation affecting the Great Basin 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 acounts for the present configuration and height ot 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 drammatically 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 as compared with 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 under the Sierra Nevada is 7.9 km/s, generally about 7.8 km/s under the structurally high Basin and Range Province, and 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 asthenspheric heat flow that is affecting the Basin and Range Province but allows the upper mantle lithosphere beneath the plate 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 lead many people to think of the Sierra Nevada as a vast storehouse of valuable minerals. 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 of these deposits are located in the country rocks. The batholith served primarily as a "heat engine" that mobilized fluids which carried and concentrated metals dispersed in the country rocks. A major exception is the tungsten skarn deposits, which were formed by the reaction of magmatic fluids with carbonates in the country rocks.

Gold

Most of the production of gold in California has been from placer deposits, and probably all of 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 Coarse Gold, Fine Gold, and Mariposa Creeks. Placer activity in this area was intense within a few months of the discovery of gold at Sutter's mill but virtually ceased by 1870. Since then a few small operations have been undertaken, apparently without notable success. 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 4-mile 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 (1988) is in mineralized country rock. The veins dip steeply northeast and pinch and swell abruptly; rarely can they be followed for more than a kilometer. 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, is reported to still be yielding excellent specimens (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 Fe-Mg 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 Fe-Mg carbonate and subordinate sericite, quartz, albite, and pyrite.

The age of the Mother Lode veins has been uncertain. 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 older veins that were deformed during the development of schistosity are present (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 108 to 127 Ma (Kistler and others, 1983). However, 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 the isotopic ages of Böhlke and Kistler (1986) on samples from other Mother Lode veins.

The 114 Ma age. coincides 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 on 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.

The only large mine within the map area 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 ounces of gold at $20.67 per ounce) is much less than that of mines farther north. Most of this 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 (1,250 ft) and more than 2,700 m (9,000 ft) 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, 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 ounce, the price of gold during most of the period between 1835 and 1934, these amounts are equivalent to 162,000 and 133,000 to 145,000 ounces of gold, respectively. Both of these deposits are in quartz veins that lie within the western margin of the zone of carbonaceous metapelite that may be the site of the Calaveras-Shoo Fly thrust fault.

South of the western metamorphic belt, gold has been mined in the Coarse Gold septum, the Mountain View Peak roof pendant, and Hildreth Mountain. Several small deposits at Grub Gulch, in the western side of the Coarse Gold roof pendant, are said to have been worked as early as 1849. Exploration there was reported to have reached a depth of 250 m and production to have totalled about $960,000, equivalent to about 46,400 ounces 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 a meter 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 ounces 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 and other mines in the Pine Creek septum. The Cardinal mine, in a spur of the Bishop Creek septum, which extends to the north across the North Fork of Bishop Creek, is reported to have yielded $1,570,000, equivalent to about 76,000 ounces of gold, between 1910 and 1938. The ore body was 4 to 8 feet 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, 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 thick and was composed 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 somewhat more than 2,000 ounces of gold and 800 ounces 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 in Fish Spring Hill, south of Big Pine, are in the Jurassic Tinemaha Granodiorite. Most of the veins strike north of west and dip 20° to 50° north. 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 is 42,437 ounces of gold, 34180 ounces of silver, and 1,951 pounds of copper (Bateman, 1965a).

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 by product. 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 of the mines exploit sulfide-bearing quartz veins that occupy zones of intense shearing. The principal sulfides in the Jesse Belle, Green Mountain, and Lone Tree Mines are chalcopyrite, pyrite, and pyrrhotite, whereas pyrrhotite is the most abundant sulfide in the Buchanan and Daulton 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 the other mines as well, was enriched material from near the surface. The Buchanan mine was opened in 1863, during the Civil War, and operated continuously until 1909; the other mines probably were located about the same time. The Daulton mine was last worked between 1942 and 1945, during World War II, and the bulk of the production at the Jesse Belle mine also occurred during that period. Incomplete records suggest that the total production from the Daulton, Jesse Belle, and Buchanan mines was between 1.4 and 1.8 kg of copper plus some gold. The Green Mountain mine reportedly had 4,000 feet 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 western 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 ppm tungsten and fail to reveal any evidence of 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 ppm tungsten. The tungsten in these highly evolved rocks probably was carried in a vapor phase.

General features of skarns

The mineral assemblages present in most tungsten-bearing skarn deposits were formed during two distinct, but probably continuous or overlapping, stages of metasomatism, which followed nonadditive thermal metamorphism that accompanied the intrusion of the granitoids. Thermal metamorphism took place with rising temperatures and most of the metasomatism 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 contant.

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 the earliest stage of metasomatism. The early anhydrous metasomatism closely followed the nonmetasomatic metamorphism at 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 anhydrous skarns were formed after the granitic rocks had begun to solidify. The anhydrous skarns were formed at relatively high temperatures estimated to range between 550 and 650 °C (Morgan, 1975; Newberry and Einaudi, 1981) and at pressures in the range of 1.5 to 2.0 kb, equivalent to depths of 5 to 7 km (Newberry, 1982).

As temperatures decreased, the anhydrous garnet-pyroxene skarns were replaced, partially or completely, with hydrous minerals such as epidote and amphiboles, and by calcite. Both the anhydrous skarns and the adjacent granitic rocks were locally silicified, and quartz veins were formed. During the silicification, 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. During the period of quartz veining and silicification, sulfides were deposited--chiefly molybdenite, pyrite, pyrrhotite, chalcopyrite, and bornite, but including sphalerite, galena, and bismuthinite. Significant amounts of by-product copper and molybdenum are recovered at the Pine Creek mine.

The richest skarns were formed in relatively pure calcareous marble, which reacted readily with the mineralizing fluids. Iron-poor, light-colored calc-silicate minerals that were formed earlier during the nonadditive thermal metamorphism, such as wollastonite, plagioclase, grossularite, idocrase, and diopside, reacted with the ore-forming fluids much less readily than did calcite to form skarn that generally contains little or no tungsten. Rocks in which 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 almost barren 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 meters or more in thickness and from a few meters to as much as a hundred meters in outcrop length. Commonly skarn bodies are sporadically distributed along contacts between marble and granite, and between ore bodies only a thin layer of wollastonite may occur along the contact. Many ore bodies are localized at irregularities in the granite contact 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 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 Morgan Creek pluton of the Cretaceous Lake Edison Granodiorite. The Morgan Creek pluton is granite rather than 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 meters and have been exploited to depths of almost 1,000 meters.

Mineralogical and chemical relations

In the formation of skarn 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 skarns, but several interesting studies have been made of Sierran skarns. Significant studies of the mineralogic and chemical relations in tungsten skarns of the eastern Sierra Nevada are summarized below.

Donut ore body of the Pine Creek Mine

In their study of the Donut ore body in the northern part of the mine (Gray and others, 1968) (fig. 59), the geological staff at the Pine Creek Mine was the first to show that a skarn ore body is compositionally zoned. This ore body occupies the margins of a blocky inclusion of marble, which in plan view averages about 25 m across. Skarn 1 to 7 m thick occupies the margins of the inclusion. Most of the skarn is composed chiefly of garnet and pyroxene, but irregularly shaped masses of hornblende skarn penetrate inward from the margins into the garnet-pyroxene skarn. Epidote skarn is also present locally along the margins, and the adjacent granite 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 marble inclusion and its complete enclosure within granite, Gray and others (1968) considered a significant contemporary temperature gradient unlikely. Apparently they did not consider the effect of decreasing temperature with time in the formation of hornblende and epidote.

Pine Creek Mine

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

Newberry (1982) recognized the early nonadditive metamorphism of the country rocks, 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 zones; a thin inner zone adjacent to granite, 1 cm to 10 m thick, composed of qarnet+quartz±pyroxene; a main zone, generally 1 cm to 10 m thick, composed chiefly of garnet+pyroxene+scheelite; and a thin outer zone adjacent to marble, 1 to 40 cm thick, composed of idocrase+wollastonite+scheelite ±pyroxene. The boundary between the inner zone and the main zone is gradational and marked by the penetration of barren veins of iron-rich garnet from the inner zone into the main zone of garnet and pyroxene, whereas the boundary between the main zone and the outer zone is abrupt and marked by the replacment of garnet by idocrase.

Garnet and pyroxene in all of 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 25-55 percent andradite ($Ca_3Fe_2Si_3O_{12}$), 35-70 percent grossularite ($Ca_3Al_2Si_3O_{12}$), and 3-10 percent combined almandine ($Fe_3Al_2Si_3O_{12}$) + spessartine ($Mn_3Al_2Si_3O_{12}$) whereas those in the inner zone contain 20-80 percent of almandine + spessartine. Pyroxene in the main zone contains 15-40 percent diopside ($CaMgSi_2O_6$), 50-75 percent hedenbergite ($CaFeSi_2O_6$), and about 10 percent johannsenite ($CaMnSi_2O_6$), and the pyroxene in the inner zone contains about 30-35 percent diopside, 50 percent hedenbergite, and 15-20 percent johannsenite.

During this early anhydrous stage, both calc-silicate and pelitic hornfelses within and adjacent to the skarn were altered to pyroxene-dominant skarn that contains little scheelite. Thin, discontinuous layers within the skarn zones probably reflect impure layers within the marble. Between the inner skarn and the granite is a discontinous zone 1 cm to 1 m thick composed of massive quartz with traces of subcalcic garnet and alkali feldspar on the skarn side and hedenbergitic pyroxene and plagioclase with minor epidote on the granite side. Locally, especially in the upper parts of the mine, zones of massive quartz with minor garnet, plagioclase, and hedenbergitic pyroxene contain substantial amounts of molybdenite and lesser amounts of chalcopyrite. Gradational contacts of the massive quartz with both the skarn and the granite show that it replaced both rocks. Across a distance 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 grades, with distance from the contact, 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 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 quartz, chalcopyrite, scheelite, and minor 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 partly altered garnet knobs 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 epidote-calcite replacement of plagioclase and garnet, and amphibole-quartz±calcite replacement of pyroxene. Apatite, pyrite, sphene, and locally, scheelite are present in small amounts.

The lowest temperature alteration products are zeolitic zones of quartz, calcite, laumontite, chlorite, kaolinite, and native copper. These alteration zones cut across all older rock and form pipelike masses, one of which extends upward from 3,300 m altitude to the surface at 4,000 m altitude.

Metals were distributed somewhat differently during the later hydrosilicate alteration than during the earlier higher-temperature anhydrous metasomatism. During the early anhydrous metasomatism, scheelite was concentrated most abundantly in the outer skarn zone, adjacent to marble, where it occurs as small subhedral (0.2-0.4 mm) 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. Deposition, which was mainly in pyroxene-dominant skarn, may have been caused by reduction of the metasomatic fluids upon coming in contact with rocks containing high ferrous iron, or simply as the 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 are typically 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. The scheelite grains are low in molybdenum relative to the scheelite in anhydrous skarn. In some places, the retrograde alteration has resulted in the depletion of tungsten, for example, in biotite-dominant zones, but generally it is enriched. Although the precipitation of late 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 ion into the ore solutions during the relacement of calcium-rich calc-silicate minerals by calcium-poor hydrosilicate assemblages.

Strawberry mine

Nokleberg (1981b) has described in detail vertical compositional changes in a layer of skarn that replaces a steeply dipping bed of marble interstratified with plagioclase hornfels at ore body no. 7 of the Strawberry tungsten mine, northwest of Bishop on the western slope of the Sierra Nevada (fig. 60). From the lower contact with granodiorite upward, four different kinds of skarns are present: hornblende skarn, pyroxene skarn, garnet skarn, and wollastonite skarn adjacent to marble. 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 skarns in relation to one another are shown in figure 61. At each downward step, the most calcium-rich mineral disappears (fig. 61A): first calcite, followed successively by wollastonite, garnet, and pyroxene. Nevertheless, until they disappear, wollastonite, garnet, and pyroxene are increasingly abundant in each replacing skarn, and garnet and pyroxene are richer in iron.

CO_2 virtually disappears in the wollastonite skarn, and calcium diminishes in each underlying skarn (fig. 61B). Al_2O_3 and SiO_2 vary by only small amounts in all the skarns, except in hornblende skarn, where SiO_2 increases. MgO increases regularly downward. Both FeO and Fe_2O_3 are at a maximum in the pyroxene skarn and decrease both upward and downward. Fe_2O_3/FeO decreases downard. 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 (1981b) attributed the compositional pattern of these skarns to infiltration diffusion. Ore fluids moved upward and reacted with marble to form wollastonite skarn. Deposition of Si, Al, Mg, Mn and Ti and the removal of large amounts of CO_2 and Ca caused significant changes in the compositions of the ore fluids. After local equilibrium was attained and maintained, another wave of mineralizing fluids worked upward, which then 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 (1981b) considered and dismissed the possibility of a temperature gradient that decreased outward from the granodiorite contact, but he seems not to have considered the possibility that the temperature decreased with time as successive skarns were formed. Another possibility that seems not to have been considered is that early introduction of meteoric water and later depletion of the supply could account for the pattern of decreasing Fe_2O_3/FeO in the garnet, pyroxene, and hornblende skarns.

<u>Skarns in the Mount Morrison roof pendant</u>

Morgan (1975) studied four skarns in the Mount Morrison roof pendant north of Bishop 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 the carbonates and that the mixing ratios were strongly influenced by the local attitude of the contact. He plotted the ratio of ferrous iron to ferrous iron plus magnesium in pyroxene against the ratio of ferric iron to ferric iron plus aluminum in garnet and found that two of the deposits lie on an oxidizing trend and that two lie on a reducing 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 carried more meteoric water to the sides of the pluton than to the top, where magmatic fluids predominated.

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 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?

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 which can be tested--if not now, at least in the forseeable 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, published by Bateman and others (1963) was that the country rocks form a complex faulted synclinorium and that magmas generated in the lower crust as the result of radioactive decay rose and crystallized to form the granitic rocks. At that time, the thinking on orogenic problems was dominated by the tectogene concept of Griggs (1939) and Vening-Meinesz (1948). The principal evidence indicating a synclinorium consisted of west-younging strata 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). The root beneath the Sierra Nevada was conceived 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 west part of the batholith where the crust is thin and heat generation low, and because the reported predominance of eastward-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 Nevada. Both subduction zones came into operation in the Early Jurassic or Late Triassic and consumed intervening oceanic crust until they collided in the Late Jurassic, causing the Nevadan orogeny. Magma generation then ceased in both arcs, but eastward subduction farther west 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 of the source reservoirs for the magmas had been formed 1700 m.y. ago. Their studies showed that initial $^{87}Sr/^{86}Sr$ increases eastward from about 0.704 in the west to about 0.708 in the east. They interpreted the 0.706 isopleth to mark the western limit of the area that is underlain by Precambrian continental crust and 0.704 to be the eastern limit of oceanic crust. Later (Kistler and Peterman, 1978), they interpreted the 0.706 isopleth to also mark the eastern boundary of two stages of rifting that occurred between 1250-800 and 600-350 m.y. ago, and the 0.604 isopleth to mark the western boundary of mantle material introduced into the lower crust during these stages of rifting. They attributed variations of 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 mapped by Silver and Anderson (1974), trending N 50° W from northeast of Hermosillo in the Sierra Madre Occidental, Sonora, Mexico to the south end of the Inyo Mountains, continues northwestward through the central Sierra Nevada. The zone of disruption follows the axis of Jurassic magmatism farther south, and they suggest it is also the locus of Jurassic magmatism within and north of the map area. Silver and Anderson (1974) have estimated 700 to 800 km of left-lateral offset along this zone between mid-Triassic and mid-Jurassic time.

In contrast with the model of Kistler and Peterman (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 incompatable elements and the crustal component to be Paleozoic and Proterozoic sedimentary strata. He dismissed material carried down a subduction zone as 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 of incompatabilities of isotopic ratios; specifically, hydrothermally altered oceanic crust has high positive values of $\epsilon_{Nd}(T)$ (ratio of Nd isotopes to the ratio in a model chondritic reservoir at the time of emplacement, minus 1) whereas the granitic rocks have negative and low positive values.

DePaolo 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 ($\epsilon_{Nd} \simeq -12$) 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 paper, DePaolo (1980) had estimated that 50 to 70 percent of the Nd in some rocks, especially from the eastern side of the batholith, may have been derived from older crust, whereas some samples from the western side contain essentially no Nd from the crust and represent wholly material extracted from the mantle and added to the crust.

Saleeby (1978) proposed large-scale wrench movements and oblique subduction, more westerly blocks having moved northward relative to easterly blocks, which may also have undergone lesser amounts of northward movement. 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 in contact with late Paleozoic to early Paleozoic oceanic crust and ophiolite to the west. The oceanic crust includes ophiolite with 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) infer from the presence of a thick section of clean radiolarian chert that the ophiolite originated and remained for some extended time beyond the reach of continental derived sediments. Nevertheless, Saleeby (1982) later postulated that the 200 and 160 m.y. old ophiolites originated close to the continental margin in basins formed as the result of early Mesozoic accretion of late Paleozoic sea floor.

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

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 westward facing, essentially homoclinal sequences in the eastern belt of metavolcanic rocks. These sequences are part of the west-facing strata that form the eastern limb of the synclinorium in the geosynclinal model (Bateman and others, 1963). Some of the rocks have Early Cretaceous isotopic ages, so the faulting must have originated after the compressional events attributed to the Nevadan orogeny. They postulate it to have occurred as the result of regional extension during the time-span when the Cretaceous granitoids were emplaced. 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). The world-wide association of batholiths having the wide compositional range of the Sierra Nevada batholith with zones of convergence at continental margins leaves little doubt but that they are related in a fundamental way. Consequently, I assume that the batholith was emplaced within the margin of the North American plate above an eastward-dipping subduction zone that generated linear belts of heat and magma. 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.

Probably heat and magma rising from the mantle caused linear belts of uplift, extension, and faulting in the crust, which then exerted control over the rise of magma in the upper crust. The cross-cutting 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 evidence shows 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-travelled terranes within the area of the Mariposa 1° by 2° quadrangle.

Isotopic dating indicates that 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 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 p. 95-96) and 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.

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 basaltic magma 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 $^{87}Sr/^{86}Sr$ of 0.70371 for the eastern part of the greenstone of Bullion Mountain (Bateman and others, 1985) and of 0.7032 for two intrusions just west of the map area (pl. 1), the Guadelupe 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, has 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. Inward increase of initial $^{87}Sr/^{86}Sr$ in the margins of the Tuolumnne Intrusive Suite shows that the magma that formed the outer part of the suite continued to assimilate crustal material as it was emplaced.

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 $^{87}Sr/^{86}Sr$ of 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 $^{87}Sr/^{86}Sr$ significantly higher ($>$ 0.710 in some intrusions), and cogenetic tonalites and granodiorites, which would suggest differentiation from a more femic parent, are lacking.

With the rise of the magmas 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 here, crystal fractionation is the separation of crystals from the coexisting melt phase by such processes as sidewall acretion, 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, general marginal, 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 magma.

Acretion of crystals at the cooler margins of a magma chamber apparently are responsible for the inward compositional changes in the Tinemaha Granodiorite and the granodiorite of McMurray Meadows, where initial $^{87}Sr/^{86}Sr$ remains constant. The efficiency with which early precipitated crystals were concentrated in the margins of these rock, especially in the granodiorite of McMurray Meadows, suggests that these processes also were operative, and even may have been dominant, in the margins of the Tuolumne Intrusive Suite where isotopic data indicate mixing. The proposal of Sawka (1985) that sidewall acretion of crystals in the Tinemaha Granodiorite led to the upward streaming of the crystal-depleted, less dense adjacent magma, as proposed by Turner (1980) and McBirney (1980) provides a reasonable explanation for the downward increase in the density and mafic mineral content of 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 (figs. 38 and 39).

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

The planar foliation in intrusive rocks that contain mafic inclusions and/or tabular and prismatic minerals together with a virtual absence of evidence of lineation shows that the intrusions expanded as they were emplaced. Ballooning is especially evident where wall rocks wrap around rounded plutons and protrusions from larger plutons, but the planar foliation requires that virtually all plutons ballooned as they were emplaced. The ballooning intrusions both displaced wall rocks and produced cleavages and steeply dipping lineations within them, indicating thinning and upward translation. Ballooning of intrusions accounts for much of the deformation in the country rocks of the batholith, even though most intrusions show little or no evidence of ductile deformation within themselves. Only intrusions that continued to expand after crystallization had advanced far enough for them to sustain a shear stress show evidence of ductile deformation and lineation.

Although ballooning provided space for individual plutons, it is unlikely that the walls of the batholith were spread apart the distance required to accomodate the plutons. The different compositions (especially the K_2O content) of the Jurassic granitoids on the two sides of the batholith makes it unlikely that they were emplaced adjacent to each other and spread apart the full distance that now separates them during intrusion of the Cretaceous granitoids. Some intrusions appear to have made some space by stoping, but this mechanism merely redistributes material and does not result in increased space for the batholith as a whole.

Space for the batholith as a whole was provided in a variety of ways: by the incorporation of crustal materials in the magmas as is indicated by isotopic data, by the upward translation of wall and roof rocks and their erosion to depths of 3.5 to 7 km as is indicated by geobarometers, expulsion of volcanic materials, chiefly in ash flows and Plinian-type eruptions, and moderate extension normal to the axis of the batholith. Passage of the Independence mafic dike swarm through the crust, apparently without producing either 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.

A possible modern example of an incipient magmatic arc that may in time produce an intrusive suite like those that were formed during the Mesozoic is the zone that extends along the east side of the Sierra Nevada through Owens Valley, Long Valley, the Mono Lake basin, and beyond. Owens Valley, Long Valley, and the Mono Basin are down-dropped 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 the zone, may be analogous to the "basic forerunners" of the Sierra Nevada batholith, and felsic volcanism in the Mono and Inyo Craters, Glass Mountain, the Long Valley caldera, a rhyolite dome 11 km south of Big Pine, and the Coso volcanic field probably are underlain by felsic plutons. A pluton comparable in size to larger Sierran plutons is known to underlie the Long Valley caldera (Bailey and others, 1976; Hill and Bailey, 1985), and another pluton probably underlies 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 of initial $^{87}Sr/^{86}Sr$, decrease of initial $^{143}Nd/^{144}Nd$, and systematic changes in the abundances of various oxides and elements in the granitic rocks, especially eastward increase of K_2O, can reflect (1) eastward increase in the thickness of the prebatholithic crust, which would cause eastward increase in the proportion of crustal to mantle material in the parent magmas, (2) changes in the composition of the prebatholithic crust, or (3) 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 crustal margin. Moore (1962) has correlated increase of K/Na in Cenozoic volcanic rocks eastward across the Sierra Nevada with increase in the present crustal thickness, and Condie and Potts (1969) have postulated a general linear correlation between the potassium content of Cenozoic volcanics 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 and an absence of mafic precursors to indicate they are the products of crystal-liquid fractionation of more mafic magmas indicate that the source materials for these granites were largely crustal and similar to the nearby Precambrian rocks (table 2). Probably only selected 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. The refractory compositions of xenoliths derived from sedimentary protoliths in the trachyandesite intrusion near Big Creek (Domenick and others, 1983) and in the trachybasalt flow at Chinese Peak (Dodge and others, in press) 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 Solder 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 and others (1978a) argued that the Jurassic monzonites in the White and Inyo Mountains are distinct from Sierran rocks and must have had a mantle source because they contain far too much K_2O and Sr to fall on the trend of eastward increasing K_2O in Sierran granitoids. Initial $^{87}Sr/^{86}Sr$ for the Jurassic granitoids has not been determined, but probably is much lower than 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 may be less depleted in the east side of the batholith than in the west side.

Whether the initial strontium ratios of the surface rocks are representative of ratios at depth is uncertain. Dominick 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 ordered 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 and in the same range as the west to east isotopic variations across the batholith at the surface. Higher $^{87}Sr/^{86}Sr$ and lower $^{144}Nd/^{143}Nd$ in the Dinkey Creek Granodiorite than in any of the xenoliths, except for a sample of sillimanite gneiss, suggests that this inferred order in depth would represent the existing arrangement in 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. Expanded descriptions can be consulted in the referenced publications. The intrusive suites are discussed in terms of the three separate areas shown on plate 1: (1) west slope of the Sierra Nevada, (2) eastern Sierra Nevada and Benton Range area, and (3) White and northern Inyo Mountains area. For each area, intrusive suites are described first, in 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 the Q-A-P diagram shown in figure 6 (p. 46). Detailed descriptions of some rocks, especially in the Yosemite region, have not yet been published; consequently their descriptions are abbreviated.

Western 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 here 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 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 are shown in figure 63.

The suite is characterized by low contents of potassium and alkali feldspar. Generally alkali feldspar is interstitial to the other minerals, indicating it began to crystallize very late; alkali-feldspar megacrysts are absent. The low potassium content and initial $^{87}Sr/^{86}Sr$ of about 0.7045 indicate source materials that contained a preponderance of mantle 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 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 and by the granodiorites of Whisky Ridge and the Pick and Shovel mine area belonging to the Shaver Intrusive Suite. Stern and others (1981) reported a slightly discordant U-Pb zircon age of 113 m.y. on this rock.

Bass Lake Tonalite (Kbl)

Rock at the entrance to 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 southwest part of the map area (pl. 1) where the rock was called the tonalite of Blue Canyon. Because the name Blue Canyon is preempted in formal lithostratigrahic nomenclature and The Gateway does not appear on current maps and is poorly situated for the type locality of this lithodeme, the rock is here 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 rocks, 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 Range were mapped as separate lithodemes by Dodge and Calk (1986, 1987). Figure 63 shows the distribution and gives the names of the larger plutons that are separated from the main mass.

Typical rock 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 more than 50, but in most samples falls between 10 and 30 (fig. 65).

Darker 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 above 30 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 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) are included here in the Bass Lake Tonalite despite the U-Pb ages.

Northeast, east, and southeast of two plutons of the younger Ward Mountain Trondhjemite, the Bass Lake is deformed to gneiss. A secondary foliation concentric to the trondhjemite plutons and a radial lineation indicate that the deformation was caused by intrusion of the trondhjemite as a highly viscous crystal mush in a late stage of crystallization. This event has been referred to as the Blackhawk deformation by Bateman and others (1983).

The large size of the Bass Lake Tonalite suggests that it is not a single intrusion. The absence of obvious contacts, abrupt compositional changes, and discontinuities in foliation patterns seem to refute this possibility. Variations in the modal abundance of minerals in connection with foliation patterns and the distribution of metamorphic remnants suggest that at a slightly higher level of erosion the tonalite would appear as 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 follows roads that extend northeast from the western metamorphic belt through Raymond to the Coarsegold roof pendant (unpublished seismic reflection data, C.M. Wentworth, Jr., 1985). 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 about 3 km depth at the contact with metamorphic rocks to about 6 km, 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 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 on 13 samples of the Bass Lake Tonalite range from 124 to 105 Ma, but if the two lowest ages of 108 and 105 Ma and the three highest ages of 119, 119, and 124 Ma are omitted, the remaining 8 ages fall between 111 and 117 Ma and average 114 Ma, which is considered the optimum age of the Bass Lake (Stern and others, 1981: Dodge, 1986). Nevertheless, a large and complex intrusion such as the Bass Lake Tonalite probably cooled to below the blocking temperature over a period of several million years. K-Ar age determinations on biotite range from 91 to 114 Ma and on hornblende from 102 to 118 Ma (Bateman and Lockwood, 1976; Evernden and Kistler, 1970; Kistler, written commun., 1976).

Granodiorite of Arch Rock and other small intrusions of biotite granite and granodiorite (Kar)

The granodiorite of Arch Rock and several other compositionally and texturally similar intrusions, but which may not be exact equivalents of the Arch Rock, intrude the Bass Lake Tonalite (fig. 63). The rocks are medium to coarse grained and range in composition from biotite granite to biotite granodiorite and tonalite (fig. 64); a few are hornblende-biotite trondhjemite. The color index generally is between 4 and 12 but is higher or lower in a few rocks (fig. 65). The U-Pb age of the granodiorite of Sawmill Mountain is 116 m.y. 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 El Capitan Granite, and so its true age must be older. The granite of Hogan Mountain has a granoblastic fabric that continues into adjacent parts of the Bass Lake Tonalite, which indicates temperatures fell to below the granite solidus before emplacement was completed. Most of the intrusions in the southern part of the Bass Lake Tonalite were deformed when the Ward Mountain Trondhjemite was similarly emplaced a little later as a highly viscous crystal mush. The modes of these deformed rocks are highly variable, ranging from tonalite to granite.

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 here 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 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, gneissic, and when viewed in thin section 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 in depth. The attitudes of contacts and of gneissic foliations in the trondhjemite, and gneissic foliations and lineations in the adjacent metamorphic rocks and 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 intruded as a viscous, largely crystallized mush, which deformed the adjacent rocks as well as itself (Bateman and others, 1983a).

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 (Stern and others, 1981) and a K-Ar age on biotite of 106 Ma have been determined (Naeser and others, 1971).

Knowles Granodiorite (Kkn)

The Knowles Granodiorite, here formally named for the settlement of Knowles, is light gray, medium grained, and equigranular. In previous publications, this rock was designated informally as the granodiorite of Knowles (Bateman and others, 1983a; Stern and others, 1981). The name is formalized here because of the economic importance of the rock. 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. 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 area in the vicinity of the Raymond quarry is designated the type locality.

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 boundary between 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 similar rock north of Eastman Lake (called the plagiogranite north of Buchanan Lake by Bateman and Sawka, 1981 and the plagiogranite north of Eastman Lake by Bateman and others, 1982) and a small stock southwest of Rabbit Hill (called the granodiorite east of Hensley Lake by Bateman and Sawka, 1981 and Bateman and others, 1982), and on plate 1 and figure 63 these rocks are shown to be parts of the Knowles Granodiorite.

The Knowles Granodiorite intrudes the Bass Lake Tonalite and the Ward Mountain Trondhjemite and truncates gneissic structures formed as the result of intrusion of the trondhjemite. A U-Pb age of 111.5 Ma (Stern and others, 1981) is mildly discordant but is supported by K-Ar ages on biotite of 113, 107, 109, 110, and 113 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 El Capitan, is more than 30 km long and averages about 5 km wide. The rocks of several plutons in the northern part of the suite, which are here 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 and others, 1983b), 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 it is megacrystic or seriate and contains small (1-2 cm) alkali-feldspar megacrysts. Biotite, generally in small books, is the principal mafic mineral and is accompanied locally 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 20, but it reaches a maximum between 4 and 9. Primary magmatic foliation is weak, doubtless because of the paucity of mafic minerals. However, a secondary cataclastic 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 of 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 Mountain as being younger than the El Capitan Granite. The El Capitan has yielded two concordant U-Pb ages of 102 and 103 Ma, and a discordant age of 97 Ma (Stern and others, 1981; Stern, written communication, 1983). K-Ar ages on biotite of 94, 92, and 85 m.y. 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 here 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 12 but most 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, the Sentinel Granodiorite and the granodiorite of Yosemite Creek, which are possible early members of the Tuolumne Intrusive Suite, and 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. It was previously informally termed the Shaver granitoid sequence by Stern and others (1981). It is here 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 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 bullseye pattern. The other location is west of Huntington Lake 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, Kdcp)

The Dinkey Creek Granodiorite, informally referred to as the granodiorite of Dinkey Creek in earlier reports, is here 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, and 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 the northern part, where the rock is more felsic and inclusions of metasedimentary rocks are abundant, and the margin of the lobe that extends toward the southwest are megacrystic and contain 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 short distances, ranging from less than 5 in a belt of megacrystic and associated equigranular rocks, which extends northeast from the east end of Huntington Lake, to more than 25 south of the Dinkey Creek roof pendant (fig. 70). Variations of bulk specific gravity within the Dinkey Creek Granodiorite (fig. 38), which closely reflect compositional variations, are discussed beginning on p. 84.

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 of 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 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 sparse discrete euhedral hornblende prisms as much as 1 cm long and and biotite books as much as 0.5 mm across contrast with the small anhedral grains of these minerals in clots. These crystals are believed to have precipitated directly from the melt phase of the magma, perhaps obtaining some constituents from the clots.

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 age on zircon of 104 Ma has been obtained (Stern and others, 1981). Although this age is discordant, it agrees well with a U-Pb age of 102 Ma on 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 from the Dinkey Creek Granodiorite range from

101 to 80 Ma on hornblende and from 94 to 78 Ma on biotite (Bateman and Lockwood, 1976; Bateman and Wones, 1972a, b; Kistler and others, 1965).

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 than adjacent equigranular Dinkey Creek Granodiorite, ranging from 5 to 10 (fig. 70), 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 Pick and Shovel Mine Area (Kps) and granite of Ordinance Creek (Ko)

The granite of Ordinance Creek and the granite 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 glomerophenocrysts of biotite set in a felsic aplitic matrix. Similar dark spots elsewhere are associated with miarolitic cavities and appear to reflect the appearance of 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 pluton has a highly irregular shape and the smaller one 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 between 1 and 12 and averages 6 (fig. 70). Most modes plot in the granite field on a Q-A-P diagram, but a substantial number plot in the granodiorite field (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 rocks 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, Shorthair 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, and a 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 aerially 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 intruded 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 ages on biotite 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), respectively, 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 indicates that all of these K-Ar ages have been reset, probably as the result of intrusion of the younger John Muir Intrusive Suite.

Possible additional units 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 assigning it to the Shaver Intrusive Suite is a concordant U-Pb age on zircon of 103 Ma which is the same age within the limits of analytical error as those on the Dinkey Creek Granodiorite 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 ranges from to 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 the pluton. The suite is shown to comprise six successively younger map units on plate 1 and figure 71: (1) the quartz diorite dikes on El Capitan; (2) the granodiorites of Illilouette Creek and Tamarack Creek, the tonalite of Crane Creek, and the Leaning Tower Granodiorite; (3) the granodiorite of Ostrander Lake; (4) the granodiorite of Breeze Lake; (5) the Bridalveil Granite; and (6) the granite of Chilnaulna 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, diorite 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 that intrude 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 marginal 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 complex of dikes, 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 (fig. 72) that contains 15 to 50 percent of biotite and hornblende (fig. 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 unassigned 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 plot modes fall in the granodiorite and adjacent parts of the granite fields (fig. 72). Most samples contain 5 to 15 percent of 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 that of the ~102 Ma old El Capitan Granite and of the ~98 m.y. old 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 fine 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. The granodiorite of Horse Ridge generally is a little darker than the Bridalveil Granodiorite, and the modes of a few samples plot in the tonalite and quartz diorite fields on a Q-A-P diagram (fig. 71).

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 Horse Ridge mass of Bridalveil Granodiorite 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 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 less than 10 km^2 (fig. 71). The older leucogranite porphyrites of Red Peak and Post Peak, which Peck (1980) asssigned 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 metamorphic fragments are so abundant as to form intrusive 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 granite 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 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. 73). 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 has been disrupted and split into two parts by a tongue of the Half Dome Granodiorite of the Tuolumne Intrusive Suite, but a crudely nested pattern is recognizable in which successively more felsic units are present toward the interior (fig. 71).

Granodiorite of Red Devil Lake (Krd)

The granodiorite of Red Devil Lake is the marginal and most mafic rock of the intrusive suite of Washburn Lake. It forms 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). It 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 is 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 in the core unit 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 in comparison with other intrusive suites, are closely associated spatially, have similar isotopic ages, and all intrude the only slightly older volcanic 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 age Ma on the granodiorite of Red Devil Lake and the 100 Ma age of the Minarets volcanic sequence (Peck, 1980; Kistler and Swanson, 1981; Stern and others, 1981; Fiske and Tobisch, 1978). 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 that of the Shaver Intrusive Suite and the coeval intrusive suite of Yosemite Valley and that of the Tuolumne and approximately coeval 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, being 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. 34) and are concentrically arranged--the oldest and most mafic units occupy 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 p. 81). The changes at contacts are merely steps in progressive changes from the center to the core of the intrusive suite. 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 has removed little of the adjacent older unit, textural and compositional changes across the contact are slight, but where a magmatic surge was transgressive and removed a significant part of an earlier 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. 33). 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 marginal parts 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 marginal parts 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. 33). These masses may have been continuous when they were emplaced and torn apart and partly reincorporated in magma during the surge that brought in the Half Dome Granodiorite.

These marginal rocks are dark gray, and their color index ranges from about 10 to 30, decreasing inward (fig. 75). Compositions range from quartz diorite to granodiorite (fig. 74). The rocks are equigranular and range from fine to medium-grained. Because of the equant shapes of grains, foliation is not readily visible in the tonalite of Glen Aulin, but it is shown by lensoid inclusions in the granodiorite of Kuna Crest. 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 mm across, and apatite occurs as tiny (0.2-0.5 mm) prismatic crystals. Some magnetite is present as equant crystals, and some titanite occurs as skeletal 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 marginal rocks intrude all of the plutonic rocks with which they are in contact that do not belong to the Tuolumne Intrusive Suite, showing that the Tuolumne Intrusive Suite is the youngest in the region. They are intruded by the Half Dome and Cathedral Peak Granodiorites. 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

Half Dome Granodiorite (Khd, Khdp)

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 at contacts with their inner parts. 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 ranges generally from 5 to 15 in the equigranular facies and from 2 to 10 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, 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 a kilometer, then remain constant farther inward, both at about 25 percent.

Tabular alkali-feldspar megacrysts about 1 cm thick and 2-3 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 amount of alkali feldspar is difficult, even when counts are made on large glacially polished surfaces, because the texture of the rock is seriate and no sharp cutoff between megacrysts 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 marginal rocks of the Tuolumne Intrusive Suite and is intruded by the Cathedral Peak Granodiorite. It has not been dated isotopically by the U-Pb method, but concordant U-Pb ages of 88 and 91 Ma on the granodiorite of Kuna Crest, 88 Ma on the tonalite of Glen Aulin, and of 86 Ma on the Cathedral Peak Granodiorite bracket its age closely.

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. At the contact of the Cathedral Peak Granodiorite with the megacrystic facies of the Half Dome Granodiorite, megacrysts are crowded together in swarms, and their abundance cannot be accurately determined. However, about 1 km inward from the contact the Cathedral Peak contains about 12 percent megacrysts. The size and abundance of megacrysts decreases inward, and at the contact with the Johnson Granite Porphyry only a few scattered megacrysts half the size of those at the outer contact with the Half Dome 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 are distinctly less abundant than in the Half Dome and form smaller crystals. Modal plots on a Q-A-P diagram range across the boundary between the granodiorite and granite fields (fig. 74). The color index ranges between 1 and 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.

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 Johnson intrude the Cathdral 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 bordered with medium-grained rock identical with the groundmass of the contiguous Cathedral Peak, and scattered angular fragments of plagioclase and quartz approximately of 2 to 4 mm across are set in an extremely fine-grained groundmass (less than 1 mm) of quartz, alkali feldspar, and plagioclase. The fine-grained groundmass indicates that the magma was quenched, the angular plagioclase and quartz fragments indicate comminution, and the miarolites 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 near-anhydrous minerals until it exceeded the amount soluble in the melt. The appearance of a separate volatile phase would quickly increase the volume of the system beyond which the wall and roof rocks could adjust without fracturing (Burnham, 1972).

Modally, the Johnson contains more alkali feldspar than the contiguous Cathedral Peak Granodiorite (fig. 84), and chemically it contains more K_2O and less CaO, Fe_2O_3, FeO, MgO TiO_2, P_2O_5, and MnO. Its minor-element content includes small amounts of V, Zn, and Sr. These chemical differences also extend into adjacent parts of the Cathedral Peak Granodiorite, showing that it too was affected, probably by circulating volatiles.

Possible earlier 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, they 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 here left unassigned.

Granodiorite of Yosemite Creek (Kyc)

The granodiorite of Yosemite Creek, called the Yosemite Creek Granodiorite of Rose by Kistler (1973) and Bateman and others (1983b), 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 (fig. 76), and the color index shows a corresponding wide range. Locally, the rock is porphyritic and 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)

D.L. Peck (written commun., 1982) mapped a contact within the Sentinel Granodiorite previously mapped by Calkins (1930) and Calkins and others (1985), that separates a large area, including the type locality at Sentinel Rock, from tonalite at Glacier Point (map unit Kk), which is clearly part of the Tuolumne Intrusive Suite (pl. 1). Typical Sentinel Granodiorite 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. 84). The color index ranges from 4 to 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 marginal 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 here 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 the Mount Morrison roof pendant and the Wheeler Crest and Pine Creek septa on the east and the Potter Pass septum on the west. 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 of the larger plutons have long, narrow shapes and trend 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 northerly than the others. All of the U-Pb isotopic ages for units of the suite are about 90 Ma.

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 is relatively leucocratic; the southeast end, which is almost detached from the main mass, is also 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 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 south 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 a centimeter 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 in the range of An_{45} to An_{30} but is more sodic in the felsic south end of the pluton where the average range is 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 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 is 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,Kmgp)

The Mount Givens Granodiorite, one of the larger plutons in the Sierra Nevada, comprises about 1,400 km^2. The pluton is elongate in a northwesterly direction, having a length of 80 km and an average width of 17 km across the southern two-thirds of the body, and almost twice that distance across a bulge 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 fall in the tonalite field (fig. 78). The Mount Givens consists of two facies, an equigranular granodiorite facies and a megacrystic facies that is mostly granite but includes some 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 granite forms several elliptical domains in the central and southern part of the pluton. The color index ranges from about 5 to 25 in equigranular granodiorite and from 2 to 12 in the megacrystic granite and granodiorite (fig. 79).

Variations of modal composition and specific gravity within the pluton are shown in figure 41. 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 more felsic rocks, granite and some granodiorite, are weakly foliated or unfoliated. The different compositional and textual variants generally grade to one another. Along a traverse from the margin to the core of the bulbous head of the pluton (fig. 40), which was studied in detail by Bateman and Nokleberg (1978), equigranular biotite-hornblende tonalite in the margin grades inward through equigranular hornblende-biotite tonalite and hornblende-biotite granodiorite to megacrystic biotite granodiorite and granite in the core. This inward gradation is interrupted at a horseshoe-shaped internal intrusive contact. Megacrystic granite on the outside is intruded by fine-grained equigranular granodiorite on the inside, which grades inward to megacrystic granite. Details of the compositional zoning in the bulbous head are given beginning on p. 85 in the section dealing with compositional zonation.

Hornblende and biotite in the hornblende-bearing rocks are generally subhedral or anhedral, and alkali-feldspar megacrysts in the megacrystic rocks exhibit few crystal faces and are poorly formed. 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 a meter 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 area of granite is in the bulbous head, and several smaller areas are present in the southern part of the pluton. The rounded shape of the bulbous head probably resulted from the the lesser density and greater viscosity of the siliceous magma, which approached the viscosity of the wallrocks and caused swelling 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 a granitic core.

The Mount Givens Granodiorite intrudes all of the granitoids along its west and north borders except the granodiorites of Red Lake and Eagle Peak, which are considered possible additional members of the John Muir Intrusive Suite. On the east it is intruded by the granodiorite of Lake Edison 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 on two samples of the Mount Givens are 88 and 93 Ma (Stern and others, 1981). The 88 Ma age is concordant and the 93 Ma age slightly discordant. K-Ar ages are 88 and 89 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 (~103 Ma) of the adjacent Shaver Intrusive Suite on the west (late Early Cretaceous).

Lake Edison Granodiorite (Kle)

The lake Edison Granodiorite is here formally named for exposures at its type locality, which is along the shores of the upper end 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 swells at both ends. A narrow tongue, called the Morgan Creek mass by Bateman (1965a), 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 importance because it is genetically related to the important skarn-type tungsten deposits of the Pine Creek roof pendant.

Typical rock of the Lake Edison Granodiorite is fine- to medium-grained, equigranular hornblende-biotite granodiorite, distinctly finer grained than the Lamarck or Mount Givens Granodiorites. Titanite is generally abundant. The color index 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 rock exhibits good foliation that strikes parallel to adjacent contacts and is vertical or dips steeply. The bulge at the north end of the pluton is concentrically zoned, and the color index falls 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; the color index decreases somewhat irregularly toward the east to less than 6, and hornblende is absent. This bulge was originally called the Basin Mountain mass of the Tungsten Hills Quartz Monzonite (Bateman, 1965) (here 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 would have 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 has 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 77 and 82 Ma and 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 (Krd)

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 because of its lesser density and developed an intrusive contact with 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 country rocks and older granitoids; the two rocks are in contact only at the south end of the largest and most southerly pluton, the Round Valley Peak pluton. The Round Valley Peak pluton is an arcuate body that encloses a lobe of the younger Mono Creek Granite on the west. Before intrusion of the Mono Creek Granite, this pluton probably had a linear shape with its long axis parallel to those of the Lamarck and Lake

Edison Granodiorites. The other two plutons lie farther northwest, where they intrude the Mount Morrison roof pendant.

Typical rock is notably equigranular and medium grained. Except for being 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. Plagioclase commonly is in equant to slightly elongate grains, which together with biotite and hornblende give the rock a "tidy" look (Bateman, 1965a). Some plagioclase grains contain mottled cores in the compositional range of An_{49} to An_{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 16 and averages about 11 (fig. 79).

The Round Valley Peak pluton is strongly foliated and compositionally zoned. In the eastern margin, adjacent to older rocks, strongly flattened mafic inclusions that appear elongate in outcrop are abundant. These inclusions give the rock a strong foliation. The marginal rock 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. The rock in the smaller masses enclosed in the Mount Morrison roof pendant 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 here 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, 1965) crops out farther south in the range front east of Mount Humphreys; another small mass (called the quartz monzonite of Turret Peak by Lockwood and Lydon, 1975) lies southwest of the south end of the main mass. The rock resembles the Cathedral Peak Granodiorite of the Tuolumne Intrusive Suite but generally is a little finer grained.

Typical rock is megacrystic, and blocky alkali-feldspar megacrysts averaging 1.4x2.5 cm as seen in outcrops are set in a medium-grained groundmass. The color index of most samples falls in the range of 5 to 16 (fig. 79). On a Q-A-P diagram, modes plot in both the granite and granodiorite fields, but more are in the granite field (fig. 77). A weak foliation parallel to the long axis of the pluton is shown 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 consistent changes in specific gravity (fig. 80). The specific gravity in the margins is greater than 2.65, whereas in the core area it ranges from of 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 part of the main pluton the volume percent of megacrysts decreases systematically inward from about 10 percent in the margins to as little as 3 percent in the core (fig. 8). An exception to this pattern is a protrusion of equigranular leucogranite from the east side of the pluton. A northwest trending zone of high values of specific gravity and volume percent of 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 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 roof pendant the pluton is bordered by a swarm of pegmatite and aplite dikes that dip gently into the pluton and disappear into the granite. 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 in the adjacent wallrocks that were caused by upward movement and expansion of the viscous Mono Creek magma (Bateman and others, 1965a). Kerr (1946) suggested that these dikes carried tungsten from the last crystallizing core of the Mono Creek Granite to form the rich tungsten skarn deposits in the west side of the Pine Creek roof pendant, 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 concordant U-Pb ages of 90 and 93 Ma (Stern and others, 1981) and K-Ar ages on hornblende of 79 m.y. and on biotite of 79 to 82 Ma (Evernden and Kistler, 1970).

<u>Evolution Basin Alaskite (Kev)</u>

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 here 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 Mount Goddard septum on the west. Like the Lamarck, it forms an elongate body but is much smaller, being about 20 km long and averaging about 5 km wide. A spur of metamorphic rock from the Mount 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 lay 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 and others, 1984a), indicating 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 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 m.y. Independence dike swarm. It is intruded by the unassigned granodiorite of Cartridge Pass and by dikelike bodies of fine-grained aplitic granite. K-Ar ages on biotite of 85, 80 and 79 m.y. have been reported by Evernden and Kistler (1970).

Small intrusions west of the Mount Givens Granodiorite

Three groups of small intrusions lie west of the Mount Givens Granodiorite. 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 a detailed study by Noyes and others (1983a, b) 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 rock 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, and the plot of modes on a Q-A-P diagram shows only moderate spread (fig. 78); most modes fall within the granodiorite field. 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. At the northern segment of the contact, the Red Lake clearly intrudes the Mount Givens, but the southern segment appears gradational, indicating a close relation between the two. 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 road cuts along State Highway 163 just south of the Big Creek crossing. Along the northwest side of the highway, flat-lying marginal dikes of the granodiorite of Red Lake can be seen to intrude the Dinkey Creek Granodiorite. Two K-Ar ages on biotite of 90 and 84 Ma have been determined on 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, and the core is megacrystic granite. Both hornblende and biotite are present. The color index ranges from 5 to 11 and generally is a little higher in the margins (fig. 78). 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. 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, but over most of the pluton it strikes northwest and is vertical or dips steeply. At the northwest end, the foliation appears to continue into a mass of Dinkey Creek Granodiorite. This relation indicates 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 intrusive rock 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 Shorthair Creek of the Shaver Intrusive Suite. A single K-Ar age on biotite of 89 Ma has been reported (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 Granodiorite 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 10 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 north margin are massive, whereas the southeast margin and both the east and west ends are foliated parallel to the external contacts of the pluton.

A K-Ar age on biotite 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 and others (1983b) concluded, chiefly from studies of the minor-element distribution, that the parent magmas for the two rocks 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 between 2 and 3 and a uniform composition and texture (fig. 79). Modes form a tight cluster in the granite field on a Q-A-P diagram (fig. 78). Most of the rock is massive, 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 marginal felsic dikes penetrate the adjacent Dinkey Creek Granodiorite. Bateman and Wones (1972b) reported a K-Ar age on biotite of 91 m.y. 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 in the suite 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 of the rocks is generally less than 6 (fig. 79), but is higher along some contacts with tonalite. Modes of samples of the leucogranite of Big Sandy Bluffs and Lion Point are scattered across the granite and granodiorite fields, whereas four modes of the leucogranodiorites of Burrough and Black Mountains plot in the granodiorite and tonalite fields (fig. 78).

Granitic rocks of the western 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, have Middle Jurassic and Late Jurassic ages, respectively. The older tonalite of Granite Creek has been deformed, presumably during the Nevadan orogeny, whereas the granodiorite of Woods Ridge is undeformed except locally close to contacts with other rocks.

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 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 parallels foliation 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 ranges between 15 and 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.

Granite of Woods Ridge (Jwr)

The granite of Woods Ridge is medium- to light-gray, conspicuously megacrystic, splotchy-appearing rock. Alkali feldspar forms erratically distributed megacrysts of variable size and shape. Most larger megacrysts are tabular, but some smaller crystals appear to have angular shapes, probably because of their diverse orientations relative to the exposed surfaces. Biotite is in tiny flakes. Modes are scattered on a Q-A-P diagram (fig. 82) in the granodiorite and granite fields, reflecting the erratic distribution of alkali-feldspar megacrysts. The color index ranges from 5 to 15 and averages about 8 (fig. 83). The granodiorite 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 centered south of the map area, but three lobes extends northward across the south boundary of map area. The most westerly 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 most easterly 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 on the geologic maps of the Millerton Lake and Raymond quadrangles in the tonalite of Blue Canyon (Bateman and others, 1982; Bateman and Busacca, 1982), most of which is here assigned to the Bass Lake Tonalite. The rock south of Black Mountain was suspected of being older than rocks here 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 on the Bass Lake Tonalite, a sample from the north shore of Millerton Lake was isotopically dated. This sample yielded a concordant U-Pb age of 134 Ma (Table 1).

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 here 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. The larger mass has an irregular but roughly ellipsoidal shape and underlies an area of about 6 km^2; the smaller 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 rock ranges widely in composition, from leucogranite porphyry to granodiorite porphyry (fig. 82). It intrudes the El Capitan Granite and has yielded a Rb-Sr whole-rock age of 108 ± 4 m.y. and therefore cannot be included with younger porphyries assigned to the volcanogenic Minarets sequence.

Tonalite of Aspen Valley (Kav)

The tonalite of Aspen Valley forms a small pluton with an irregular shape in the northwest 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 ranges from 10 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 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 30 (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. Typical rock is equigranular and is characterized by euhedral hornblende and biotite crystals. The color index ranges widely from about 3 to 23 (fig. 83). The Bearup 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. Both rocks are medium-grained, equigranular hornblende-biotite granodiorite and tonalite (fig. 82), but they differ in composition and appearance. The granodiorite of Grizzly Creek has a color index that ranges between 10 and 30 (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 rocks intrude the granite of Shuteye Peak of the Shaver Intrusive Suite.

Granodiorite of Camino Creek (Kca)

The granodiorite of Camino Creek is fine- to medium-grained rock of variable composition and texture. Typical rock is hornblende-biotite granodiorite, but ranges from quartz diorite 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 Mount Goddard septum

Granitic rocks of four different compositions intrude the Mount 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. The granodiorite of Goddard Canyon, the leucogranite of Hell For Sure Pass, and the fine-grained quartz syenite are sheared pervasively. This shearing continues into the metamorphic rocks of the roof pendant, indicating a period of regional deformation. 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 on the largest mass of fine-grained quartz syenite, they are assigned to the Jurassic.

Fine-grained quartz syenite (Jqs)

This fine-grained leucocratic rock forms several long thin bodies, the largest of which is 10 km long and 1 km wide. Modes are widely scattered on a Q-A-P diagram, but most modes plot in the quartz syenite field or adjacent parts of the syenogranite field (fig. 82); the color index is highly variable (fig. 83), probably because of strong alteration. The ragged texture does not appear igneous, and it is possible that the potassic composition of this rock reflects secondary enrichment. 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 elongate when it was intruded, but undoubtedly shearing has modified its shape. The rock is strongly altered and contains much chlorite, epidote, and other alteration products; during mapping it was called the "cruddy granodiorite." The small number of 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 is a medium-grained equigranular rock with a color index that rarely exceeds 3 (fig. 82). 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 Mount Goddard Roof pendant. 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, most modes plot within the granite field, but some plot in the granodiorite field (fig. 82); the color index generally ranges between 1 and 8 (fig. 83). An observation by Lockwood and Lydon (1975) that the contact of the fine-grained rock with volcanic rocks is gradational suggests that th 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 Mount Goddard septum and 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. The plutons are separated from one another by the younger Mount Givens Granodiorite and the leucogranite of Graveyard Peak and originally may have been 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 and Mono Creek Granite of the John Muir Intrusive Suite and by the leucogranite of Graveyard Peak.

Leucogranite of Graveyard Peak (Kgp)

The leucogranite of Graveyard Peak is medium grained equigranular, 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 forms several small bodies 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 and Mono Recesses Granodiorites. 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 Mount Goddard septum (Kwd, Kfp, Kmr, Ktu, Kub)

Southwest of the Mount Goddard septum, along and close to the south boundary of the map area is a group of five intrusions that cannot at present be assigned to intrusive suites (fig. 84). The intrusive relations at ontacts 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 suggests that the granodiorite of upper Blue Canyon and the granite of Tunemah Lake may belong to the same suite and that the granodiorite of Mount Reinstein and the leucogranite of Finger Peak may both belong to a younger suite. The spatial relations of the Mount Givens Granodiorite to the granodiorite of Mount Reinstein and the leucogranite of Finger Peak are quite remarkable. 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 within the map area from other plutonic rocks by a metamorphic septum, but south of the map area it intrudes and is intruded by several plutonic rocks (Moore, 1978). Modes of samples from these intrusions and their color indices are shown in figure 85. Petrographic descriptions have not been published, and no isotopic ages have been determined.

Granitic rocks of the eastern Sierra Nevada and the Benton Range

Scheelite intrusive suite

The Scheelite Intrusive Suite is here 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, and apparently is not present there. It undoubtedly continues northward beyond the map area. 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 underlies about 3,000 km^2, mostly as a single unit, the Wheeler Crest Granodiorite. The Tungsten Hills Granite underlies about 250 km^2 at the south end of the suite, and the granite of Lee Vining Canyon underlies about 200 km^2 at the northwest end.

Scattered U-Pb ages on these rocks allow the possibility that the Wheeler Crest Granodiorite is a few million years older than the Tungsten Hills Granite and the granite of Lee Vining Canyon. The most likely age of the Wheeler Crest Granodiorite is ~214 Ma whereas the 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. Although the two granites are separated about 50 km, 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 (₨wc)

In earlier publications, outcrops in the eastern escarpment of the Sierra Nevada were formally called Wheeler Crest Quartz Monzonite (Bateman, 1961, 1965a; Rinehart and Ross, 1964; Huber and Rinehart, 1965); identical rocks just east of the Sierra Nevada were informally called quartz monzonite of Wheeler Crest (Rinhart and Ross, 1957), and those farther east in the Benton Range were called the granodiorite of the Benton Range (Rinehart and Ross, 1957; 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 southwest part of the quadrangle to the quartz monzonite of Wheeler Crest and outcrops in the northeast part to the granodiorite of the Benton Range even though they considered the two units equivalent. The name Wheeler Crest Granodiorite is here applied to all of 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, megacrystic, and has a color index that ranges between 5 and 20 and averages about 10. 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 decrease in the abundance of megacrysts. The texture of equigranular rock is identical with 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 ubiquituous. Rocks having a color index less than 10 commonly contain little or no hornblende, and rocks in which the color index is above 10 generally contain more hornblende than biotite. The general compositional range of zoning in plagioclase is from about An_{40} to An_{20}, but the range of different grains varies widely. Mottled cores of An_{40} to An_{45} and above 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 borders larger grains, indicating that some of the foliation may be secondary, probably the result of forcible intrusion of 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 and small bodies of diorite and gabbro. It is 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 m.y. on a different size fraction) and of 207 Ma and discordant ages of 201 and 161 Ma (Stern and others, 1981; Chen and Moore, 1982). Several K-Ar ages on hornblende range from 211 to 98 Ma (Kistler and others, 1965; Evernden and Kistler, 1970; Crowder and others, 1973). The younger hornblende K-Ar ages were determined on samples collected from Wheeler Crest and are thought to have been reset during intrusion of the nearby Cretaceous John Muir Intrusive Suite; the three oldest K-Ar ages of 211, 211, and 199 Ma are from localities far from younger intrusions. K-Ar ages on biotite range from 215 to 71 Ma.

Granite of Lee Vining Canyon (₨ lv)

The granite of Lee Vining Canyon forms a long, narrow pluton that occupies the lower slopes of 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 marginal zones exposed along Lee Vining Canyon and west of Grant Lake. The texture varies from medium-grained 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; 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 laden with recrystallized and partly digested inclusions and that the texture is aplitic.

The granite of Lee Vining Canyon is intruded by the granodiorite of Mono Dome, the granite 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 report a concordant age of 210 Ma on 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 was originally called the Tungsten Hills Quartz Monzonite by Bateman (1965), who showed it to be much more extensive than it is on plate 1. The rocks here excluded are those west of the Pine Creek septum and their southward continuation around and south of Mount Humphreys, which have been reassigned to the Lake Edison Granodiorite of the John Muir Intrusive Suite. The lithologic designation of the Tungsten Hills is changed to reflect the current IUGS classification (Streckeisen, 1973).

Typical Tungsten Hills Granite is medium-grained, seriate or weakly megacrystic, light-colored rock with a color index in the range of 3 to 10 and averaging about 6 (fig. 85). Mafic inclusions are scarce, and mineral grains show no evidence of preferred orientation; hence a foliation is not apparent. Subhedral to anhedral grains of quartz, alkali feldspar, and plagioclase are present in almost equal amounts. Plagioclase is zoned in the general range of An_{34} to An_{20}, not including rims of subsolidus albite. Mottled cores were not observed. Biotite is the principal mafic mineral and locally is accompanied by minor 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 m.y. and a discordant age of 167 Ma (Stern and others, 1981; Chen and Moore, 1982). K-Ar ages on biotite of 76 and 77 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 here formally named for exposures along and west of Palisade Crest, a part of the Sierra divide west of Big Pine. The type area is the eastern escarpment of the Sierra Nevada south of Big Pine Creek 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 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 include the leucogranites of Red Mountain and Taboose Creeks (fig. 86). The Inconsolable Quartz Monzodiorite (formerly the Inconsolable Granodiorite; Bateman, 1965a) was originally included in the suite (Bateman and Dodge, 1970), but isotopic dating by the Rb-Sr method indicates that this intrusion has a Cretaceous age (R.W. Kistler, written communication, 1983).

The granodiorite of McMurry Meadows is nested within the Tinemaha Granodiorite and is the younger unit. Although the grain size and texture of these units are different, both are characterized by a slightly higher content of alkali feldspar than of quartz, wide but overlapping ranges in the abundance of plagioclase, and the presence of augite cores in much of the hornblende. Although both rocks are called granodiorite, few modes of either rock 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 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, undoubtedly it is continuous beneath the alluvial deposits of Owens Valley with exposures in Fish Springs Hill and the Inyo Mountains (pl. 1). The rock has much the same appearance everywhere despite a wide range of composition, especially in the plagioclase content. Typical rock is seriate to weakly porphyritic and contains blocky anhedral to subhedral alkali-feldspar megacrysts as much as 1 1/2 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 14. Quartz generally is slightly less abundant than alkali feldspar, and plagioclase is present in highly variable amounts, producing a modal plot on a Q-A-P diagram that is elongate away from the plagioclase corner (fig. 86). The rock is called granodiorite, although modes fall in the monzodiorite, granodiorite, quartz monzonite, and granite fields.

Plagioclase is commonly 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 west side and horizontal or gently dipping foliation in the east side, 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 155 Ma (Stern and others, 1981), and 164 Ma (Chen and Moore, 1982), K-Ar ages on hornblende of 187, 184, 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 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 it is older than 148 Ma. Compositional variations within the Tinemaha Granodiorite and their origin are discussed on p. 87-88.

Granodiorite of McMurry Meadows (Jmc)

The granodiorite of McMurry Meadows forms an elliptical pluton that is intrusive into and enclosed by the Tinemaha Granodiorite. It underlies an area of about 30 km^2 and is strongly zoned, grading through a narrow transitional zone from quartz monzodiorite in the margins to granite in the interior. Thus, it is bimodal, the margins being quartz monzodiorite and the interior granite (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 and conspicuous alkali-feldspar megacrysts. The plagioclase in rocks of both compositions is highly zoned in the range of An_{50} to An_{20}. Tiny cores in the range of 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 (Jrm) and Taboose Creeks (Jtb)

The closely related leucogranites of Red Mountain Creek and Taboose Creek 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 intruded, the leucogranite of Taboose Creek probably was entirely enclosed within and at least partly overlain by the leucogranite of Red Mountain Creek (Bateman, 1965a).

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. In fact, the leucogranite of Taboose Creek consists of two rocks, one of which intricately intrudes the other in narrow, irregular branching dikelets, perhaps reflecting the loss of volatiles. In outcrop and hand sample, the younger rock appears lighter colored than the older rock, but the two rocks are indistinguishable under the microscope.

On the north, northwest, and southwest sides, the plutons are bordered by schist, and the disposition of roof remnants of schist and Tinemaha Granodiorite strongly indicates that before erosion schist overlaid the leucogranites just above the present level of exposure. The enclosed position of the leucogranite of Taboose Creek suggests that it 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 map area, the leucogranite is intruded by younger granitoids. The Jurassic age of these intrusions is the only criteria that suggests they may 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, 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. Probably they 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. An 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 (1965a) 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 20 (fig. 87). Mafic dikes, presumably of the Independence dike swarm, are locally abundant. Shear zones are common and affect the dikes as well as the granite. The granite of Chickenfoot Lake is intruded by the Morgan Creek mass 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. The rock is medium-grained biotite leucogranite with a very low color index (fig. 87). It closely resembles the leucogranites of Rawson Creek and Graveyard Peak, but it is separated by many kilometers from these rocks and has a much older concordant U-Pb age of 161 Ma (Stern and others, 1981).

Inconsolable Quartz Monzodiorite (Kin)

The Inconsolable Quartz Monzodiorite, formerly called the Inconsolable Granodiorite by Bateman (1965), is here renamed because of changes in the classification system since it was originally named (Streckeisen, 1973). It forms an elongate pluton that lies northwest of the Tinemaha Granodiorite along the Sierran crest. Almost all of 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 results partly from an average color index ranging from 12 to 30 and partly from the prevalent gray to dark-grayish-red color of the feldspar. In many places, especially near the southeast margin, small but conspicuous reddish plagioclase grains are scattered throughout the rock. The rock contains abundant mafic inclusions that are progressively flatter 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 either, but the amount of plagioclase is quite variable, causing elongation of the modal plot toward plagioclase (fig. 87). Plagioclase is zoned in the average range of An_{50} to An_{30}, and mottled cores having higher An content are common. Both quartz and alkali feldspar are interstitial to plagioclase. Biotite predominates among the mafic minerals, but both augite and hornblende are present. Augite appears to be more abundant than hornblende, which indicates 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. It also is in contact with the ~164 Ma Tinemaha Granodiorite of the Palisade Crest Intrusive Suite, but the contact is featureless, and the relative ages of the two units have not been determined by field observations. Nevertheless, the Inconsolable Quartz Monzodiorite has been determined to have an Rb-Sr whole-rock age of 105±11 Ma and is therefore Cretaceous and the younger rock (R.W. Kistler, unpub. data, 1983). Until this determination was made, the Inconsolable Quartz Monzodiorite was considered to be the oldest unit of the Jurassic Palisade Crest Intrusive Suite.

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 westward into the high country west of Coyote Flat, including Mount Alice. The rock is equigranular and medium grained; the grain size generally ranges between 3 and 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, fall within the range of An_{30} to An_{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 observatons. 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 83, 87, and 89 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. These are light- to dark-gray medium-grained rocks. The granodiorite of Mono Dome grades to tonalite, quartz monzodiorite, and quartz diorite (fig. 87), which contain 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. The granodiorite of Tioga Lake is in contact only with metamorphic rocks. R.W. Kistler (written commun., 1983) has determined a whole rock Rb-Sr age of 93+6 Ma for the granodiorite of Mono Dome.

Granite of June Lake (Kjl)

The granite of June Lake was mapped originally as part of the Wheeler Crest Granodiorite (Huber and Rinehart, 1965; Kistler, 1966a) but is now recognized as a separate intrusion. 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 is younger than 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 largest of these are the granite of Mono Lake and the leucogranite of Ellery Lake. Neither of these rocks nor the leucogranite of Williams Butte has 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 poor, and its age relative to the ages of these rocks has not been determined. Presumably the leucogranite of Ellery Lake is the younger rock. The leucogranite of Williams Butte intrudes the granodiorite of Mono Dome, but the granite of Mono Lake intrudes only metamorphic rocks.

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 of 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 along the east side of the Bishop Creek septum and the Cartridge Pass along the south boundary of the map area just west of the Sierra Nevada divide. Most of the granodiorite of Cartridge Pass lies south of the map area. Both plutons are compositionally zoned. The Coyote Flat pluton has a marginal zone that is darker and finer grained than the interior, and the Cartridge Pass pluton is continuously zoned to a granite core (Moore, 1963). The two modally analyzed samples from the Cartridge Pass pluton and the three from the Coyote Flat pluton that plot closest to the plagioclase corner on a Q-A-P diagram are from marginal parts of the plutons (fig. 87). Close grouping of the other modes from the Coyote Flat pluton 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. 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 grades outward toward the grain margins to compositions as sodic as An_{20}.

Both plutons appear to intrude all of 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 age on biotite 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 northern Inyo Mountains area

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 afinities 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 G. H. Anderson (1937) proposed that his Boundary Peak Granite, which included all of 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 granodiorite of Cabin Creek as well as the granite of Pellisier Flats. Emerson (1966) showed that the granites of the northern White Mountains solidified from magmas and has convincingly degranitized them.

Soldier Pass Intrusive Suite

The Soldier Pass Intrusive Suite is here formally named for Soldier Pass, and the area between Buckhorn Spring on the southeast side of Deep Spring 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, the monzonite of Joshua Flat, the granodiorite of Beer Creek, and the Cottonwood Granite crop out within the map area where they underlie an area of about 25 km^2; the others 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 uniquely among the granitoids of the map area, in these rocks 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 are both included in the quartz monzonite of Beer Creek (Nelson, 1966b; 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, coarse-grained megacrystic rock on the north and fine- to medium-grained, equigranular, gray rock on the south. Modal analyses of the samples he collected confirm the existence of 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 the monzonite of Joshua Flat occupies small areas on either side of Deep Spring 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 feldspars generally are in nearly equant but ragged grains, whereas most of the quartz is interstitial, indicating that it was the last mineral to begin to crystallize. The modes of 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). However, Sylvester and others (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 is accompanied locally by small amounts of biotite. Titanite and opaque minerals are abundant.

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 and others, 1978a), which has yielded a U-Pb age of 179 Ma. Stern and others (1981) reported a concordant U-Pb age of 167 Ma on the Joshua Flat, Sylvester and others (1978a) reported concordant U-Pb ages of 173 and 178 Ma on different fractions of the same sample, and Gillespie (1979) reported a concordant U-Pb age of 159 Ma. The rocks has also yielded K-Ar ages ranging from 157 to 175 Ma on biotite and from 172 to 188 Ma on hornblende (McKee and Nash, 1967). The true age probably is ~170 Ma.

Granodiorite of Beer Creek (Jbe)

The granodiorite of Beer Creek, formerly part of the quartz monzonite of Beer Creek (Nelson, 1966b; Krauskopf, 1971), is medium gray, fine to medium grained, and in outcrop and hand specimen has a notably ragged texture. 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 western side of the intrusion and 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, and the color index ranges from 8 to 16. Much of the quartz is interstitial, indicating late crystallization. In the eastern facies, quartz is evenly distributed as subequant grains, the color index ranges from 4 to 10, and biotite generally is the only mafic mineral. Both facies contain abundant titanite and opaque minerals. The texture and composition of the western facies suggest it is transitional between the eastern facies and the monzonite of Joshua Flat, and indicate a relationship between the granodiorite of Beer Creek and the monzonite of Joshua Flat.

The interstitial habit of quartz in the western facies indicates that quartz was just beginning to precipitate from the melt phase when crystallization was nearing completion, whereas its greater abundance and uniform distribution in subequant grains in the eastern facies indicates that crystallization was more advanced there. These considerations suggest that the rock in the western facies crystallized in 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 age on zircon of 161 Ma on the granodiorite of Beer Creek. Several K-Ar ages on biotite ranging from 155 to 174 Ma and on hornblende from 162 to 180 Ma have also been reported (Crowder and others, 1973; Evernden and Kistler, 1970; McKee and Nash, 1967). Some of these determinations undoubtedly were on samples of the Cottonwood Granite.

Cottonwood Granite (Jcc)

The Cottonwood Granite was originally called the Cottonwood Porphyritic Adamellite by Emerson (1966). The name is here changed to the Cottonwood Granite to reflect the nomenclature of the IUGS classification (Streckeisen, 1973) (See fig. 6). Its name probably was derived from the upper reaches of Cottonwood Creek north of lat 37° 30' N., which is also presumed to be the type locality. Typical Cottonwood Granite is light gray, medium to coarse grained, and has a seriate to porphyritic 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 generally foliation is weak or absent. Mafic and 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 Creek and Indian Garden Creek. Stern and others (1981) have determined a concordant U-Pb age on zircon from this rock (incorrectly reported to be from the granodiorite of Beer Creek) of 168 Ma and a discordant age of 172 Ma.

Unassigned Jurassic plutonic rocks

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 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 of Krauskopf, 1971, Crowder and Ross, 1973, and Crowder and Sheridan 1972, and to the Barcroft Granodiorite of Emerson, 1966) occurs in two masses that are separated by the younger granite of McAfee Creek. Its dark gray color is caused by a high color index that ranges from 7 to 30 and a bluish-gray color of 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 were formed from hornblende. Crowder and Ross (1973) reported little pyroxene in the western part of the larger mass, but Emerson (1966) reported augite in about 20 percent of the samples he examined from the eastern part of the western mass and from the eastern mass.

The quartz monzonite of Mount Barcroft is intruded only by the Cretaceous granite of McAfee Creek. Stern and others (1981) and Gillespie (1979) determined U-Pb ages on zircon of 161 and 165 Ma, respectively. Similarity of isotopic age and rock composition suggest that the quartz monzonite of Mount Barcroft is related to the Palisade Crest Intrusive Suite 30 km farther south.

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. Typical rock is medium-gray, medium-grained biotite-hornblende granodiorite, but locally the rock is megacrystic. Like the quartz monzonite of Mount Barcroft, which it resembles, quartz and the 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_{33} to An_{39} (Emerson, 1966). These features suggest a close relation between the Mount Barcroft and 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 granite of McAfee Creek, which now separates them, was emplaced. 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) have determined K-Ar ages of 88 Ma on biotite and of 153 Ma on hornblende.

Granite of Sage Hen Flat (Jsf)

The granite of Sage Hen Flat (Sage Hen Adamellite of 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) has reported a U-Pb age on zircon of 144 Ma and Crowder and others (1973) reported K-Ar ages of 133 Ma on biotite and 141 Ma on hornblende.

Cretaceous granitic rocks of the White and northern Inyo Mountains

A cluster of intrusions of biotite granite crop out in the northern White Mountains, and three isolated plutons farther south 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, Leidy Creek, McAfee Creek, east of Dyer, 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 Creek and Marble Creek 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. These outward dips contrast with the steep or vertical dips of contacts of most intrusions in the Sierra Nevada. They indicate that the White Mountains granitic rocks have been less deeply eroded since they were emplaced than most of the granitoids in the Sierra Nevada, which have vertical or steep contacts, and that only their upper parts are exposed.

Barton (1987) showed that the megacrystic granite of Birch Creek is peraluminous, and Emerson (1966) and Crowder and Ross (1972) 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. None of the other Cretaceous granites of the White Mountains has been studied isotopically, but they are compositionally similar to granites farther east and northeast in the Basin and Range Province, which DePaolo and Farmer (1984) report have initial neodymium and strontium isotopic ratios similar to the ratios in adjacent Precambrian rocks at the the time the granites were emplaced.

Granite of Pellisier Flats (Kpe)

The granite of Pellisier Flats underlies an area of about 125 km^2 in the north end of the White Mountains. Typically, it is an equigranular medium-gray rock with a color index that ranges from 3 to 25 (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 an untidy appearance. Despite the dark color, much of the rock contains abundant alkali feldspar, in places in grains large enough to produce a submegacrystic texture. Modes are extremely variable and on a Q-A-P diagram form a large plot in the granite, syenogranite, quartz monzonite, and quartz monzodiorite fields (fig. 88). Compared with 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 rock elsewhere and almost no hornblende.

On the west the granite dips gently to moderately westward under metamorphic rocks whereas, on the east the contact is variable and dips toward the east in some places and toward the west in others. A zone of mixed metavolcanic and felsic granitic rocks occupies an area of about 25 km^2 in the southern part of the intrusion. This zone has been interpreted to be a fragmented and possibly foundered remnant of the roof rocks, which was intruded by the granite of Pellisier Flats and drenched with 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 in the southern part of the granite support this interpretation. The core facies in the northern part probably represents the deepest exposures.

The texture of the granite of Pellisier Flats everywhere is granoblastic, indicating that the rock has been strongly deformed, and 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 intruded. The age of the Pellisier Flats remains in doubt. If D_1 foliations in metavolcanic rocks along the west side of the White Mountains, which Hanson and others (1987) report are older than the ~160 Ma quartz monzonite of Mount Barcroft, continue into the quart monzonite of Pellisier Flats, the quartz monzonite of Pellisier Flats is of Jurassic rather than Cretaceous age as shown in this report. A discordant U-Pb age of 90 Ma (Stern and others, 1981), two K-Ar ages on hornblende of 92 and 100 Ma (Crowder and others, 1973), and an age of 157 Ma on biotite from a sample collected at the north end of the White Mountains (Evernden and Kistler (1970) have been reported for the Pellisier Flats.

Granites of Boundary Peak (Kwb), Marble Creek (Kwm), Leidy Creek (Kwl), McAfee Creek (Kwa), east of Dyer (Kwd), and Indian Garden Creek (Kwig)

All of these rocks are leucocratic biotite granites that have varying textures but only slightly different compositions (fig. 94). Peraluminous compositions and $^{87}Sr/^{86}Sr$ greater than 0.71 for the granite of Boundary Peak indicate that unlike the the Jurassic granitoids, the source for the Cretaceous magmas were Precambrian crustal rocks like those exposed to the east and northeast. Probably all the Cretaceous granites were intruded during a brief time span. Crowder and others (1973) reported K-Ar ages on biotite from these granites ranging from 73 to 87 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 is medium grained, equigranular, and massive. It is elongate toward the northwest and on the west end and southwest sides dips gently beneath the granite of Pellisier Flats, The trace of the contact on the northeast side across rugged topography indicates that it dips steeply or is vertical.

Toward the southeast, beyond a belt of metamorphic rocks, is the granite of Marble Creek. The position and shape of this intrusion along the northeast flank of the White Mountains 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, which may be part of the same intrusion. West of the granite of Marble Creek and separated from it by metamorphic rocks is the small granite of Leidy Creek (called the Leidy Adamellite by Emerson, 1966), which is fine to medium grained and 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, felsic granite of McAfee Creek. The northeastern part of this granite, adjacent to the granite of Marble Creek, is weakly megacrystic, but most of the granite is equigranular. The 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 is equigranular, fine- to medium-grained felsic rock (Emerson, 1966). The average grain size is 1 to 2 mm. In addition to its indeterminate contact with the granite of McAfee Creek, it intrudes the Jurassic Cottonwood Granite. Crowder and others (1973) have reported K-Ar ages on biotite from the granite of Indian Garden Creek of 95 and 121 Ma. These ages are distinctly older than those of any other granite in the northern White Mountains.

Megacrystic granites of Papoose Flat, Birch Creek, and Redding Canyon (Kpp)

The granites of Papoose Flat and of Redding Canyon 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 is also similar except that it is megacrystic only locally. Although here called granite, on a Q-A-P diagram, modes of the Birch Creek cross the boundary of 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 of the rocks contain biotite, and muscovite is present locally. The accessory minerals, apatite, zircon, and opaque minerals, occur in clusters with biotite, epidote, and chlorite (Sylvester and Nelson, 1966; Sylvester and others, 1978b; Nelson 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 southern ends of major anticlines, and both expanded westward, displacing and disrupting the adjacent strata to the west. The western margins of both plutons have been ductilely deformed, the grain size reduced, and the rock strongly foliated and lineated. K-Ar ages on biotite of 78 and 82 Ma have been determined for the Birch Creek pluton (McKee and Nash, 1967) and of 80 Ma on the Papoose Flat pluton (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 Birch Creek pluton was emplaced in the near-vertical east limb of the anticline along a bedding plane within the Reed Dolomite and expanded toward the northwest, bowing the adjoining strata outward and forming faults and secondary northeast-trending folds; then bending them 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 and others, 1978b). 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 of orthopyroxene can be seen.

Diorite and gabbro (dg)

Small bodies of dark fine- to coarse-grained rocks are widely distributed within the map area but are especially abundant on the two sides of the batholith. Compared with the granitoids, these mafic rocks have been neglected during most studies of the Sierra Nevada batholith, and a definitive study of them is yet to be made. Wes Hildreth (written commun., 1986) suggested that many of the diorites and gabbros fed basaltic and andesitic stratovolcanoes that were subsequently largely destroyed as the enormous reservoirs of granitic magma grew, which gave rise to ash-flow fields.

Many of the masses of diorite and gabbro intrude or are spatially related to masses of metamorphic rock. The interior of the batholith, occupied by the Shaver Intrusive Suite and the Mount Givens Granodiorite of the John Muir

Intrusive Suite in the Sierra National Forest area and by the Tuolumne Intrusive Suite in the Yosemite area, is almost devoid of plutons of diorite and gabbro. Mayo (1941) referred to these masses as "basic forerunners," and it appears to be true that most of them are older than the granitic rocks with which they are in contact. Nevertheless, some masses of diorite have been shown to be younger than the contiguous granitic rock, as in the massive face of El Capitan in Yosemite Valley, and it seems probable that the intrusive relations of some other masses have been misinterpreted. The temperatures of gabbroic and dioritic magmas could have been high enough to mobilize earlier solidified granitic rocks and cause ambiguous intrusive relations.

The compositions of these dark 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 An 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 rocks exhibit unusual textures. Layering is common, and many layered sequences exhibit "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 is the result of flow sorting (Bagnold, 1954; Bhattacharji and Smith, 1964). Some hornblende gabbros contain prisms of hornblende an inch or more long and less than half as wide. In other gabbros, large hornblende crystals are poikilitic and enclose small plagioclase grains. Some tabular bodies are composed of hornblende needles that lie in a plane but are randomly oriented within the plane. The possibility that some of these textures are the result of recrystallization at the time of emplacement of later enclosing granitic rock has not been evaluated.

With increasing amounts of biotite, quartz, and alkali feldspar, the hornblende gabbros grade to diorite, quartz diorite, and mafic tonalite. The color index in these rocks generally ranges from 40 to 20. Plagioclase compositions range from from An_{50-40} in the inner parts of grains to An_{30-20} in the margins, but mottled cores as calcic as An_{60} have been identified. 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 that amount.

<u>Mafic dikes</u>

Swarms of mafic dikes cut many of the granitic rocks but are restricted in their distribution. Mafic dikes may be common in one part of an intrusion but absent elsewhere. They rarely cut felsic rocks and are absent in some entire suites, such as the Tuolumne Intrusive Suite. Mafic 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. Generally the dikes cut across foliation in the host rock and have sharp walls, indicating that the dikes were emplaced after consolidation of the host granitic rock. However, a few "synplutonic" dikes appear to have been deformed with the enclosing granitic rock during a late stage of consolidation. Dikes of several different ages have been recognized in the central Sierra Nevada, and this relation together with the "synplutonic" dikes leads to the supposition that the exposed dikes represent late intrusions of mantle of lower crustal material, most of which intruded, partially melted, and mixed at depth with lower crustal rocks to produce the parent magmas for the granitoids.

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. Contacts between the felsic material and the mafic dike rock are generally crenulate, suggesting the coexistence of two magmas. 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 uncertain. It is possible that some, and perhaps most, of the felsic material in irregular veins was derived from selective melting of host rock fragments that were incorporated in the mafic dike.

The mafic dikes are of several ages as would be expected if they were intruded coevally with various host granitoids. However, the Independence dike swarm (Moore and Hopson, 1961; Chen and Moore, 1979), the oldest, most extensive, and best known swarm, appears to have been intruded when no major granitoids were emplaced. This swarm is about 4 km wide and extends southward from near Bishop in the eastern Sierra Nevada and White Mountains into the Mojave Desert. Dikes of this swarm intrude Triassic and Jurassic granitoids of the eastern Sierra Nevada and are cut off by Cretaceous intrusions. Chen and Moore (1979) reported that dating of three silicic dikes, which they interpret to be part of the swarm, gave concordant U-Pb ages of 148 Ma. Many dikes of this swarm are strongly sheared, especially along their margins. The great extent of the swarm and the apparent absence of coeval granitoid intrusions suggest that it was emplaced during a period of crustal extension and may have broken through to the surface.

The Lamarck Granodiorite of the John Muir Intrusive Suite is one of the Cretaceous intrusions that cuts off the mafic dikes of the Independence dike swarm. However, the Lamarck Granodiorite is intruded by younger mafic dikes, which are cut off, in turn, by the Lake Edison Granodiorite of the same suite. Other major intrusions of Cretaceous age that are 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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Table 1.--Previously unpublished U-Pb age data

Sample number----	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
$^{208}Pb/^{232}Th$	147.6	
Parts per million		
Pb	14.6800	4.0653
U	444.6900	189.4189
Th	425.4700	0.0000
Atomic ratios		
$^{208}Pb/^{206}Pb$	0.432620	0.110470
$^{207}Pb/^{206}Pb$	0.102450	0.060360
$^{204}Pb/^{206}Pb$	0.008670	0.000820

Table 2.--Major-element oxide analyses of country rocks within and east of the Sierra Nevada

[Isotopic data on strontium and neodymium from Kistler and Peterman (1973) and DePaolo (1981). T=100 Ma]

	FD-32, 33, 34, 42		FD-38	FD-39, 46		BM	FD-56
	Range	Average		Range	Average		
SiO_2	52.7-66.1	60.5	56.5	74.6-74.9	74.8	72.7	40.0
Al_2O_3	14.4-15.4	15.0	15.0	11.6	11.6	6.4	7.1
Fe_2O_3	2.9-4.4	3.5	4.1	1.4-2.4	1.9	.63	1.0
FeO	2.5-5.5	3.7	4.8	1.3-2.0	1.7	1.7	.88
MgO	2.8-5.9	3.8	5.3	1.5-1.8	1.7	2.2	8.5
CaO	2.2-6.7	4.4	6.3	.76-.94	.85	9.0	18.4
Na_2O	2.1-3.4	2.6	2.2	.95-1.5	1.0	.32	.46
K_2O	.94-1.6	1.3	.96	2.6-2.7	2.7	2.0	1.3
H_2O+	2.2-3.4	2.7	2.3	1.6-2.9	2.3	.62	1.2
H_2O-	.30-.80	.62	.52	.21-.47	.34	.04	.19
TiO_2	.60-.90	.64	.85	.53-.55	.54	.36	.38
P_2O_5	.16-.17	.17	.16	.09-.14	.12	.15	.21
MnO	.09-.17	.12	.13	.06	.06	.00	.07
CO_2	<.05-.08	<.05	<.05	<.05	<.05	3.5	20.5
$^{87}Sr/^{86}Sr_{(T)}$	.7041-.7059	.7047	.7048	.7120-.7135		.7132	
$\varepsilon Sr_{(T)}$		-9		+60 - +131			
$\varepsilon Nd_{(T)}$		+4.6		-2.2 - -10.8			

	FD-57 to FD-63		PC-1 to PC-9		PC-10 to PC-25		PC-27
Average	Range	Average	Range	Average	Range		
SiO_2	33.3-73.2	58.5	59.8-88.3	70.0	48.0-77.2	60.0	66.9
Al_2O_3	10.6-16.4	13.6	5.8-17.5	14.2	4.1-26.2	14.5	15.0
Fe_2O_3	.78-3.8	1.3	.65-1.9	1.1	.90-2.9	1.9	2.0
FeO	1.1-3.2	2.5	.68-7.0	3.7	2.2-12.3	5.3	3.6
MgO	1.0-6.0	2.7	.26-4.6	1.7	2.6-5.5	4.2	1.3
CaO	.55-23.2	6.7	.20-1.9	.94	1.6-8.5	4.7	3.3
Na_2O	.71-2.2	1.6	.09-4.1	1.6	0.6-3.6	1.8	3.2
K_2O	2.2-3.9	3.3	2.2-4.6	3.9	.60-6.7	2.6	2.3
H_2O+	1.3-3.0	1.2	.83-3.6	1.9	.64-3.7	2.3	.72
H_2O-	.15-.24	0.3	.06-.18	.10	.12-.70	.28	.08
TiO_2	.46-1.3	0.9	.27-1.4	.71	.44-1.6	1.0	.99
P_2O_5	.08-.20	0.8	.04-.21	.11	.08-.28	.13	.15
MnO	.04-.07	.06	.00-.36	.10	.04-.23	.11	.07
CO_2	<0.5-19.4	5.8	.06-.65	.19	.01-1.9	.57	.02
$^{87}Sr/^{86}Sr_{(T)}$						.7150	.7270

	PC-30a to PC-31 Range	Average
SiO_2	60.9-83.8	69.2
Al_2O_3	6.3-19.1	14.2
Fe_2O_3	.60-3.6	2.1
FeO	1.8-4.1	2.6
MgO	.90-3.0	2.0
CaO	.58-1.5	1.1
Na_2O	.87-2.6	1.4
K_2O	1.6-3.3	2.4
H_2O+	.83-3.2	2.4
H_2O-	.14-.28	.21
TiO_2	.49-1.1	.70
P_2O_5	.00-.09	.04
MnO	.04-.08	.06
CO_2	.06-.78	.31
$^{87}Sr/^{86}Sr_{(T)}$	.7360	
$\varepsilon Sr_{(T)}$		+450
$\varepsilon d_{(T)}$		15.6

Ages and locations of samples:
All samples are chip samples that were collected across lithologic units or groups of units.
FD-32, 33, 34, and 42: Mesozoic volcanic and plutonic strata west of the Melones fault zone along California State Highways 108 and 132, and along Merced County road 116.
FD-38: Greenstone of Bullion Mountain along California State Highway 140.
FD-39 and FD 46: "Calaveras" strata along California State Highways 120 and 140.
BM: Pennsylvanian and Permian(?) strata in Bloody Mountain block of the Mount Morrison roof pendant.
FD-56: Paleozoic ranging in age from Middle Cambrian to Permian exposed in Mazurka Canyon, southern Inyo Mountains.
FD-57 to FD-63: Proterozoic and Early Cambrian strata in the White Mountains.
PC-1 to PC-9: Precambrian rocks from Pleasant Canyon, Panamint Range. PC-1 and PC-7 are of sheared granite. The others samples are of sedimentary rocks of the Proterozoic Pahrump Group.

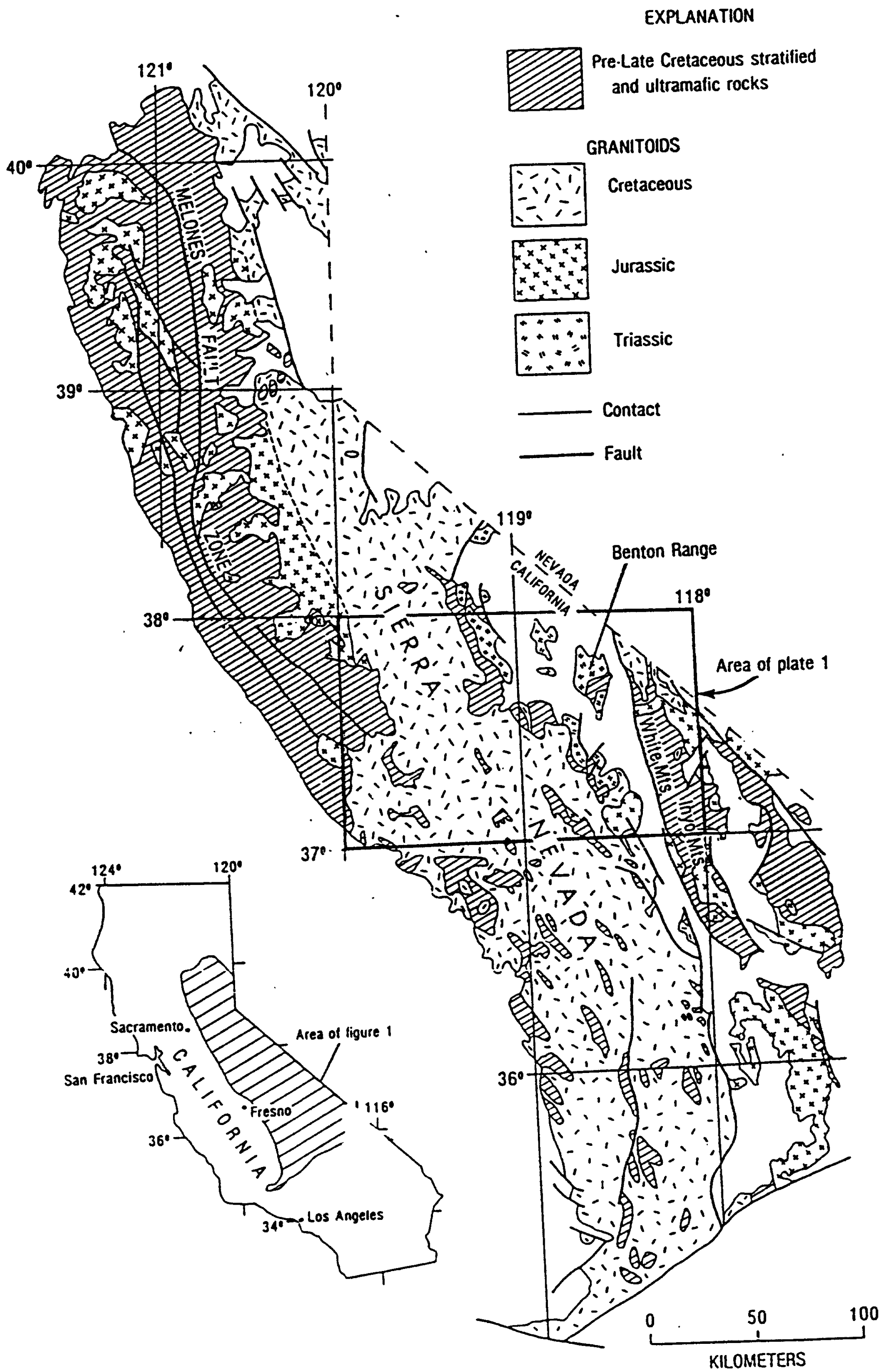

FIGURE 1.--Map of Sierra Nevada and adjacent areas in eastern California showing location of Mariposa 1° by 2° quadrangle.

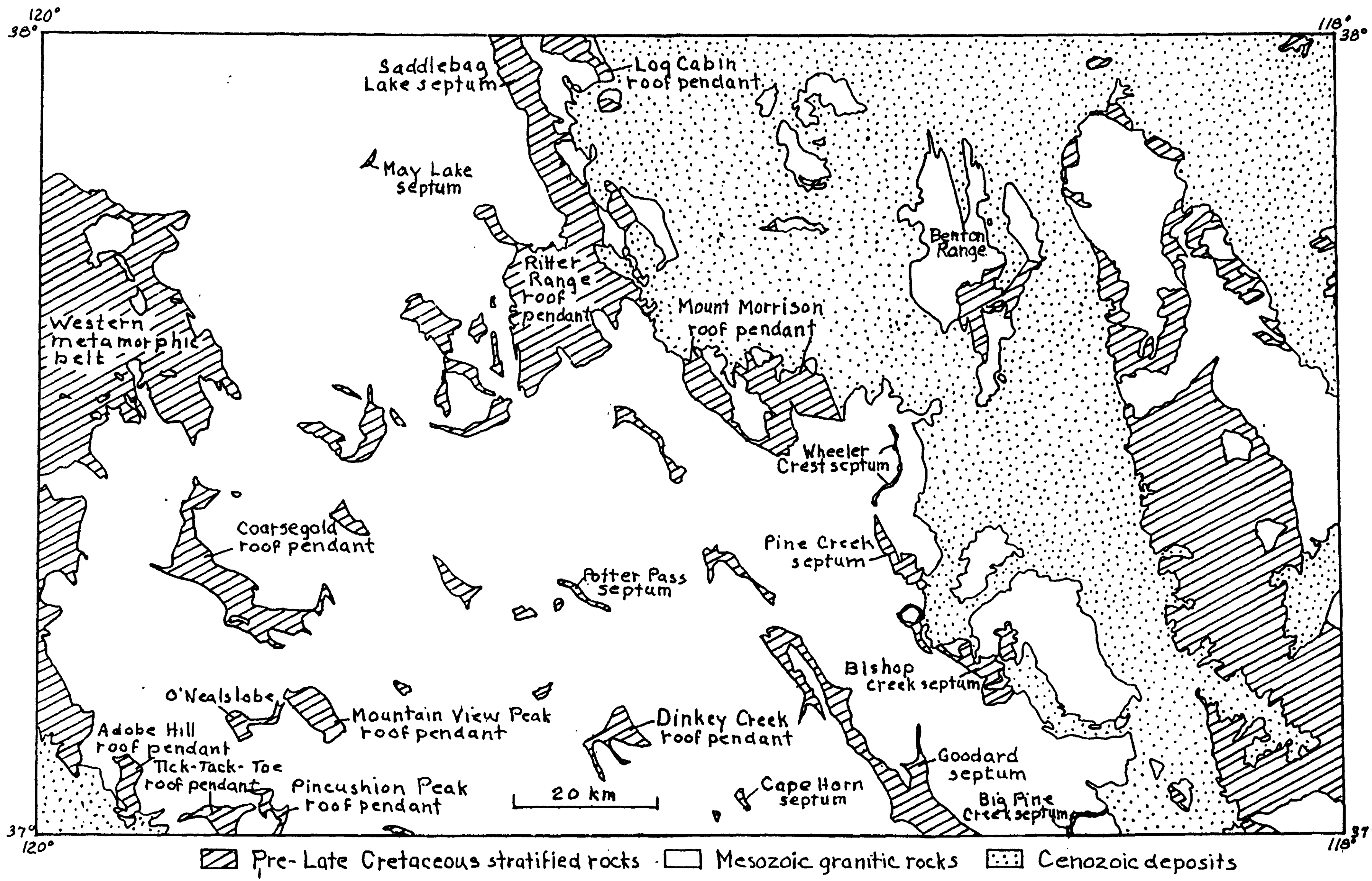

FIGURE 2.--Simplified geologic map showing distribution of country rocks within Mariposa 1° by 2° quadrangle.

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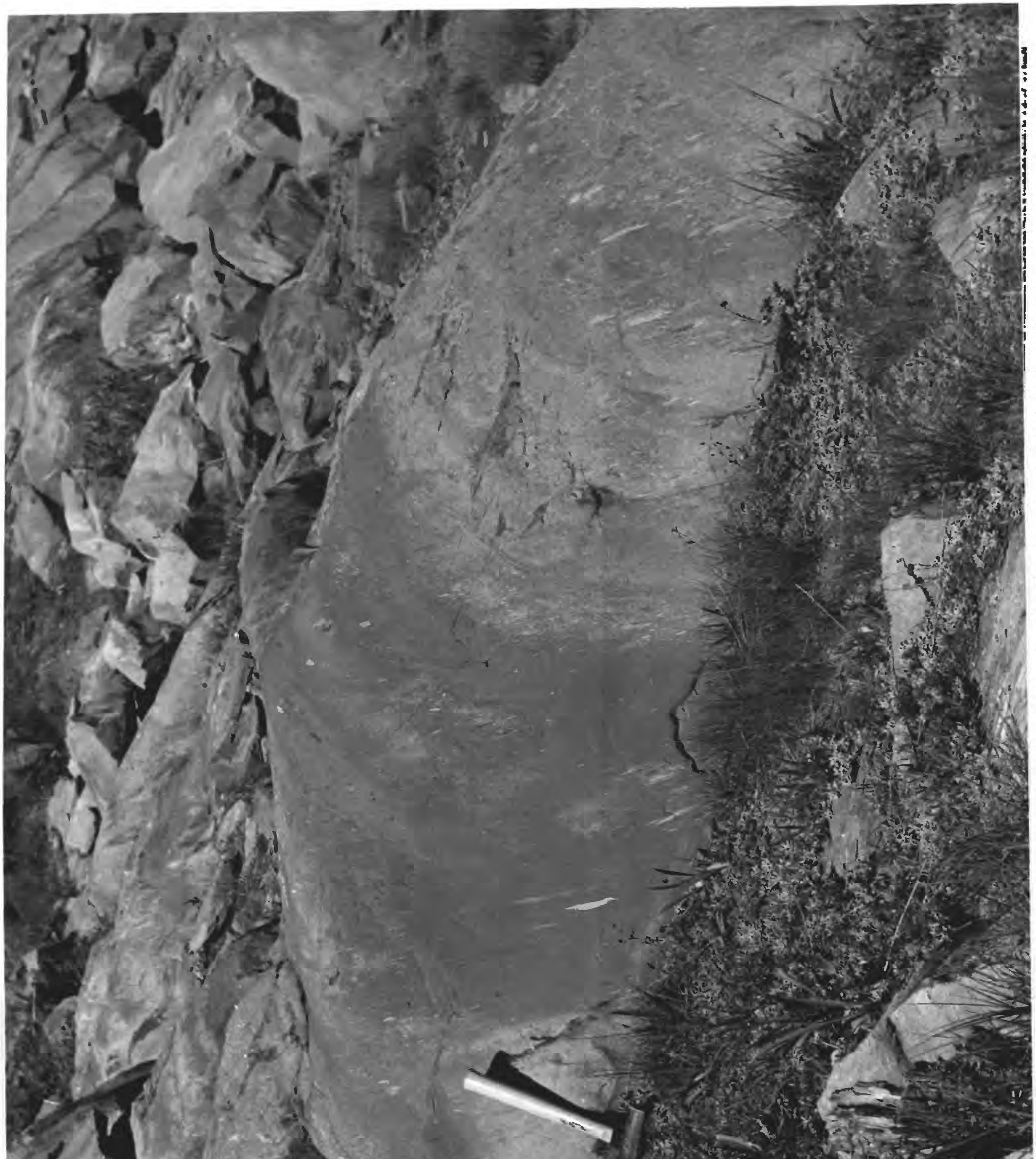


FIGURE 3.--Tuffaceous metavolcanic rocks in Goddard septum showing bedding, cross-cutting cleavage, and downdip lineation.

FIGURE 4.--Folded rhythmically layered Triassic chert along Merced River.

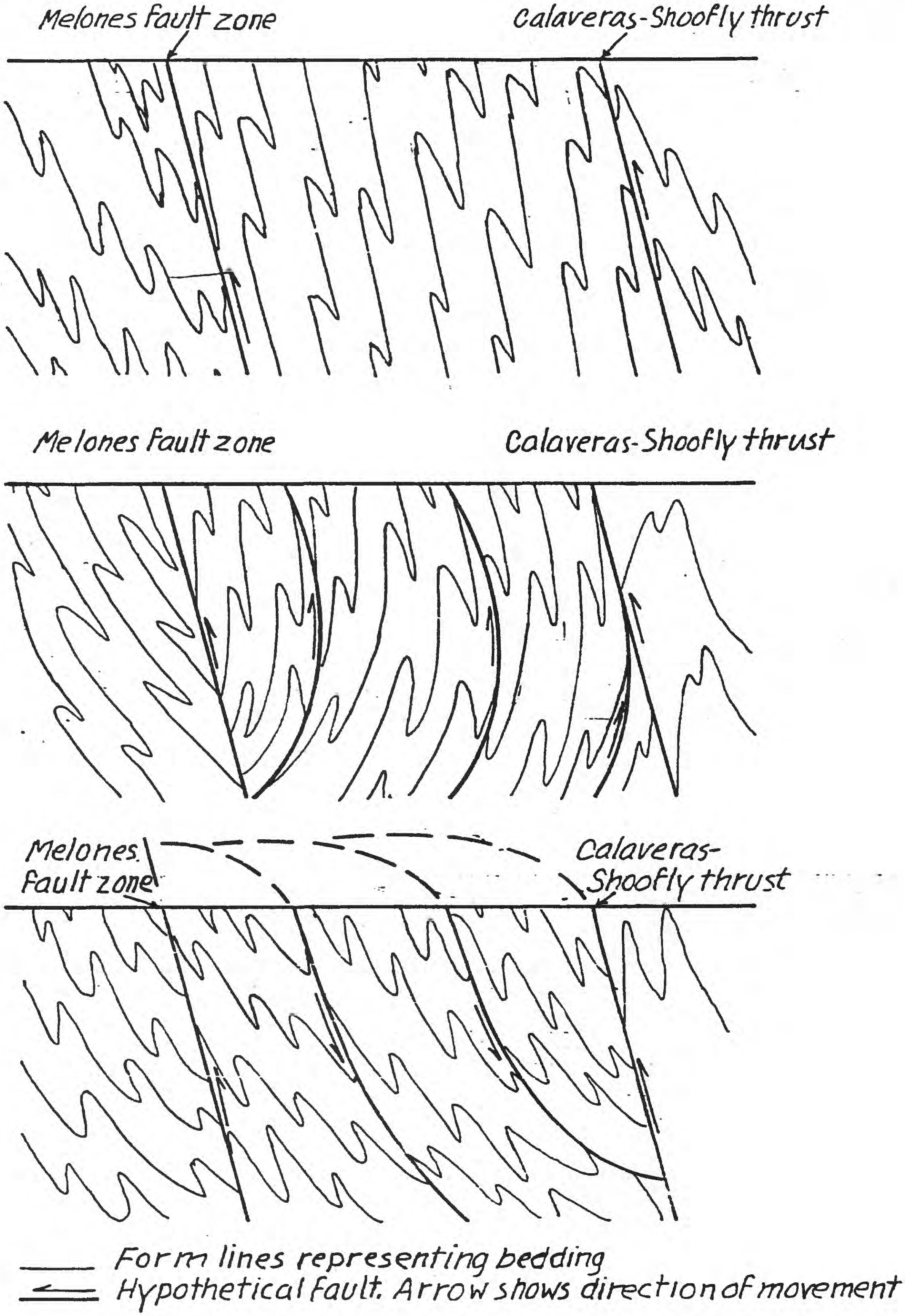


FIGURE 5.--Hypothetical cross sections showing three possible interpretations of structure in western metamorphic belt between Melones and Calaveras-Shoofly faults. For details, see discussion in text.

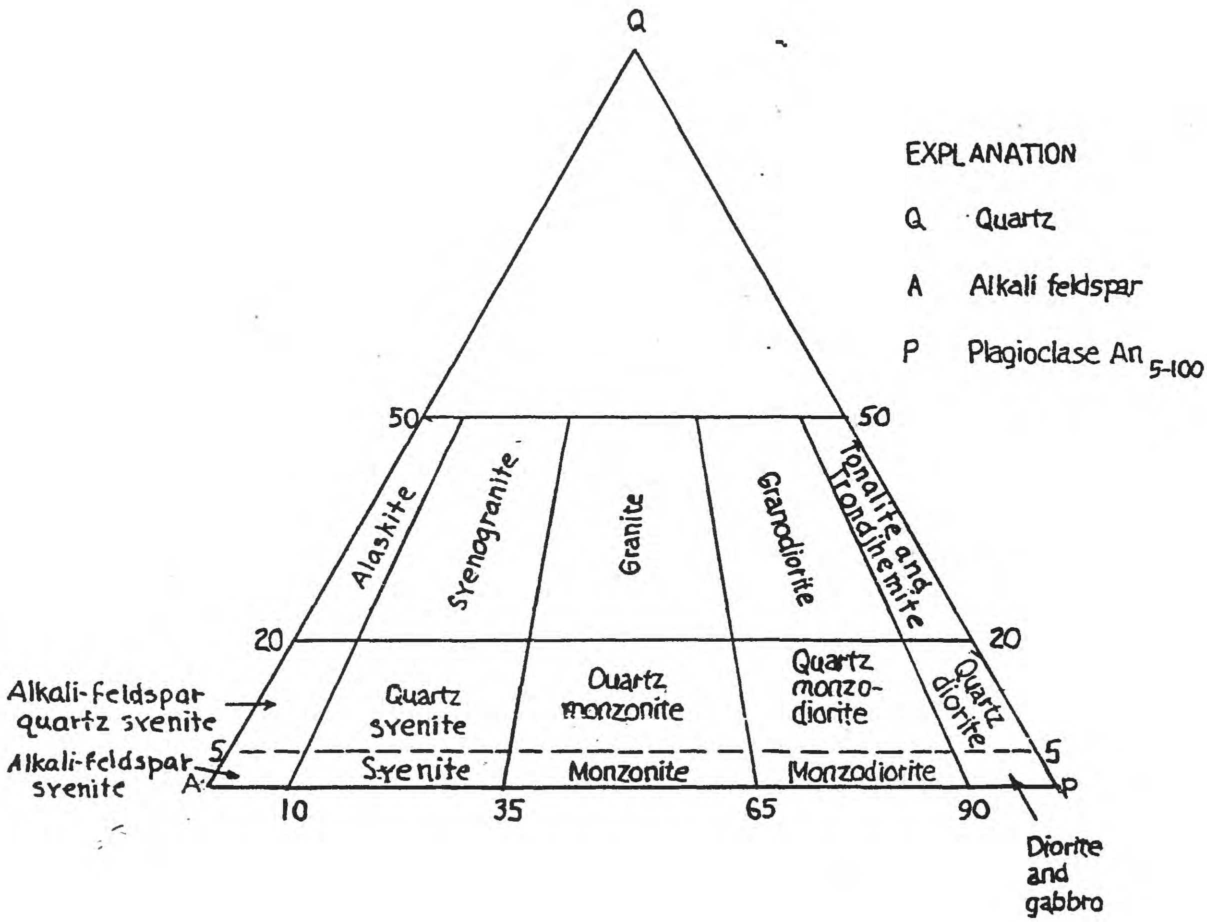


FIGURE 6.--Classification of Sierran plutonic rocks. Modified from Streckeisen (1973).

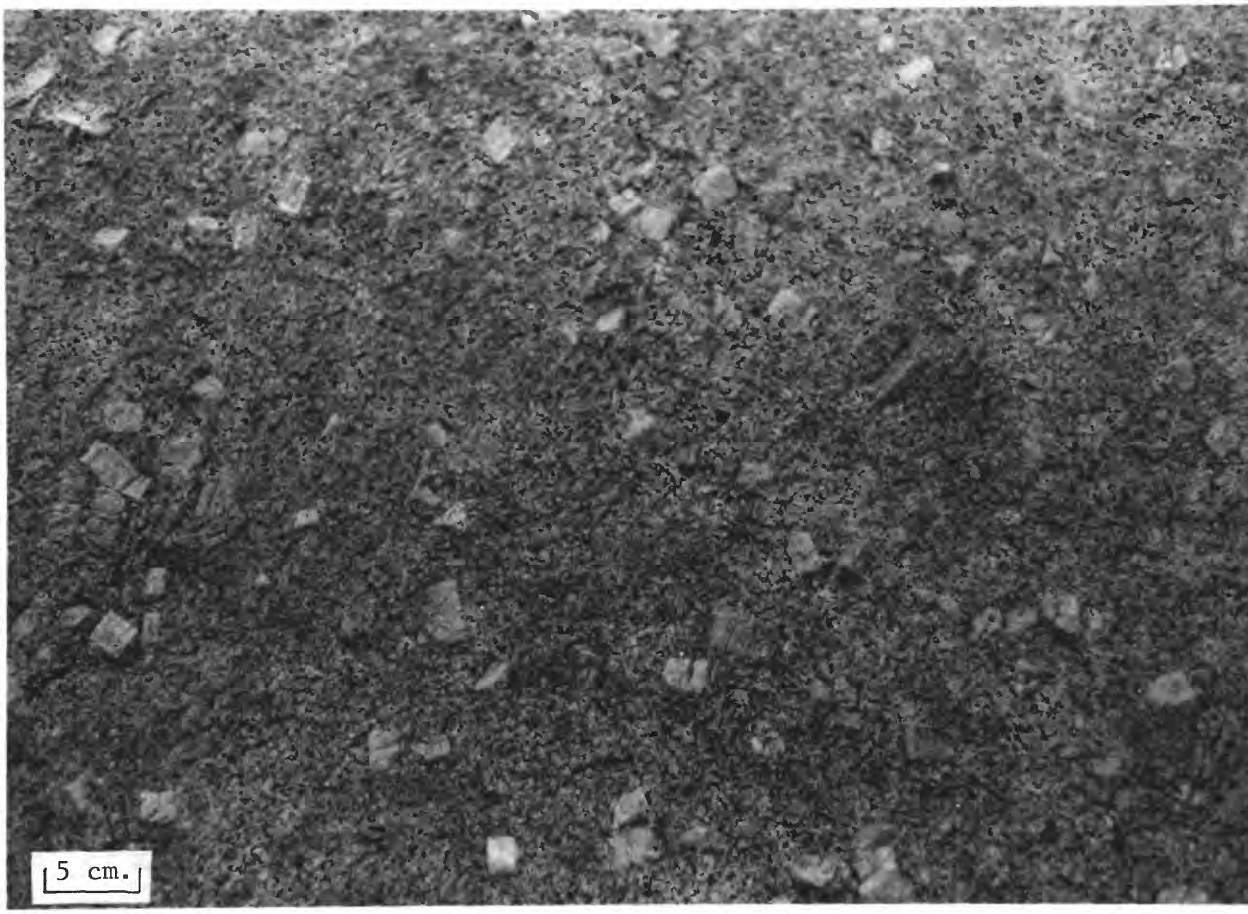


FIGURE 7.--Typical distribution of megacrysts in outer part of Cathedral Peak Granodiorite.

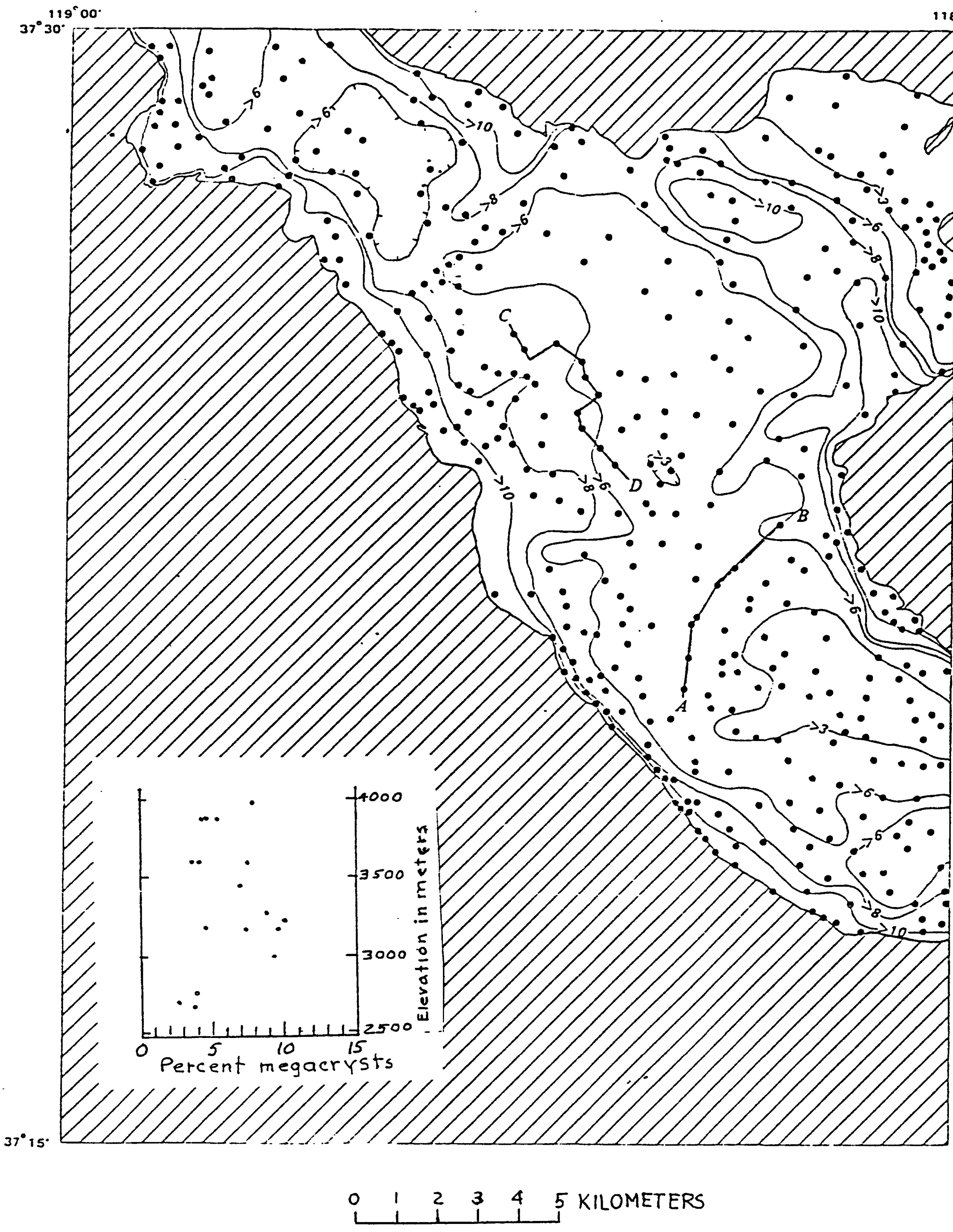


FIGURE 8.--Abundance of alkali-feldspar megacrysts in central and southern part of Mono Creek Granite. Dots show sample localities. Isopleths show percent megacrysts in rock. Insert shows abundance of alkali-feldspar megacrysts relative to elevation along traverses A-B and C-D. Modified from Lockwood (1975).

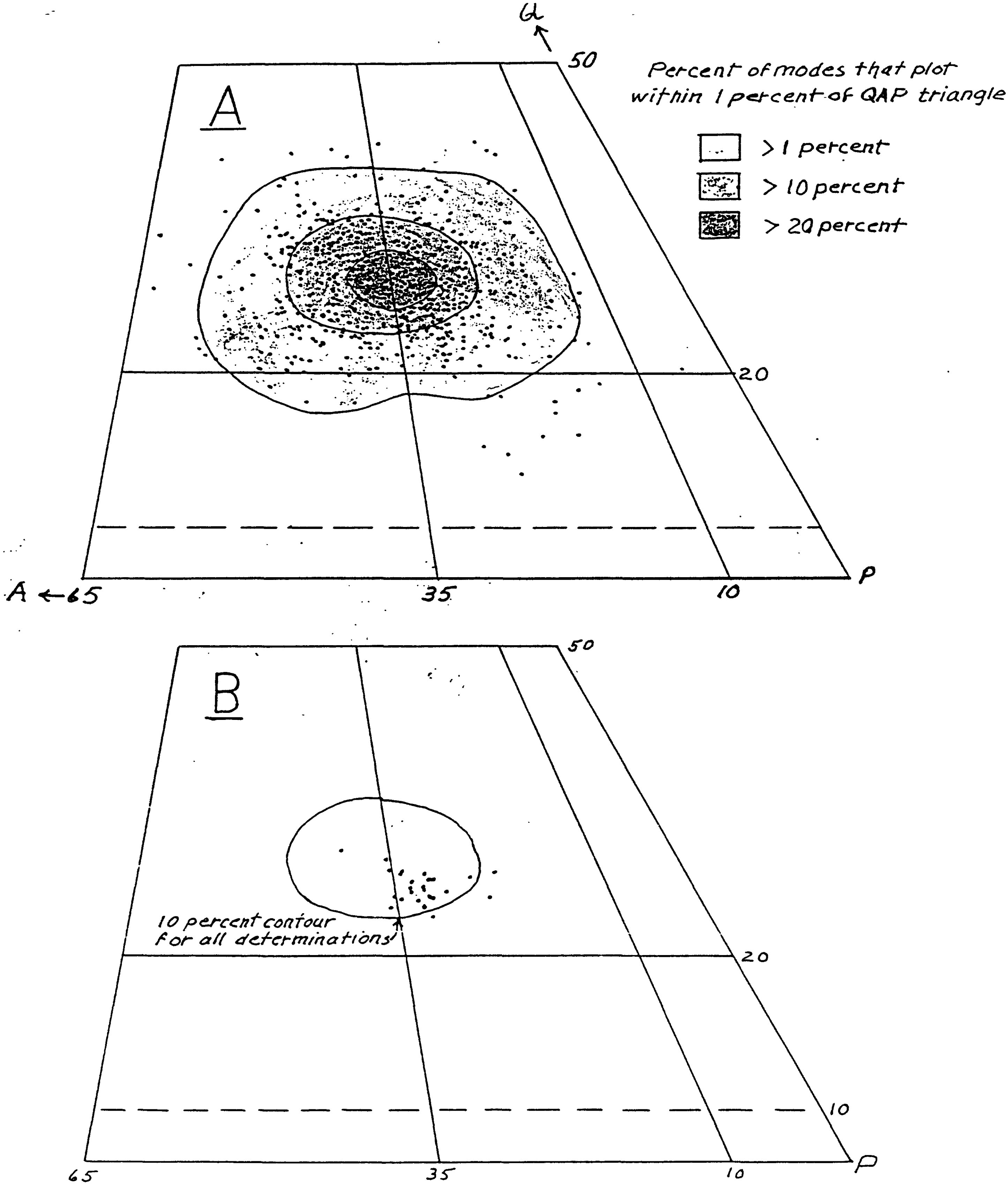


FIGURE 9.--Modes of plutonic rocks containing alkali-feldspar megacrysts. (A), All rocks within map area (Mariposa 1° by 2° quadrangle, plate 1), including Mono Creek and Cottonwood Creek Granites, Cathedral Peak and Wheeler Crest Granodiorites, granodiorite of McKinley Grove, megacrystic facies of Mount Givens, Half Dome, and Dinkey Creek Granodiorites, and megacrystic facies of granodiorite of Eagle Peak. (B), Superior modes determined in the field by Bateman and Chappel (1979) on glacially polished outcrops of Cathedral Peak Granodiorite.

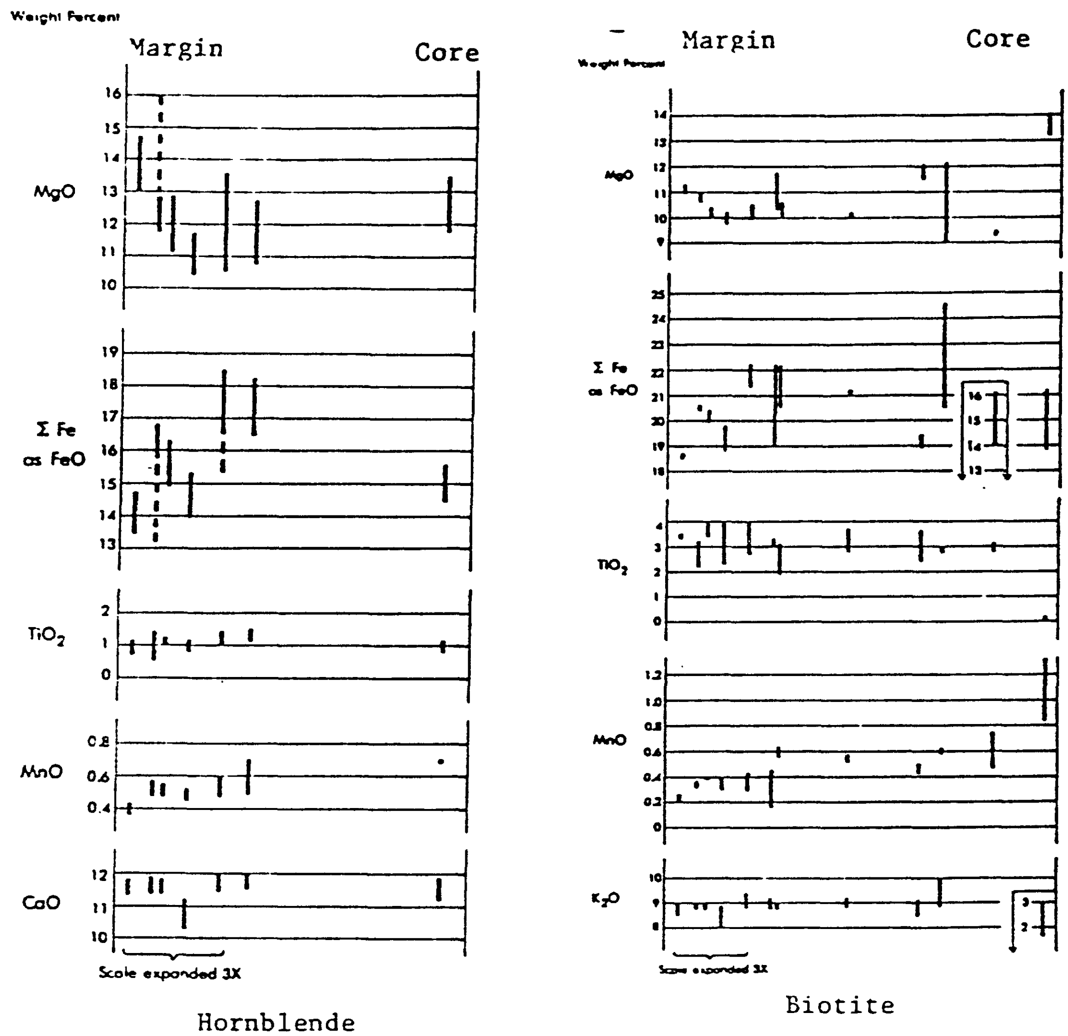


FIGURE 10.--Composition of hornblende and biotite in samples collected from margin to core of compositionally zoned bulbous head at north end of Mount Givens Granodiorite. Dashes indicate range in one or two determinations of several. (From Bateman and Nokleberg, 1978).

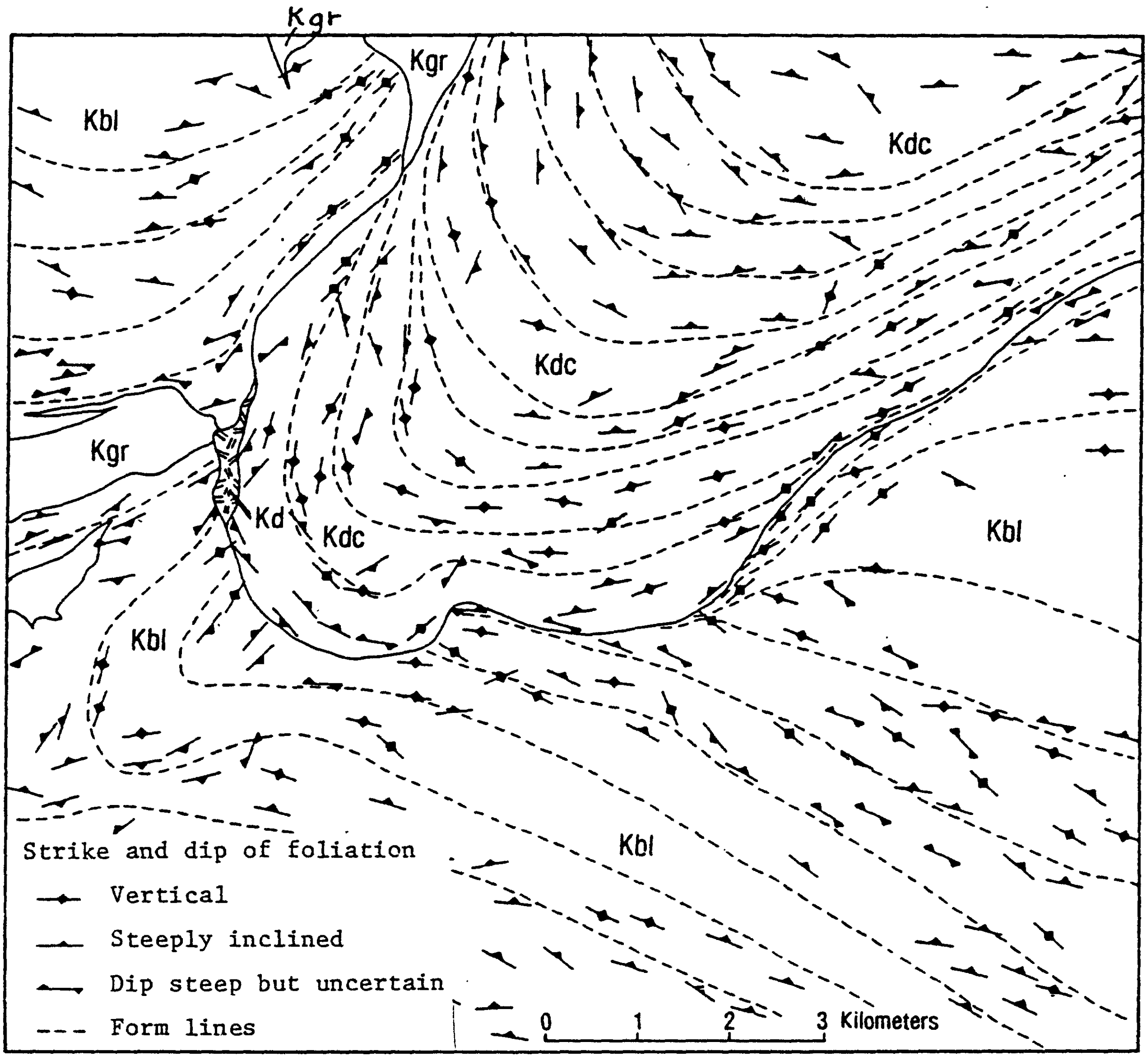


FIGURE 11.--Original and deformed foliation patterns south of Shaver Lake. Arcuate pattern of foliation in Dinkey Creek Granodiorite is original, whereas foliation pattern in older Bass Lake Tonalite (Kbl) was disturbed during emplacement of Dinkey Creek Granodiorite (Kdc). Kgr, younger granite; Kd, older diorite.

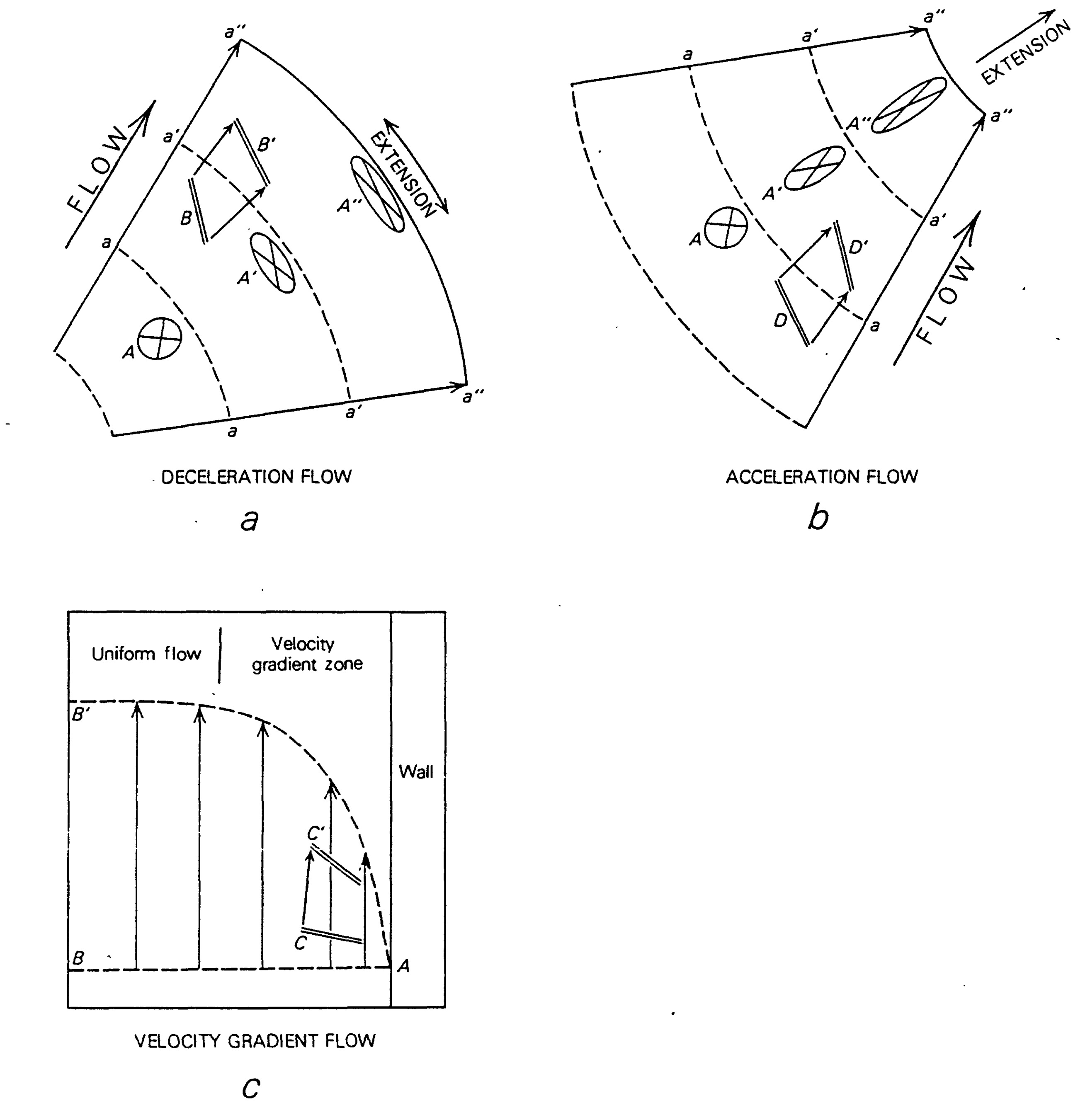


FIGURE 12.--Three types of nonuniform flow that can produce foliation and lineation in plutonic rocks (modified from Mackin, 1947). Diagrams represent either horizontal or vertical sections. See text for discussion.

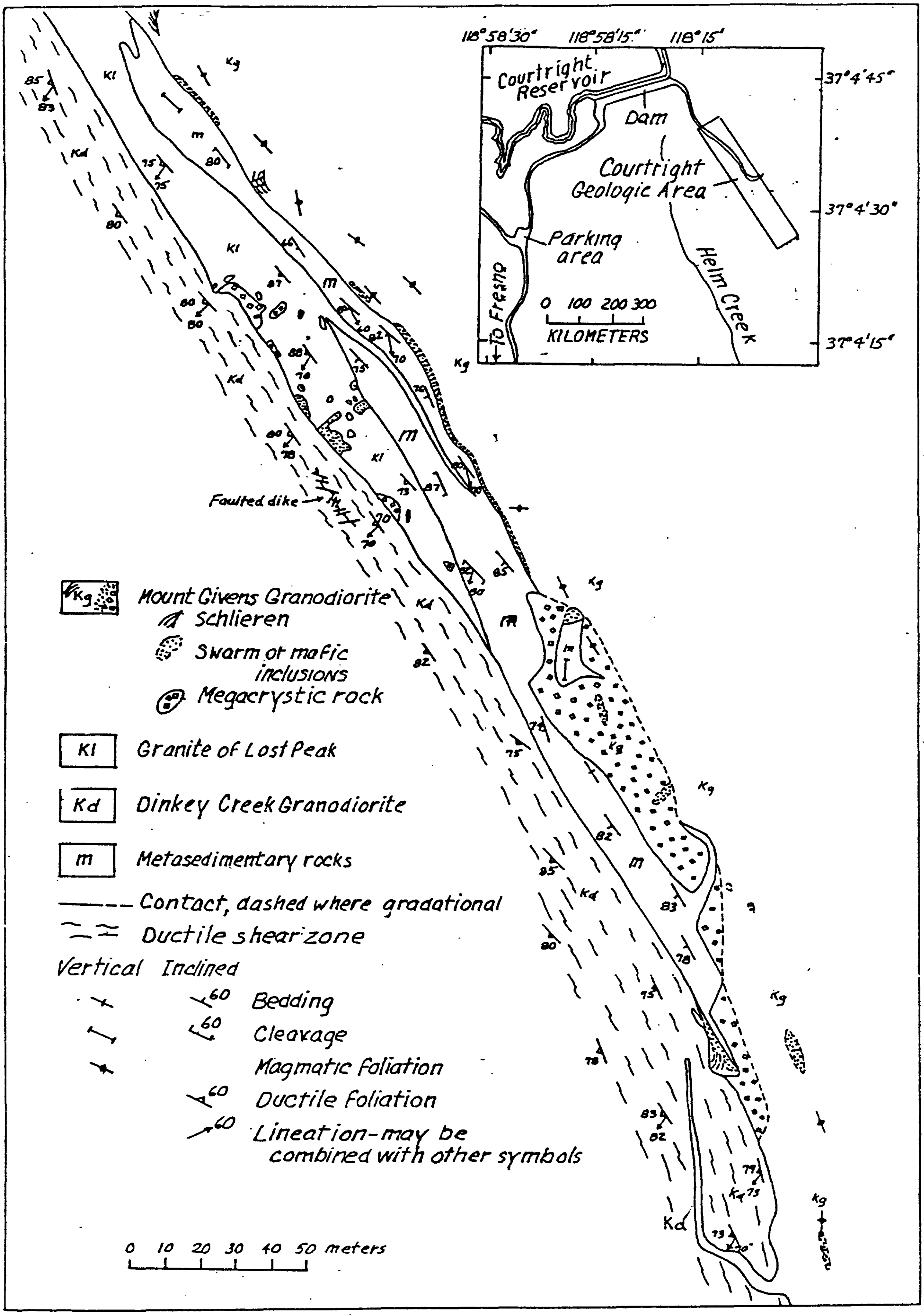


FIGURE 13.--Geologic map of Courtright Geologic Area at the south end of Courtright Reservoir.

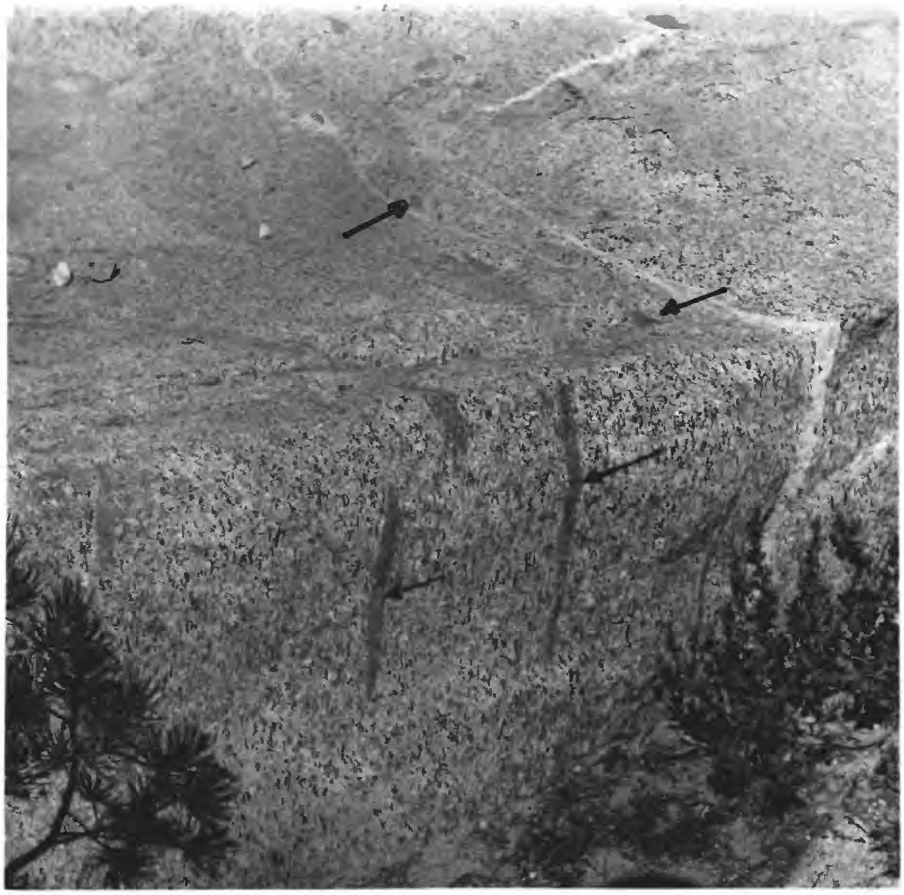


FIGURE 14.--Dinkey Creek Granodiorite showing mafic inclusions (arrows) drawn into bladelike forms as a result of ductile deformation. Courtright Geologic Area.

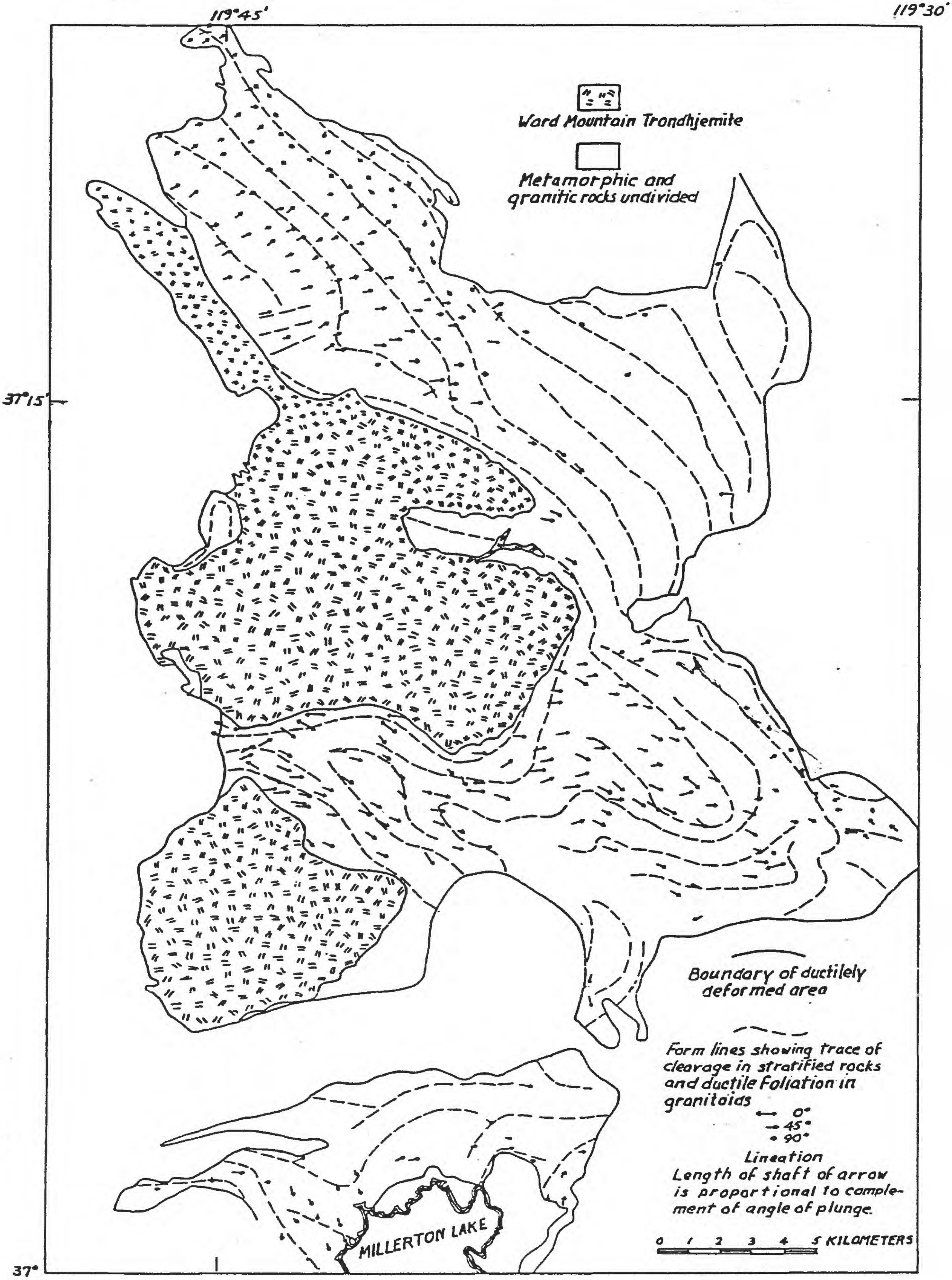


FIGURE 15.--Patterns of ductile foliation and lineation in Bass Lake Tonalite and metamorphic rocks adjacent to Ward Mountain Trondhjemite. (From Bateman and Nokleberg, 1978).

Figure]6.--Oriented double-convex mafic inclusions in Dinkey Creek Granodiorite. Southwest shore of Shaver Lake east of Shaver Lakes Heights.

FIGURE 17--Mafic dike in left side of photograph truncates foliation in Base lake tonalite shown by oriented double-convex mafic inclusions. Roadcut along road to North Fork about 9 km southwest of North Fork. Strontium isotopic ratios on these rocks are plotted in figure 18.

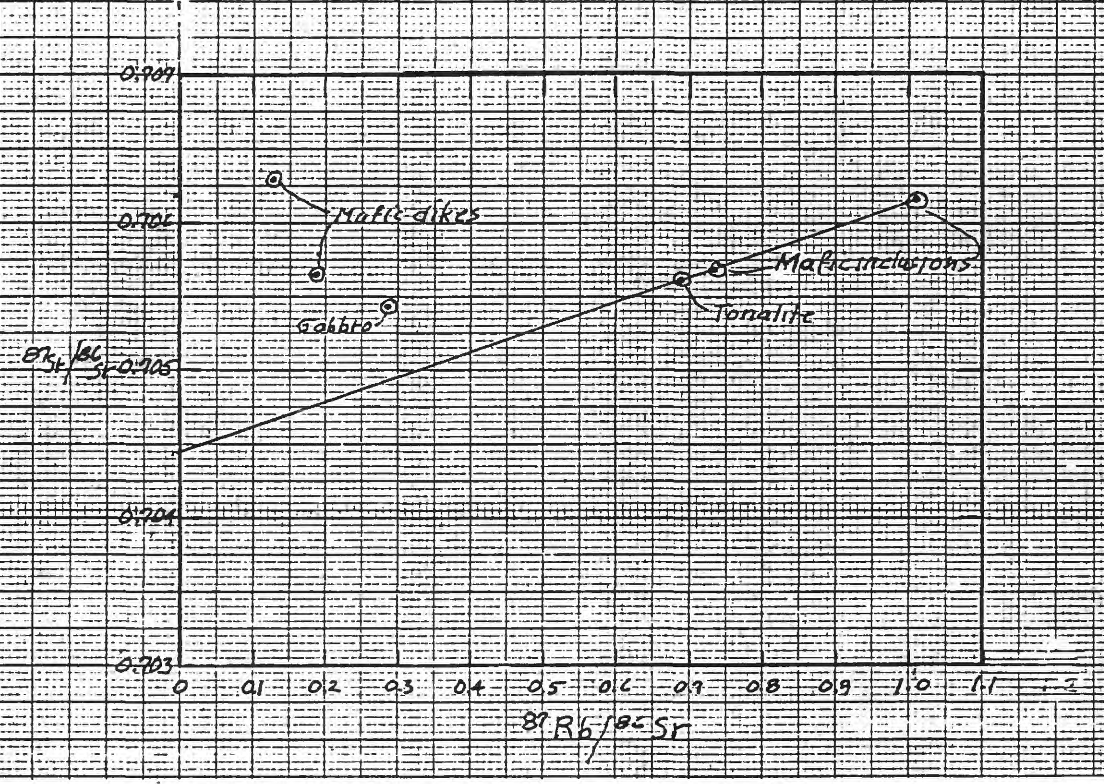


FIGURE 18.--Plot of strontium isotopic ratios of Bass Lake Tonalite, mafic inclusions, and mafic dike shown in figure 17, and nearby body of gabbro. Isochron representing an age of 114 Ma (from U-Pb data) is drawn through plots of mafic inclusions and tonalite.

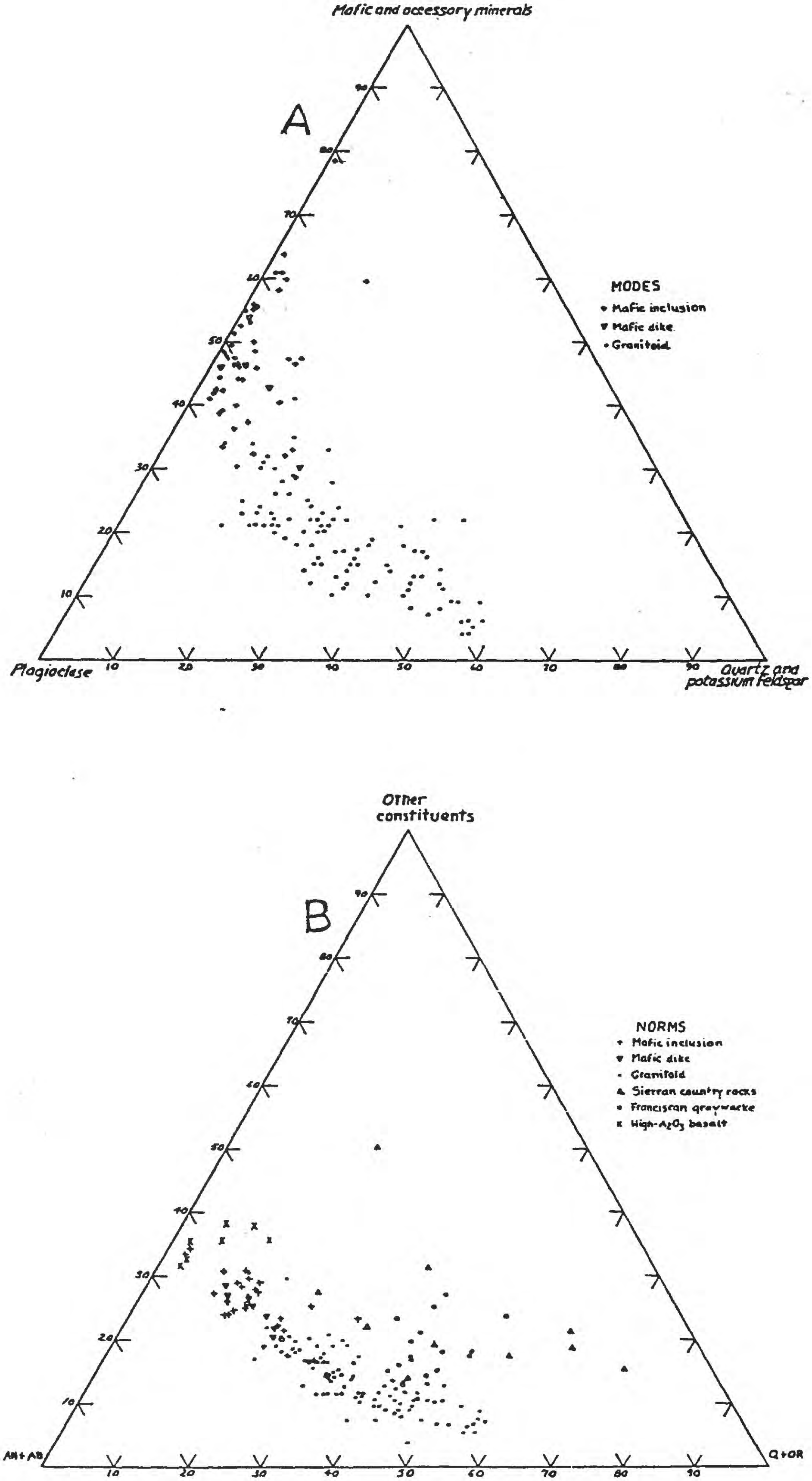


FIGURE 19.--Plots of chemically analyzed samples of granitoids (Bass Lake Tonalite, Dinkey Creek Granodiorite, and Mount Givens Granodiorite) and of mafic inclusions and mafic dikes within those rocks. A, modal plot. B, normative plot, which includes norms of country rocks within and east of Sierra Nevada (table 3), Franciscan graywackes (Bailey and others, 1964) and typical high-Al_2O_3 basalt (Carmichael and others, 1974).

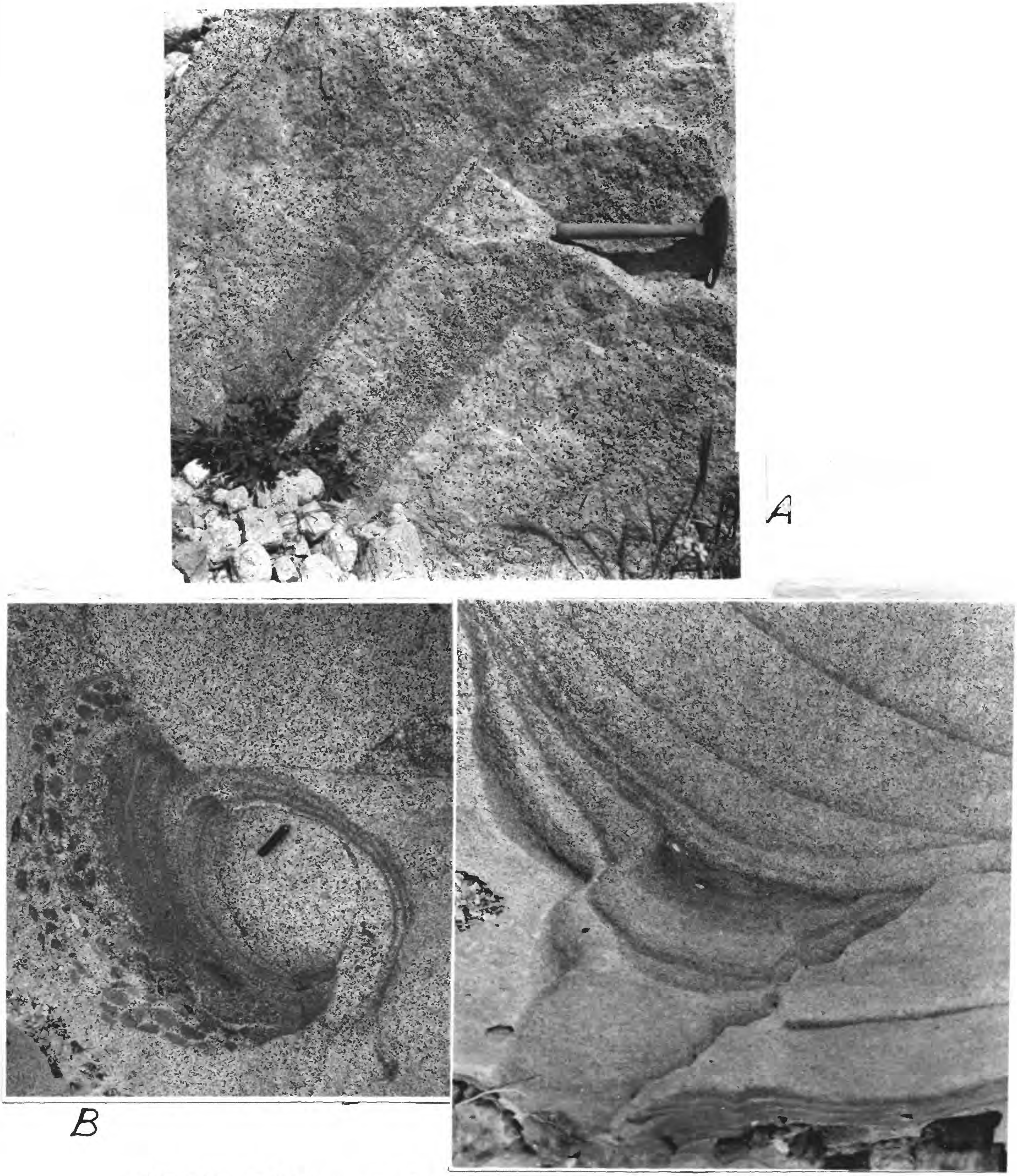


FIGURE 20.--Schlieren. A, planar schlieren in Half Dome Granodiorite at Tenaya Lake. B, tubular (faerie ring) and C, trough schlieren in Mount Givens Granodiorite at Courtright Geologic Area near Courtright Reservoir.

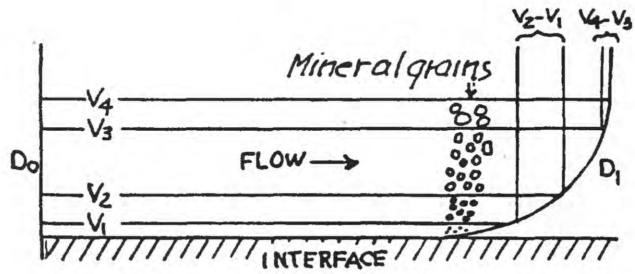


FIGURE 21. Flow sorting of mineral grains in magma. Flow is assumed to be from line D_0 toward the curve D_1 and parallel to an interface that corresponds to the boundary surface between two layers--one almost solid and the other fluid but laden with crystals. The distance between the D_0 and the curve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

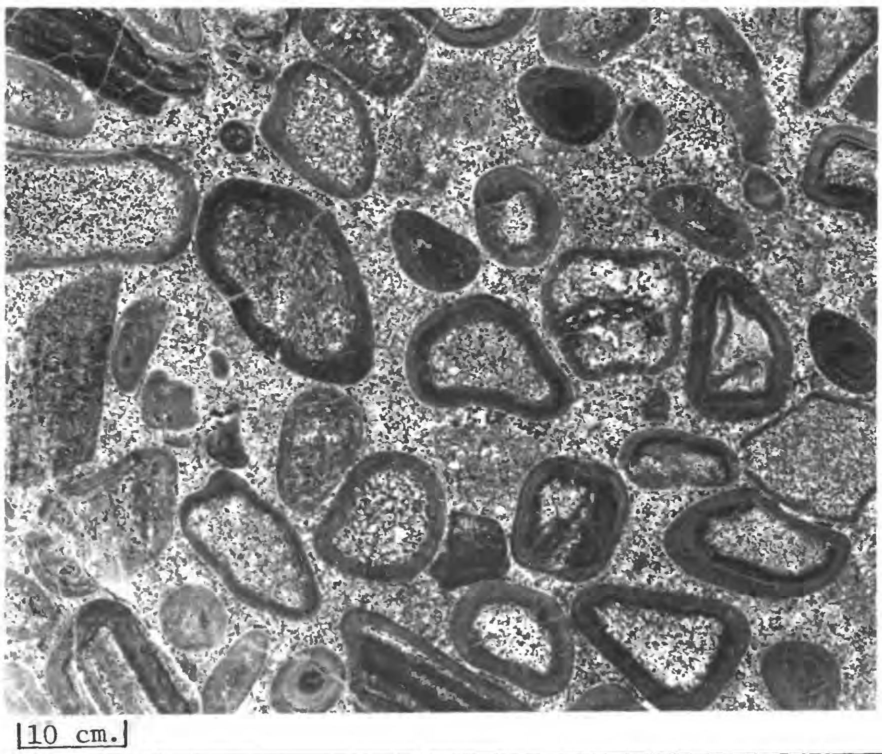


FIGURE 23.--Orbicules in mafic facies of Dinkey Creek Granodiorite.

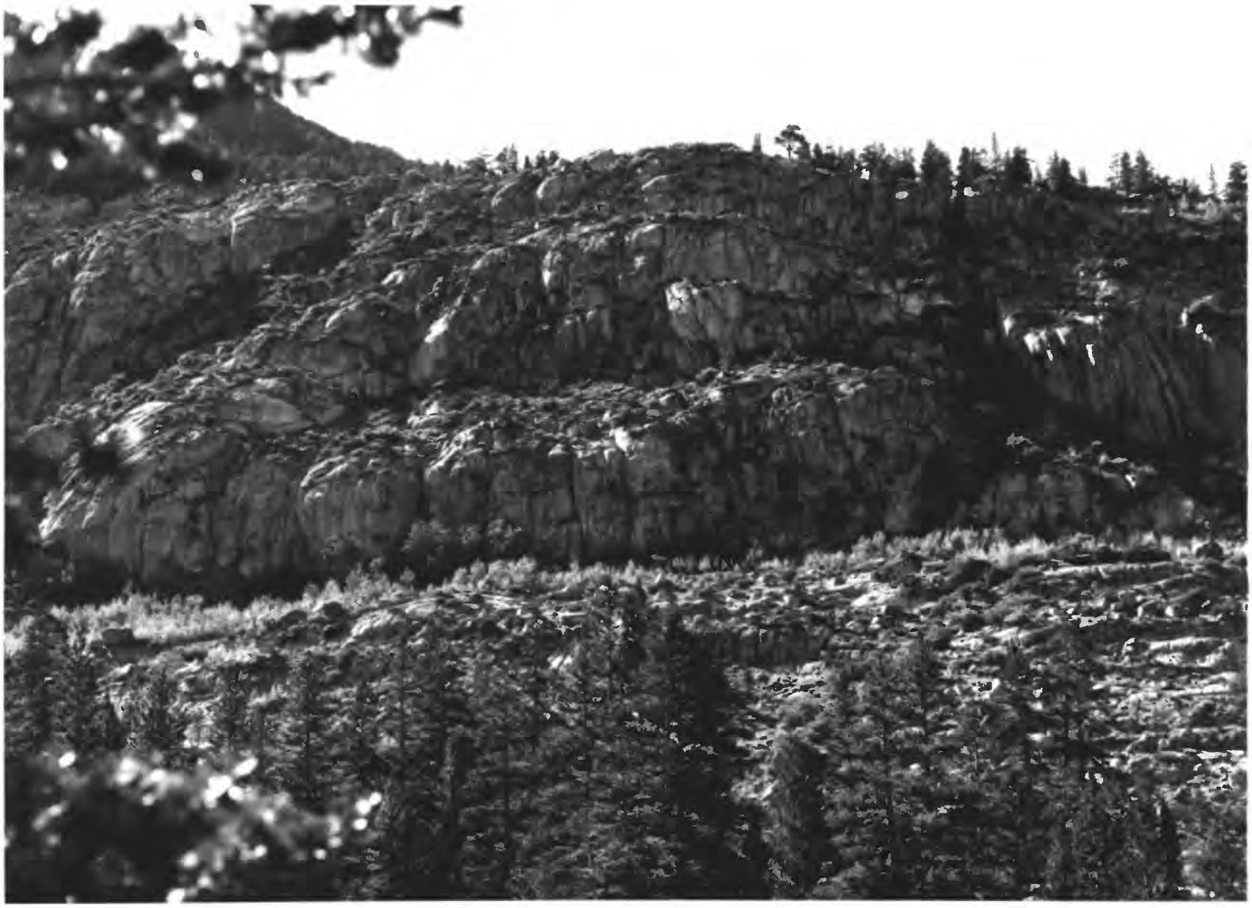


FIGURE 24.--Joint sets in Lamarck Granodiorite.

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

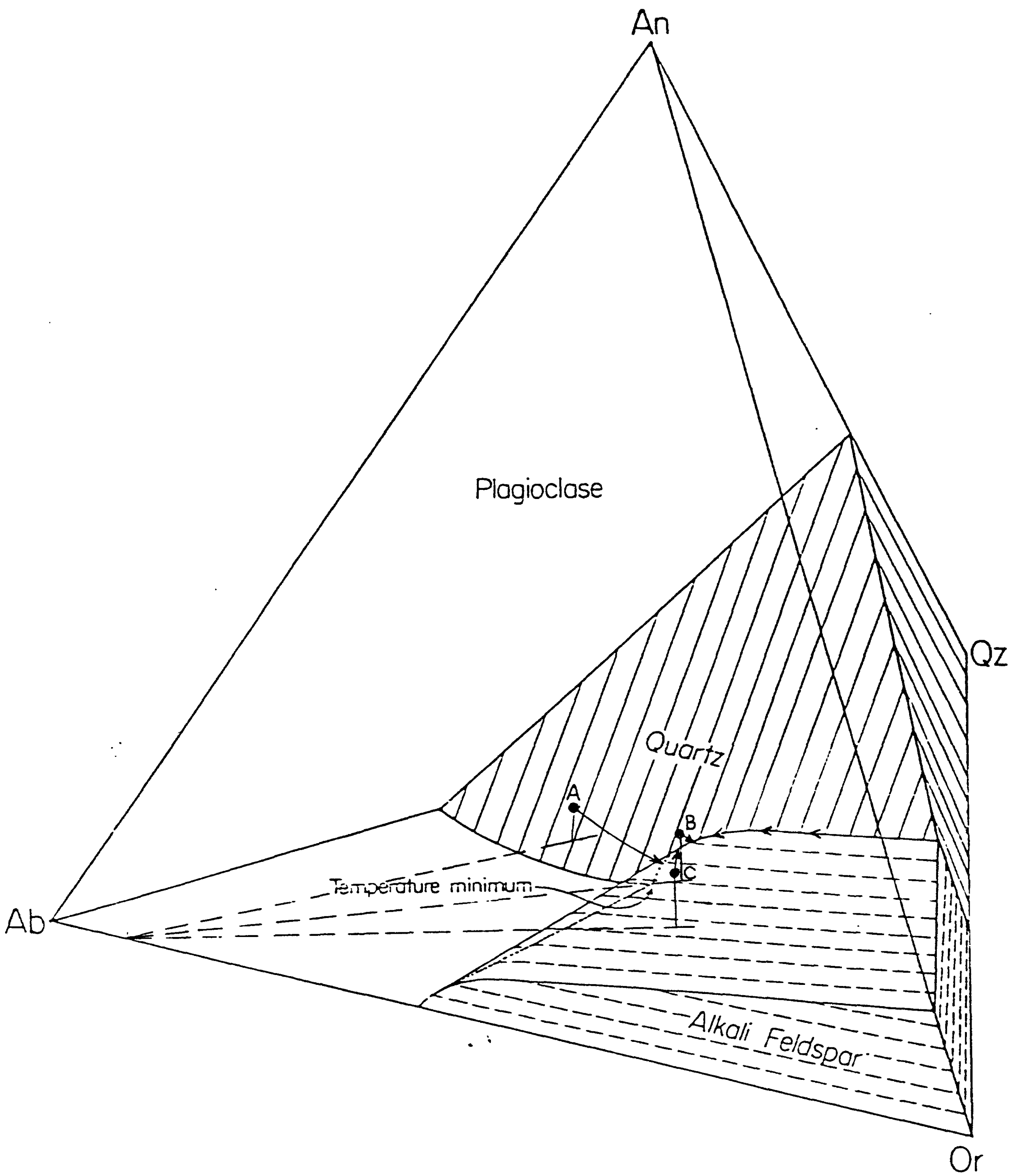


FIGURE 26.--Salic tetrahedron showing liquidus-phase volumes. A, B, and C are approximate positions where, with falling temperature and crystallization of plagioclase, granitic melts with low, medium, and high potassium contents would intersect surfaces bounding plagioclase liquidus phase-volume. Dashed lines are projected traces of these melts on the base of the tetrahedron. Arrows show subsequent paths of melt on the bounding surfaces.

Fine Gold Intrusive Suite (27A

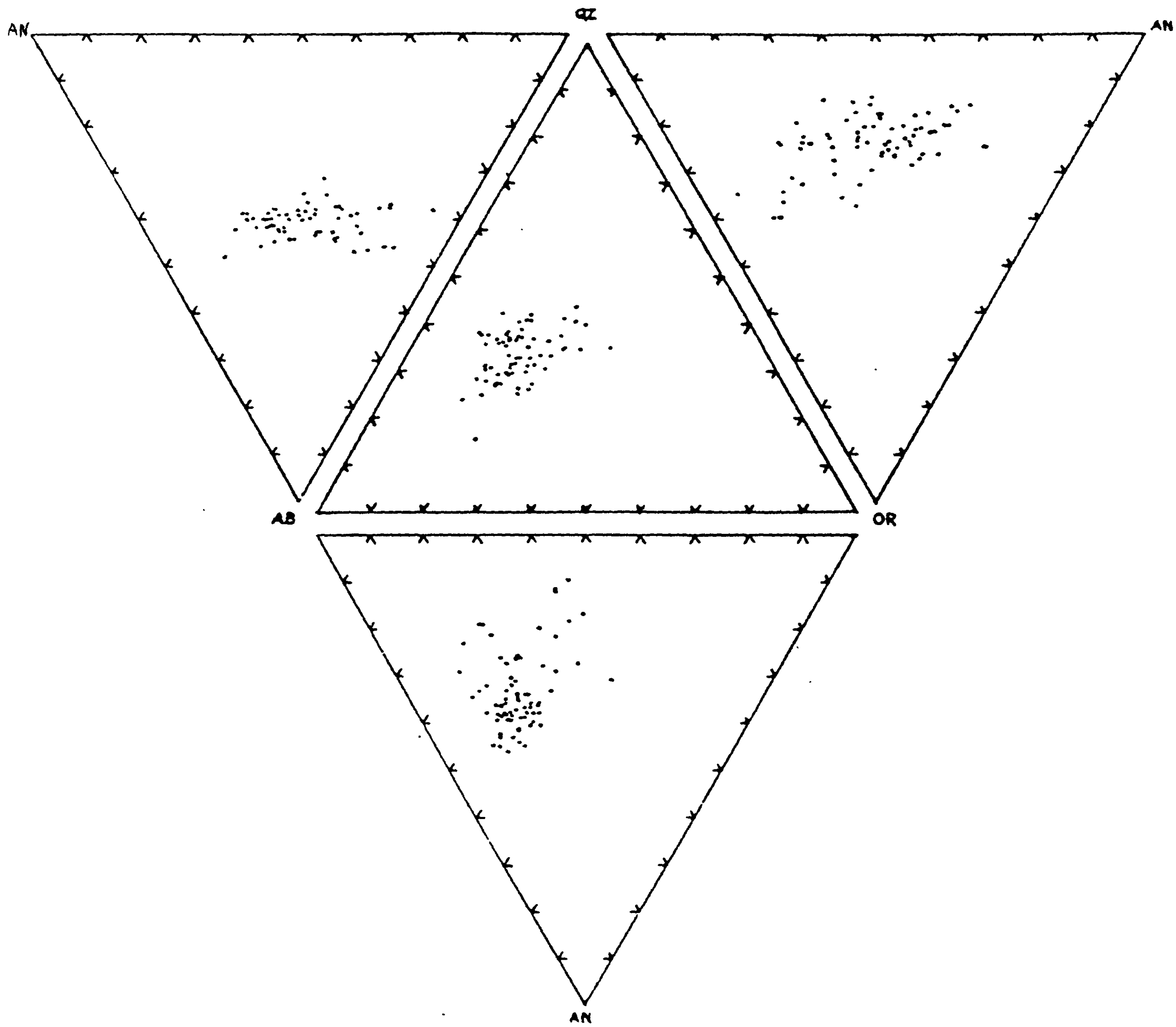
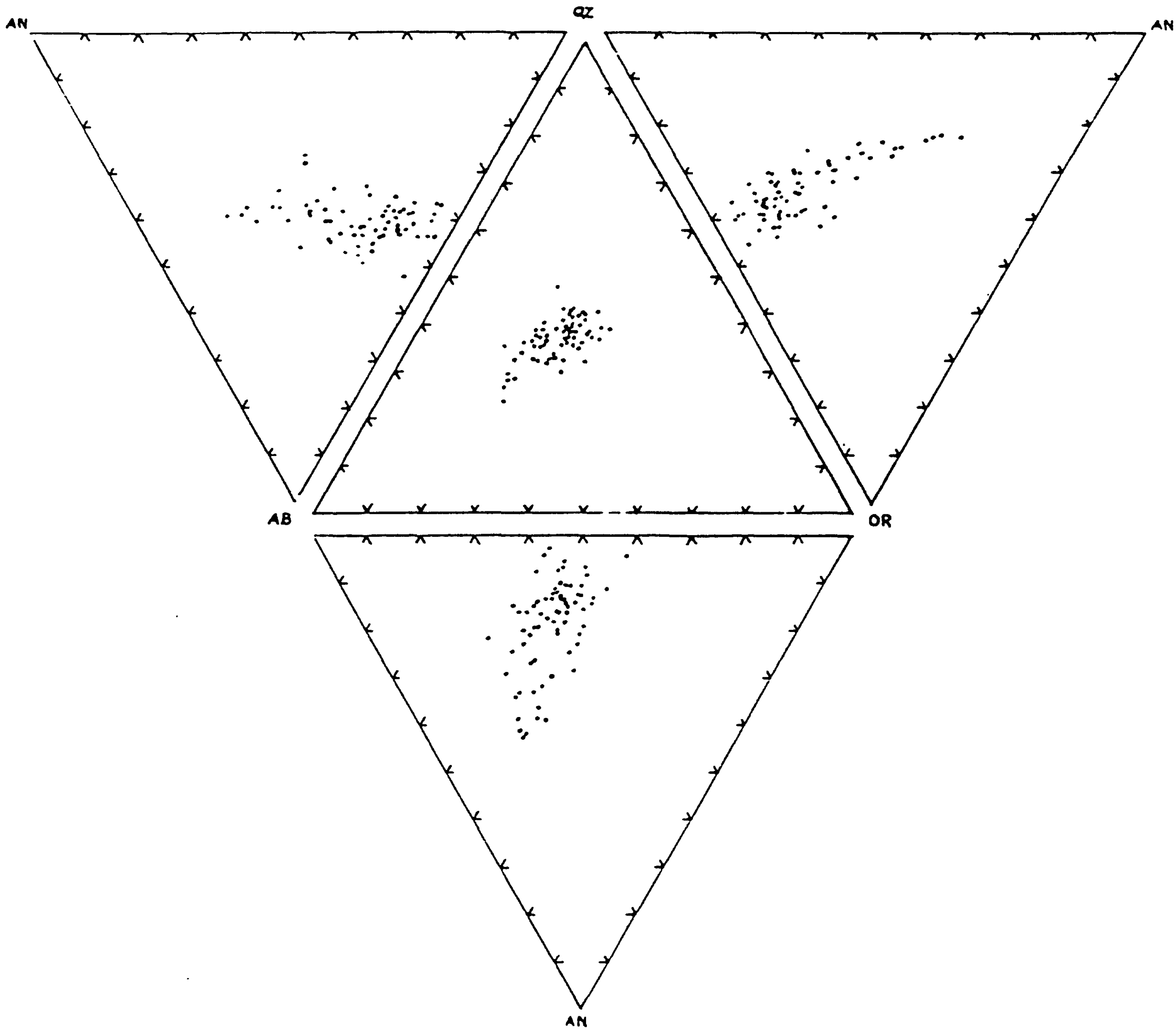
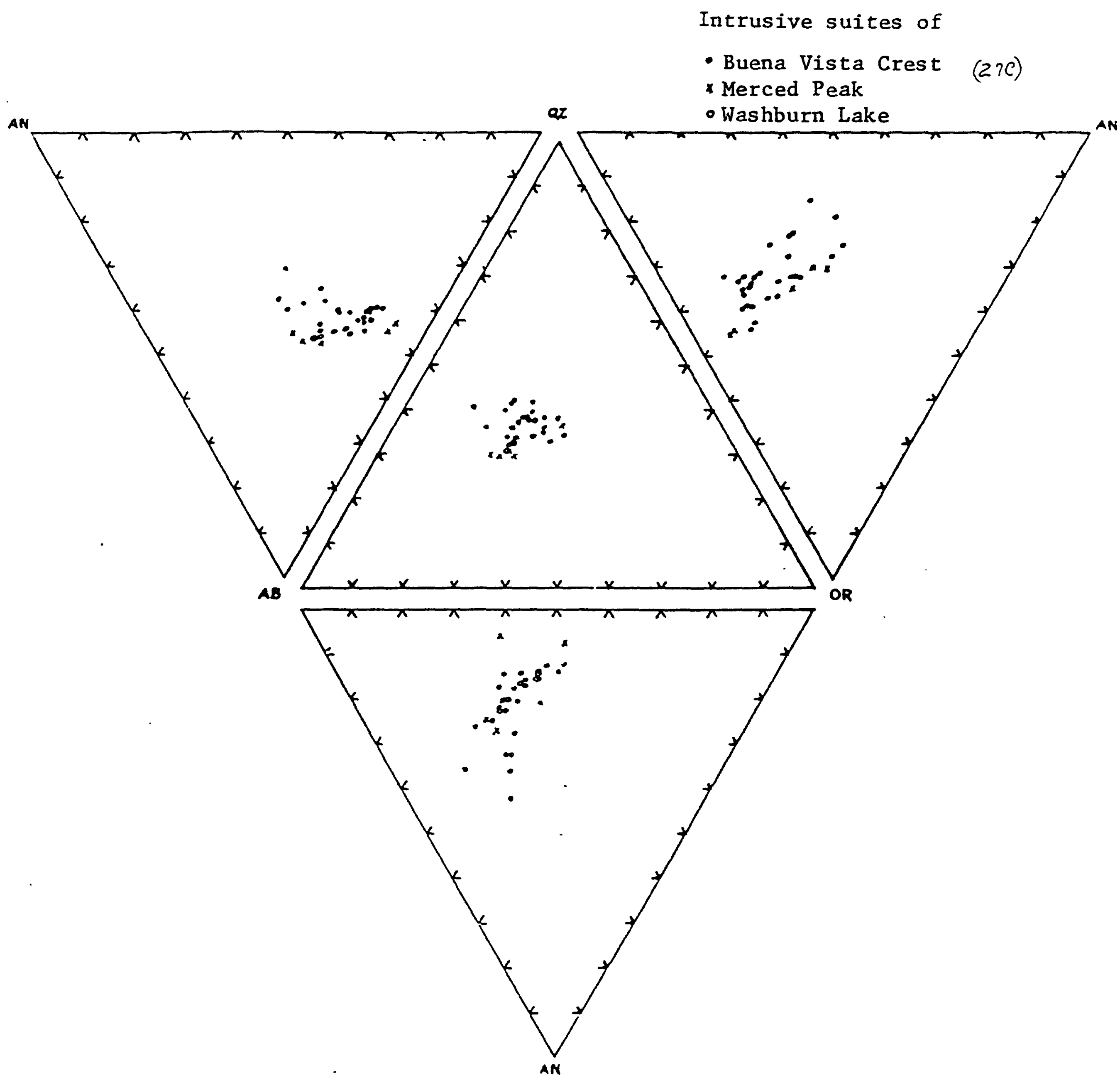
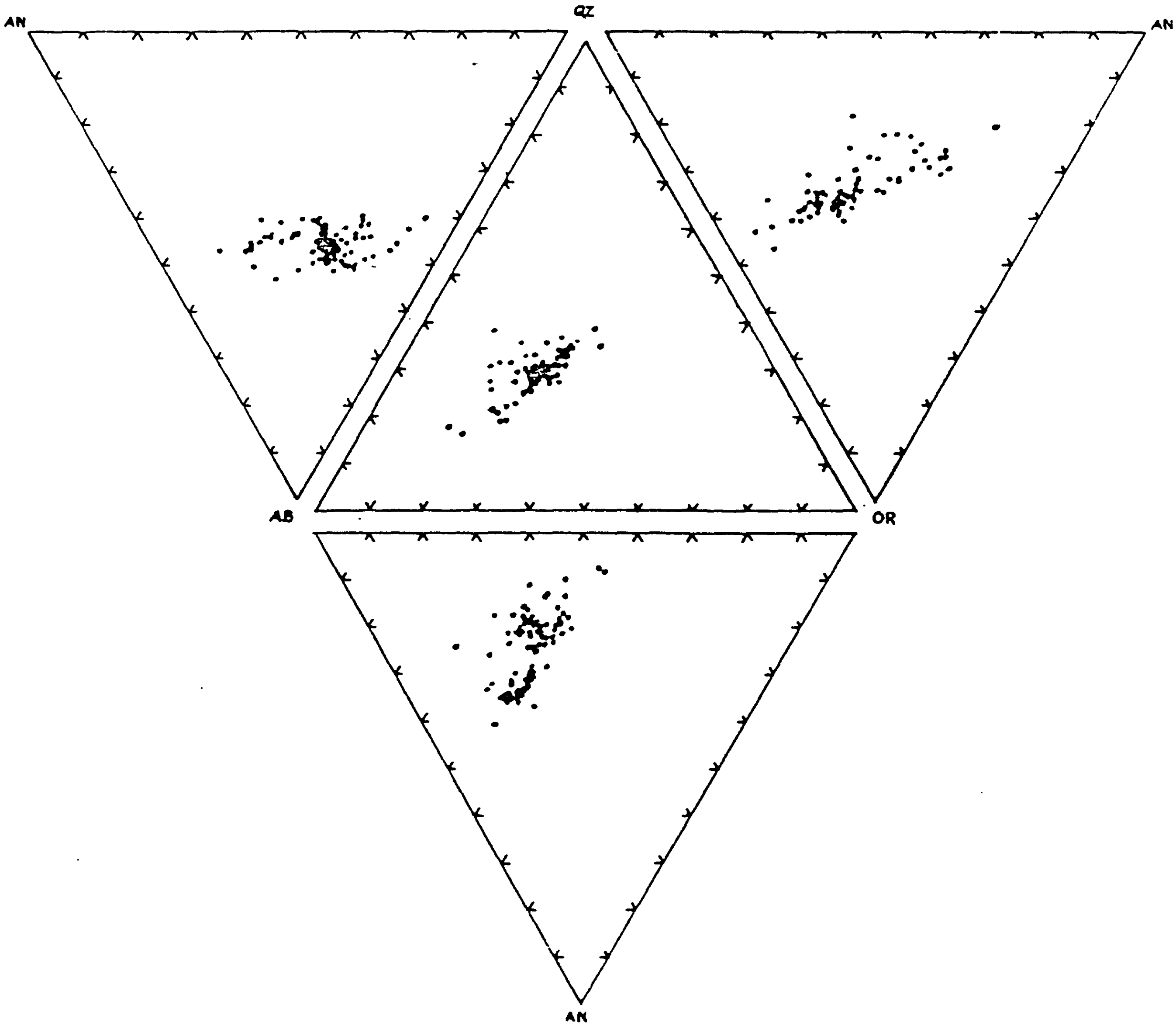


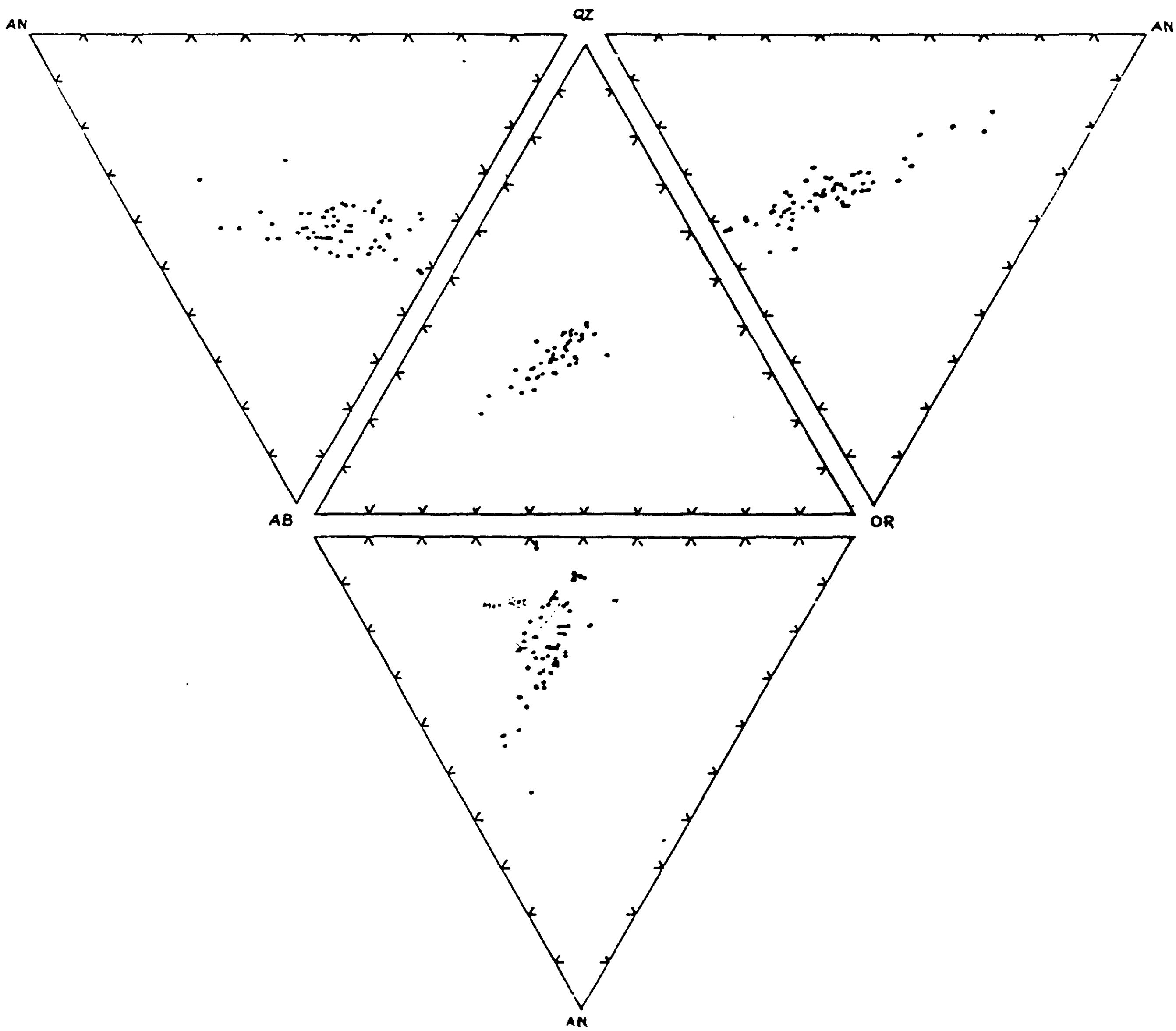
FIGURE 27.--Projections of CIPW norms of intrusive suites on faces of salic tetrahedron.

Shaver Intrusive Suite-
intrusive suite of Yosemite Valley (27B)




Intrusive suites of
• Buena Vista Crest
x Merced Peak
o Washburn Lake
(27C)
QZ
AN
AN
AB
OR
AN


QZ
AN
AN
AB
OR
AN


AN
QZ
AN
AB
OR
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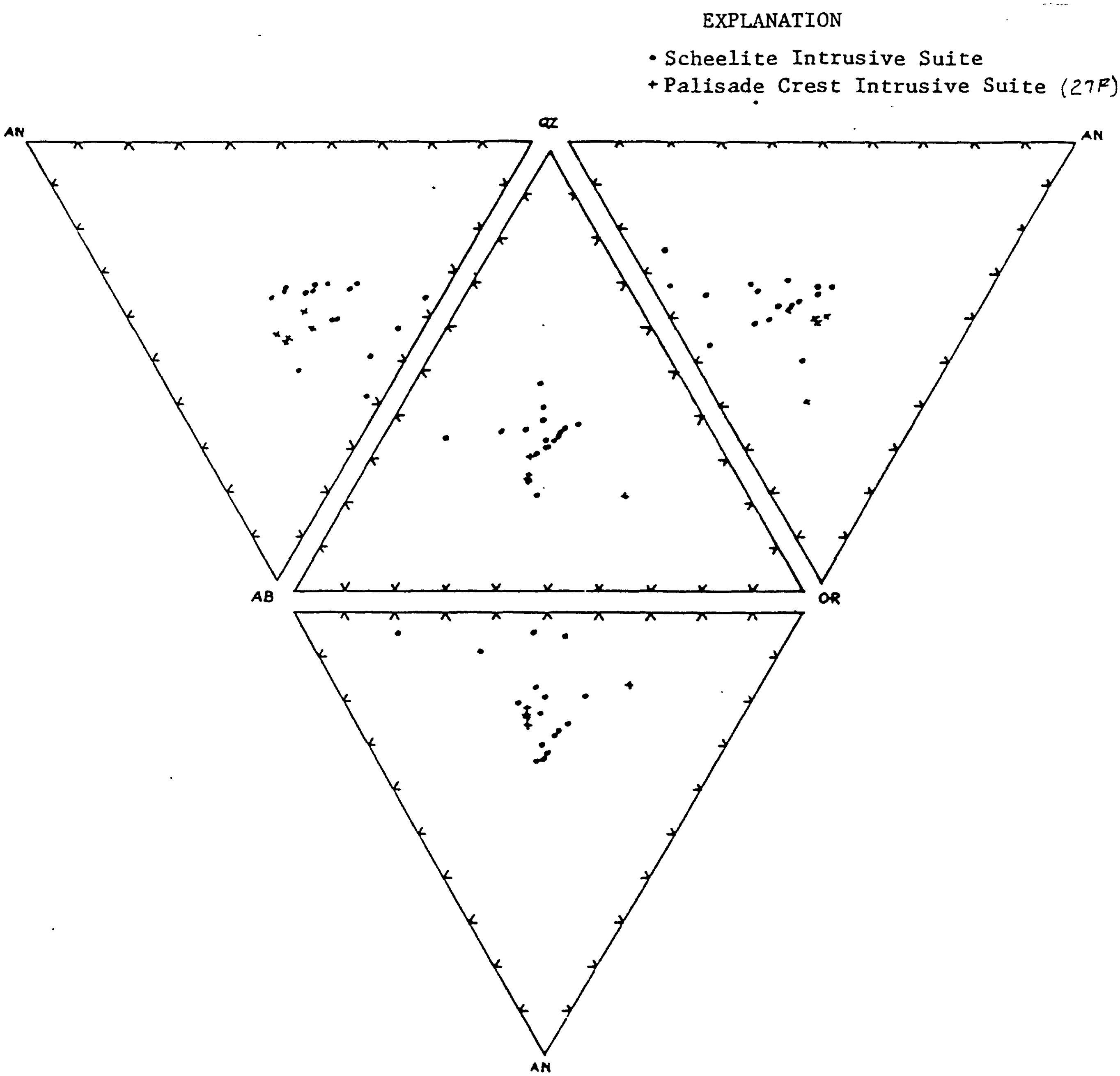
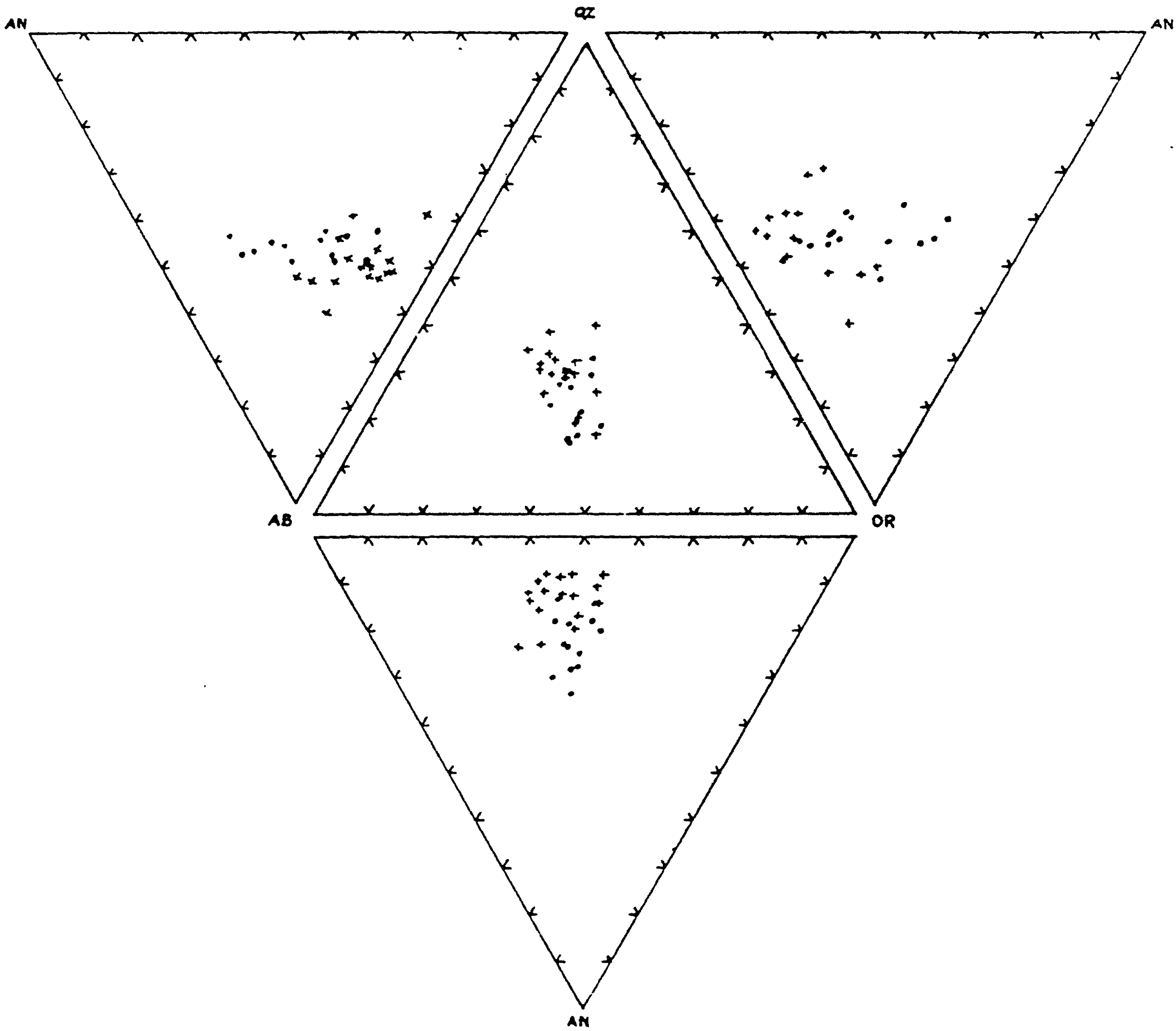

EXPLANATION
• Scheelite Intrusive Suite
+ Palisade Crest Intrusive Suite (27F)
AN
QZ
AN
AB
OR
AN

EXPLANATION (27G)

+ Cretaceous granites of the White Mountains
• Soldier Pass Intrusive Suite and unassigned
  Jurassic intrusions of the White Mountain



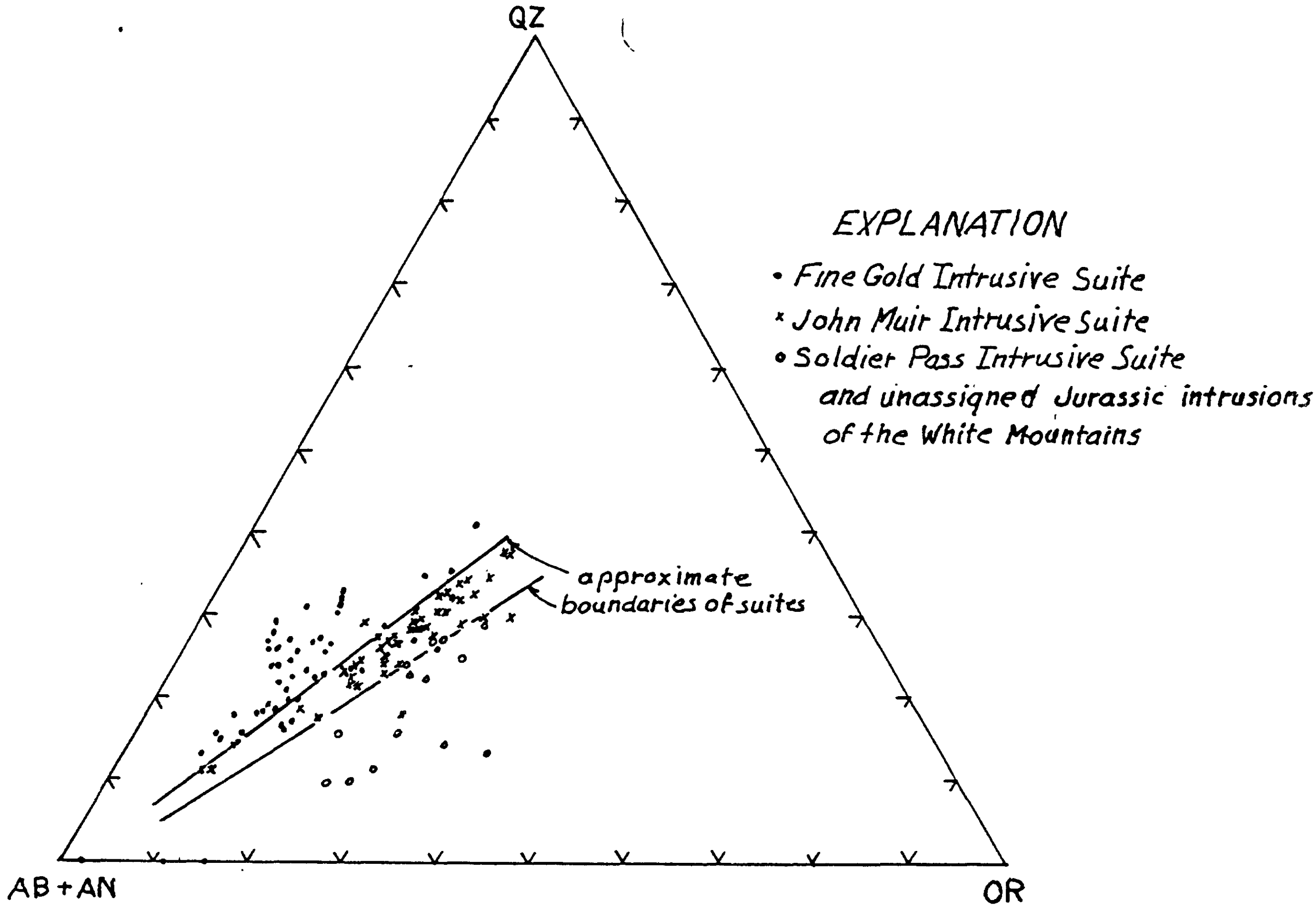


FIGURE 28.--Norms of representative intrusive suites on diagram representing salic tetrahedron collapsed on Qz-Or hinge line so that An an Ab coincide.

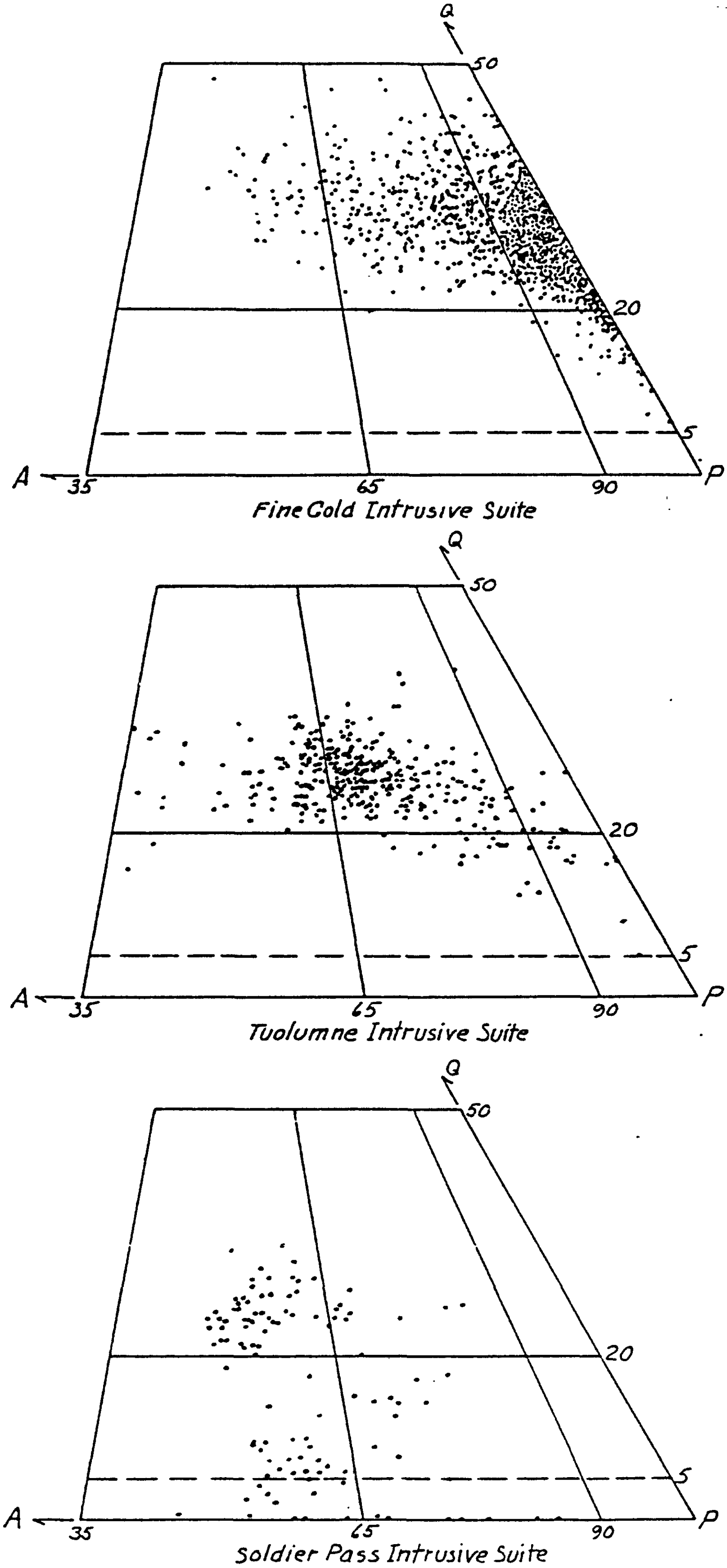


FIGURE 29.--Modes of representative calcic, calc-alkalic, and alkali-calcic intrusive suites on Q-A-P diagram.

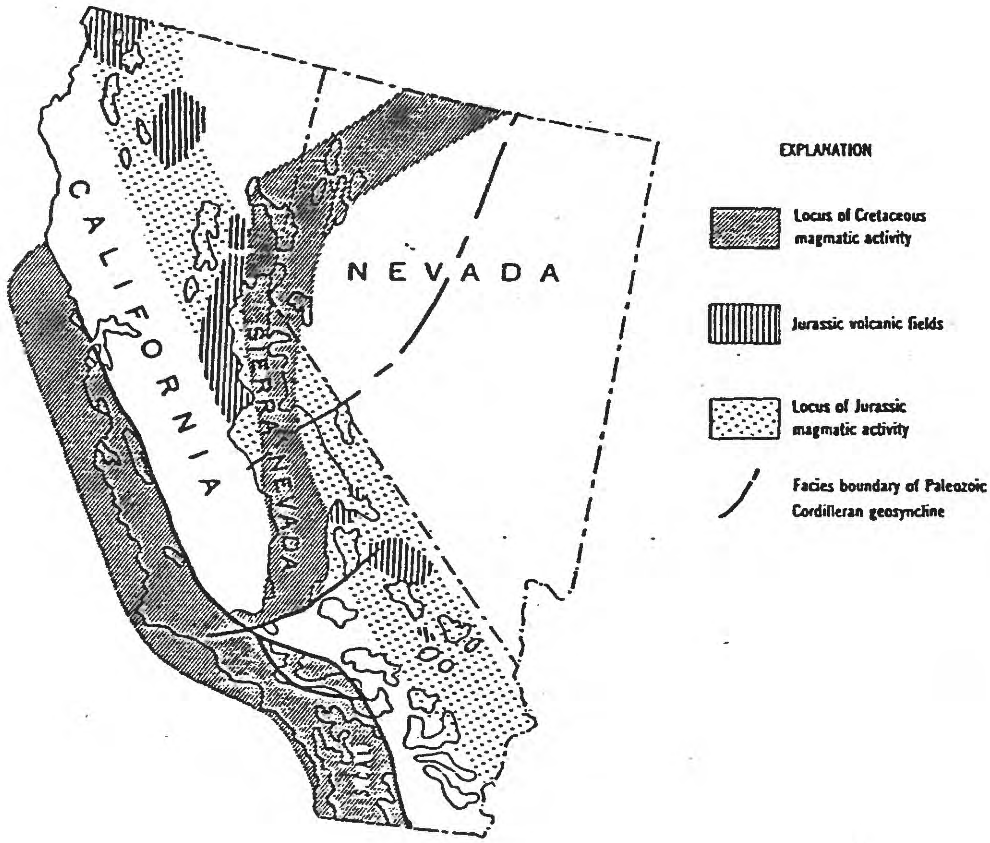


FIGURE 30.--Loci of Jurassic and Cretaceous plutonism in California and Nevada. Modified from Kistler and others (1971).

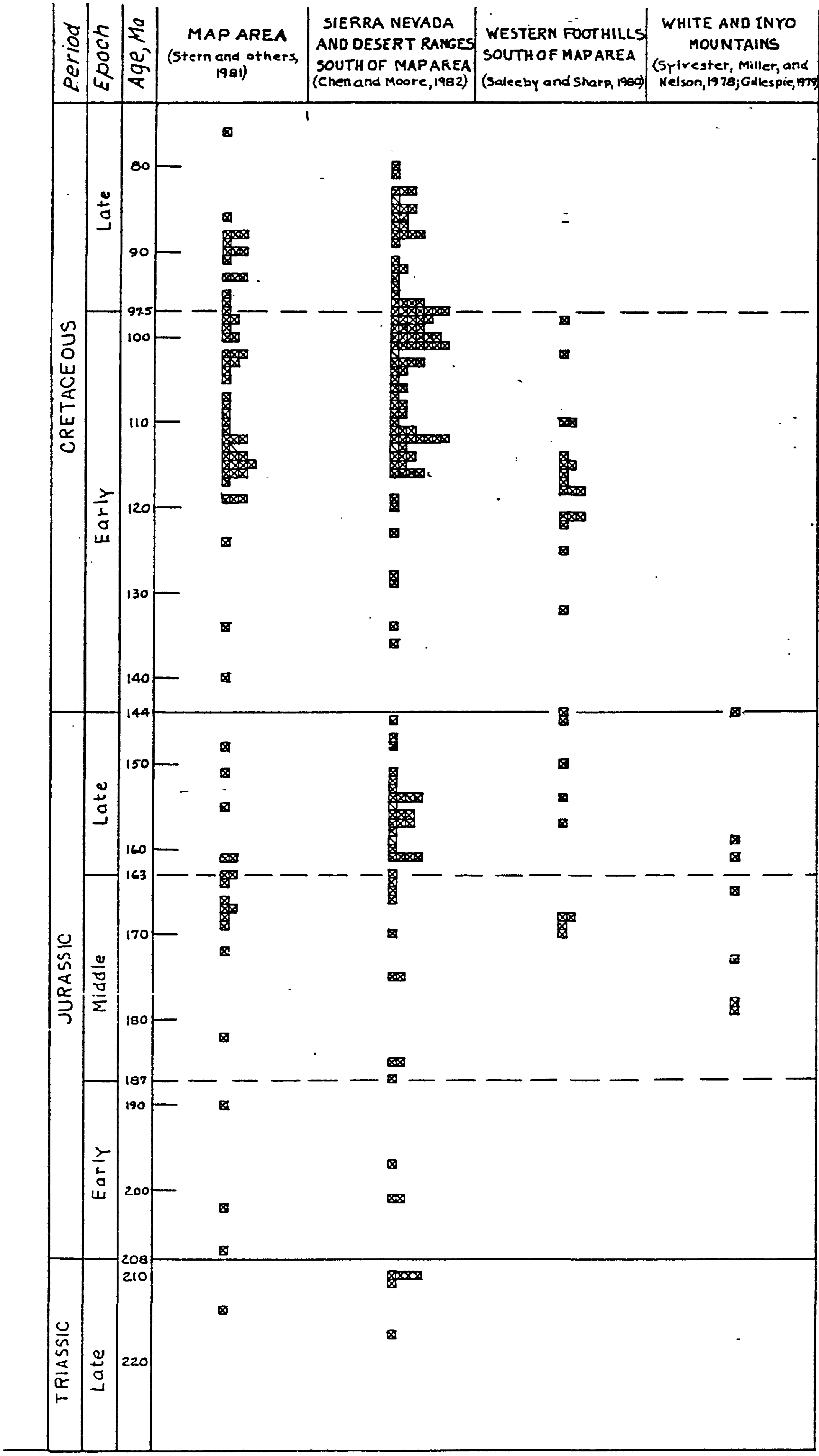


FIGURE 31.--Summary of U-Pb ages published prior to 1986 on Sierra Nevada granitoids. Ages of period and epoch boundaries from Palmer (1983).

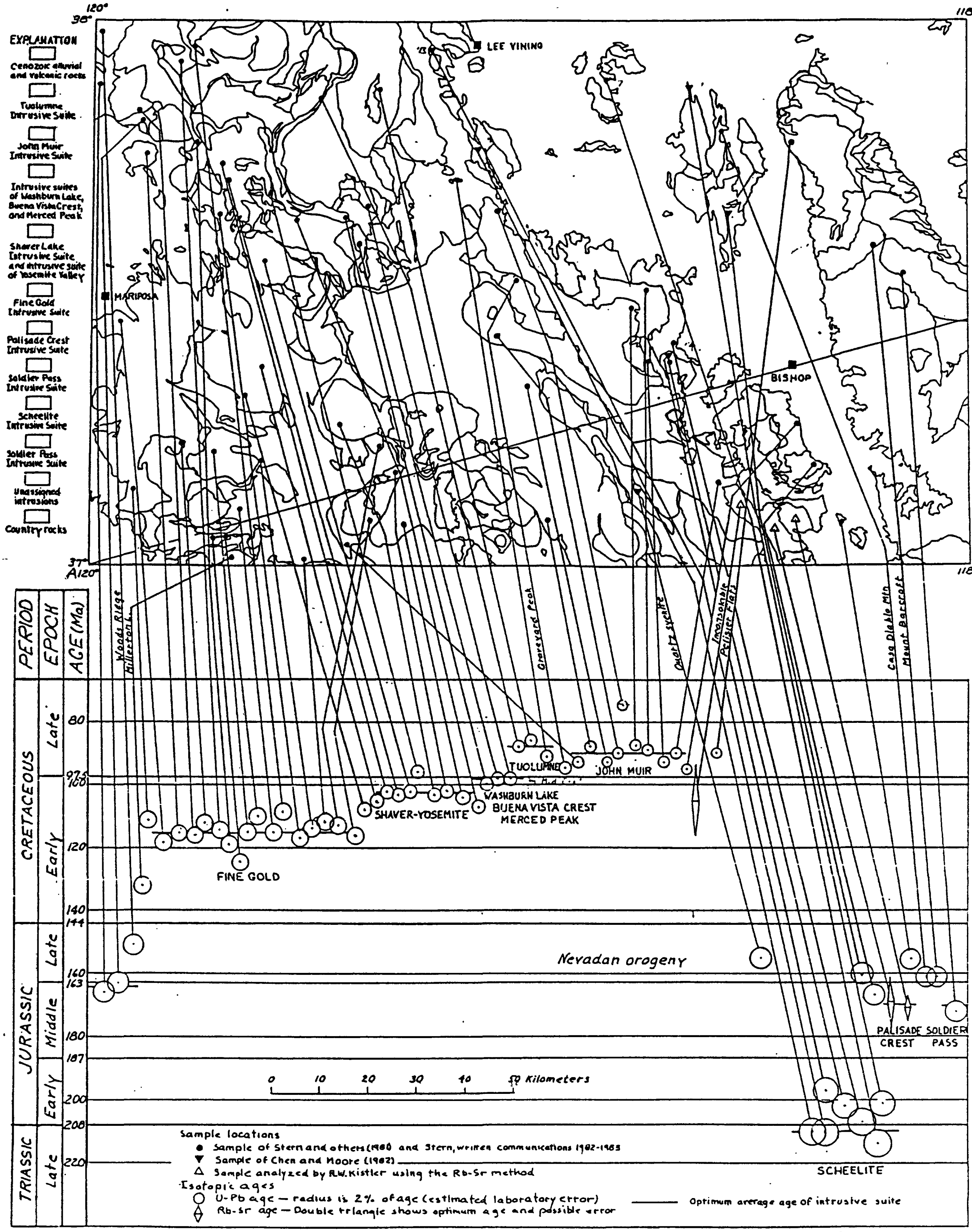


FIGURE 32.--Simplified geologic map of Mariposa 1° by 2° quadrangle showing distribution of country rocks, intrusive suites, unassigned intrusions, and localities of samples dated by the U-Pb and Rb-Sr methods. Contacts within intrusive suites and between unassigned intrusion are shown to facilitate identification of units sampled on plate 1. Chart shows U-Pb and Rb-Sr ages. Line A-B on map is line of profiles shown in figures 46, 47, 48, and 57.

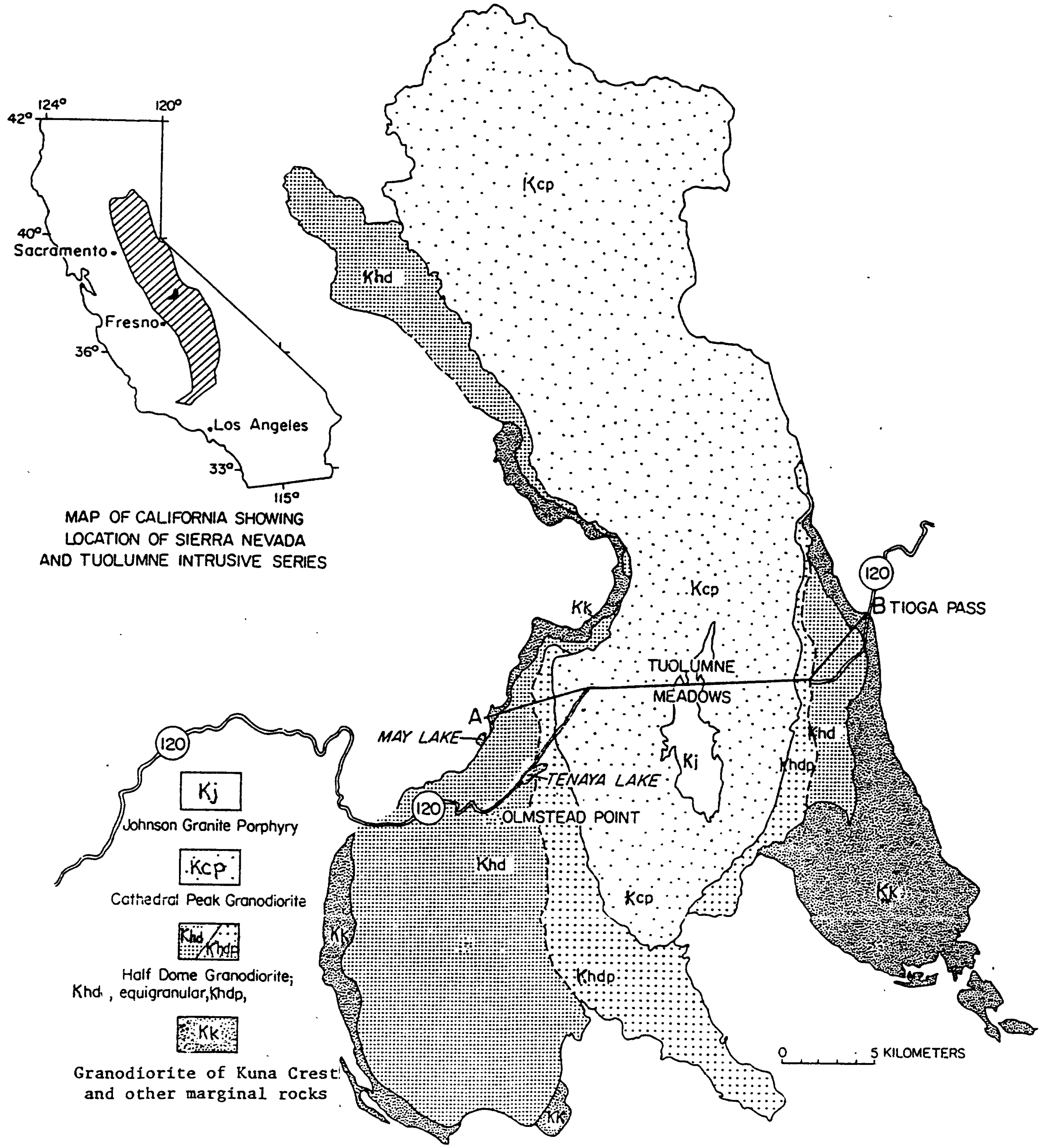


FIGURE 33.--Generalized geologic map of Tuolumne Intrusive Suite showing location of traverse A-B shown in figures 35, 36, and 37. Dashed contact between Khd and Khdp is gradational.

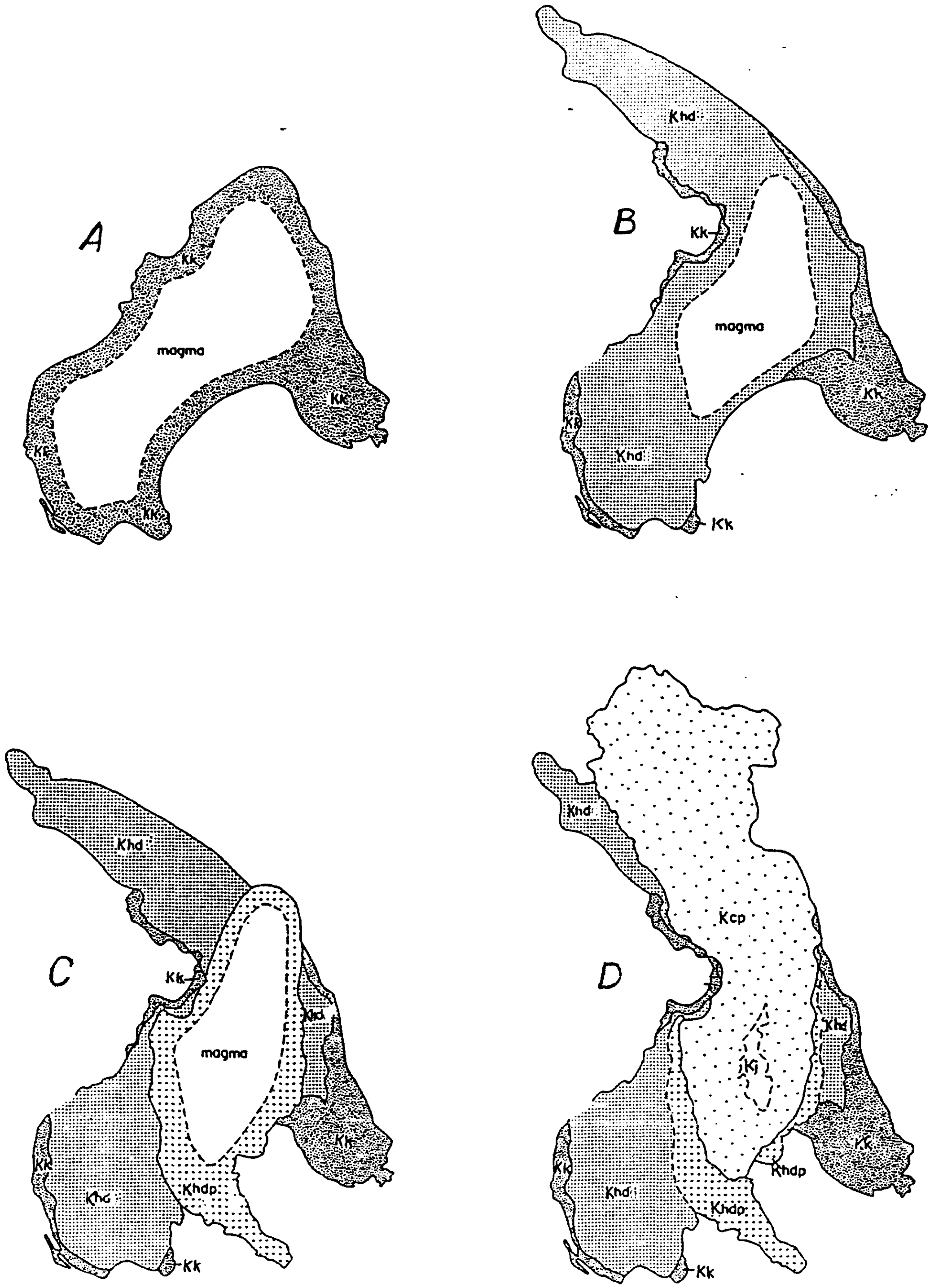


FIGURE 34.--Diagrams showing stages in emplacement of Tuolumne Intrusive Suite (from Bateman and Chappell, 1979). A, Initial intrusion and solidification of granodiorite of Kuna Crest and other marginal granitoids (Kk); B, Surge of magma followed by solidification of equigranular facies of Half Dome Granodiorite (Khd); C, Second surge of magma followed by solidification of megacrystic facies of Half Dome Granodiorite (Khdp); D, Third surge of magma followed by solidification of Cathedral Peak Granodiorite (Kcp). Eruption and solidification of Johnson Granite Porphyry (Kj) was final stage of emplacement.

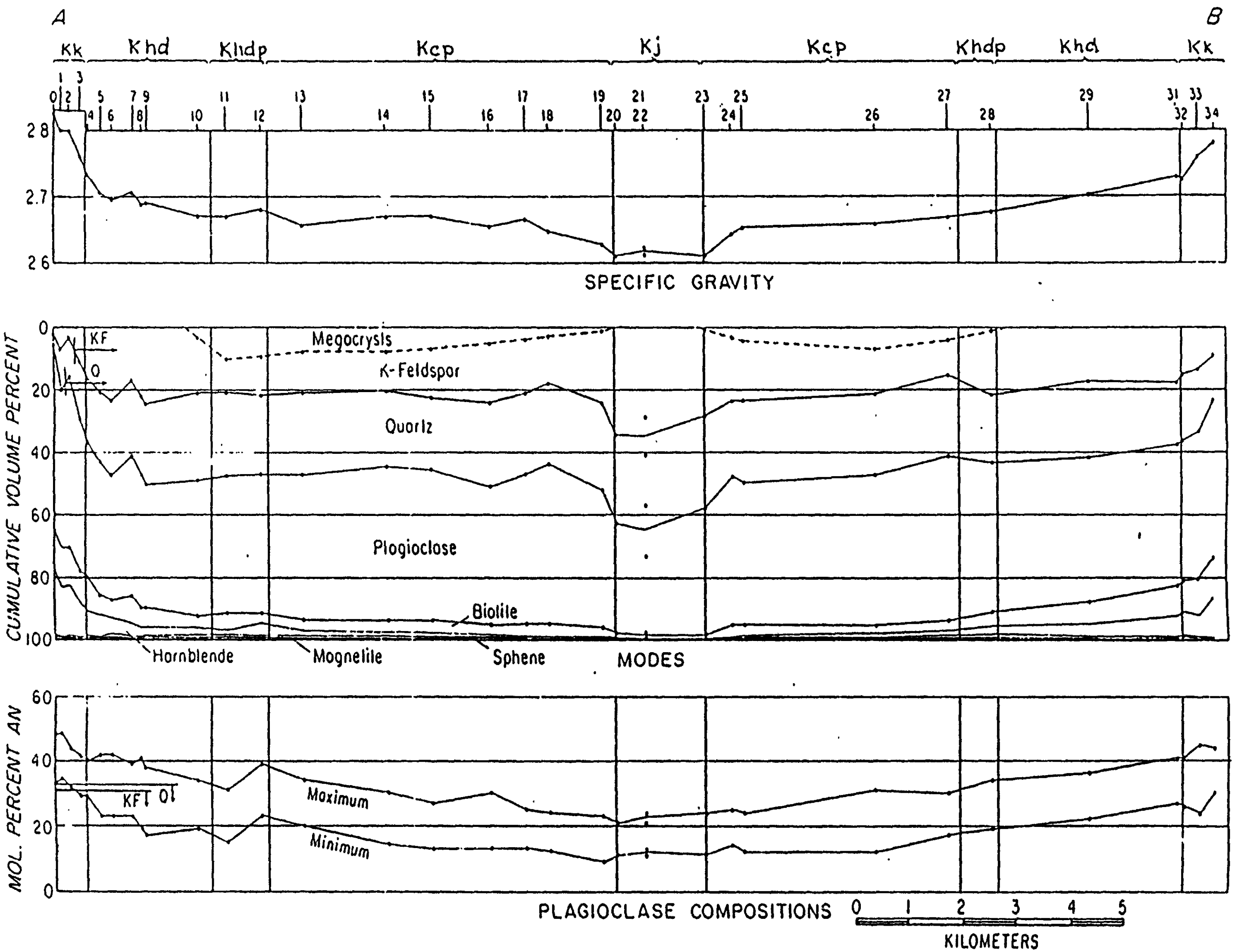


FIGURE 35.--Specific gravity, modes, and plagioclase compositions along traverse A-B figure 33. Kk, granodiorite of Kuna Crest and other marginal granitoids; Khd , equigranular facies of the Half Dome Granodiorite; Khdp, megacrystic facies of the Half Dome Granodiorite; Kcp, Cathedral Peak Granodiorite; Kj, Johnson Granite Porphyry. Q and Kf indicate composition of plagioclase crystallizing at the beginning of crystallization of quartz and alkali feldspar.

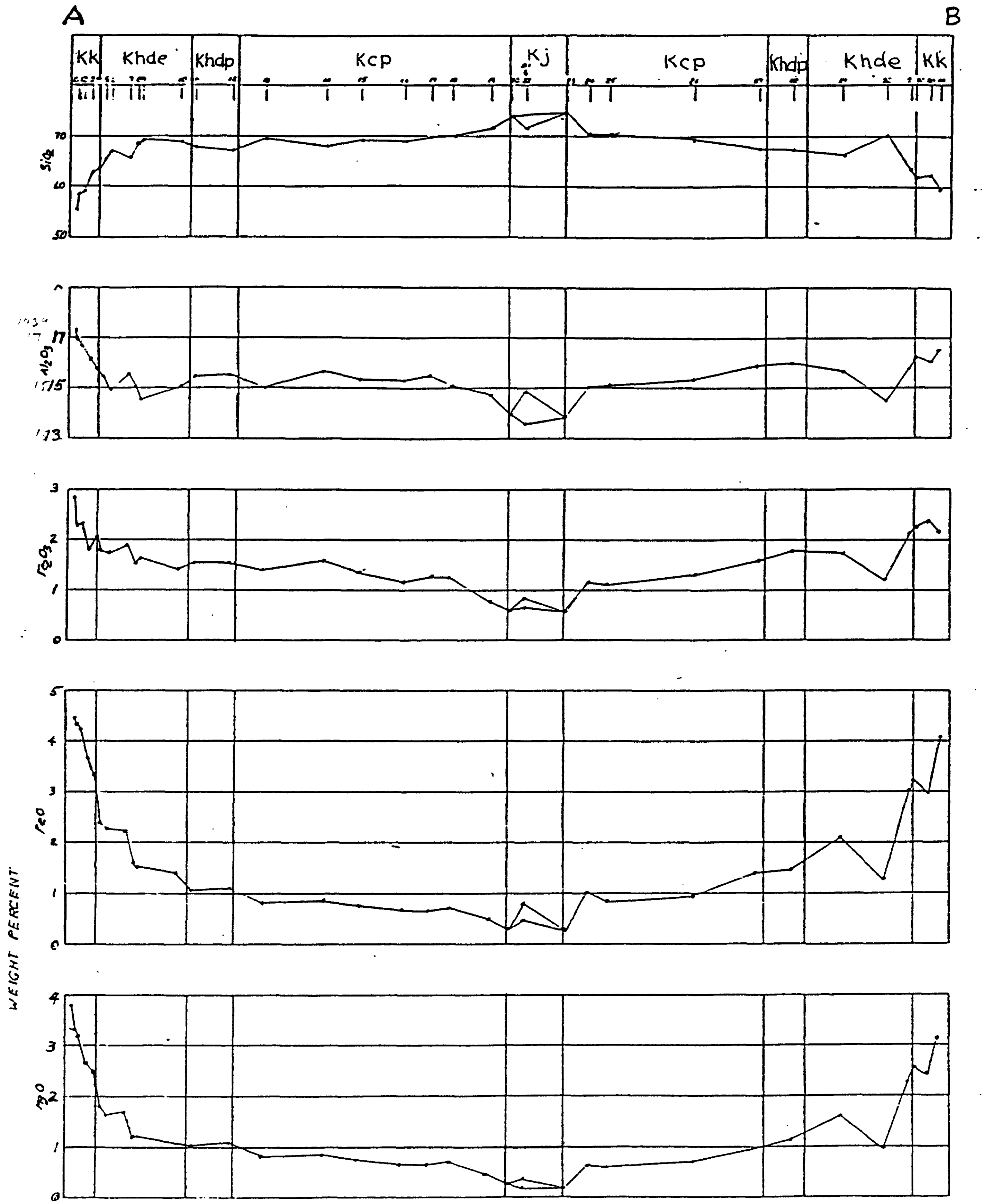
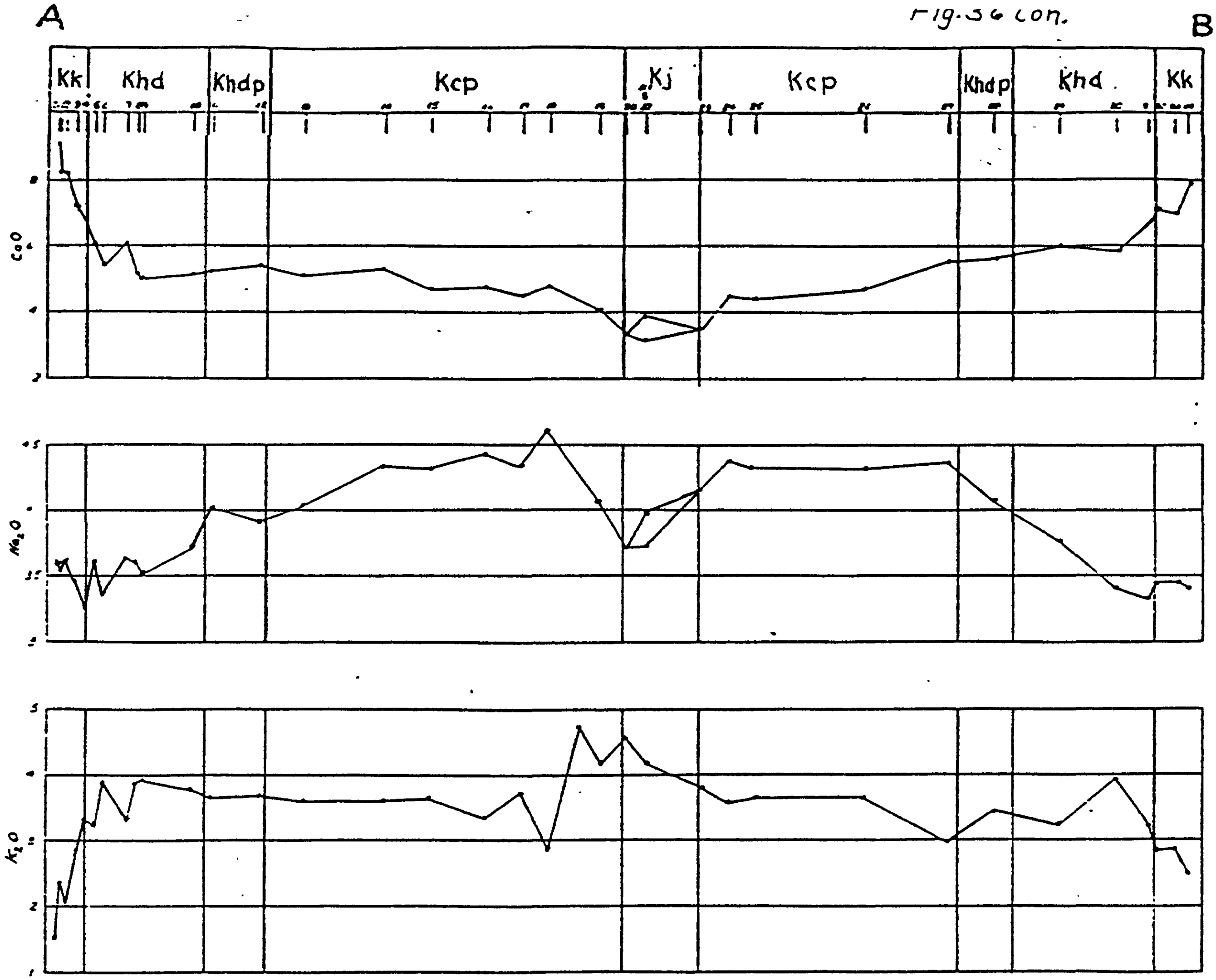


FIGURE 36.--Variations of major oxides along traverse A-B, figure 33. Vertical scale is logarithmic. Kk, granodiorite of Kuna Crest and other marginal granitoids; Khde, equigranular facies of the Half Dome Granodiorite; Khdp, porphyritic facies of the Half Dome Granodiorite; Kcp, Cathedral Peak Granodiorite; Kj, Johnson Granite Porphyry. Numbers along traverse are sample localities.


A
Fig. 56 con.
B
Kk
Khd
Khdp
Kcp
Kj
Kcp
Khdp
Khd
Kk
CaO
Na_2O
K_2O

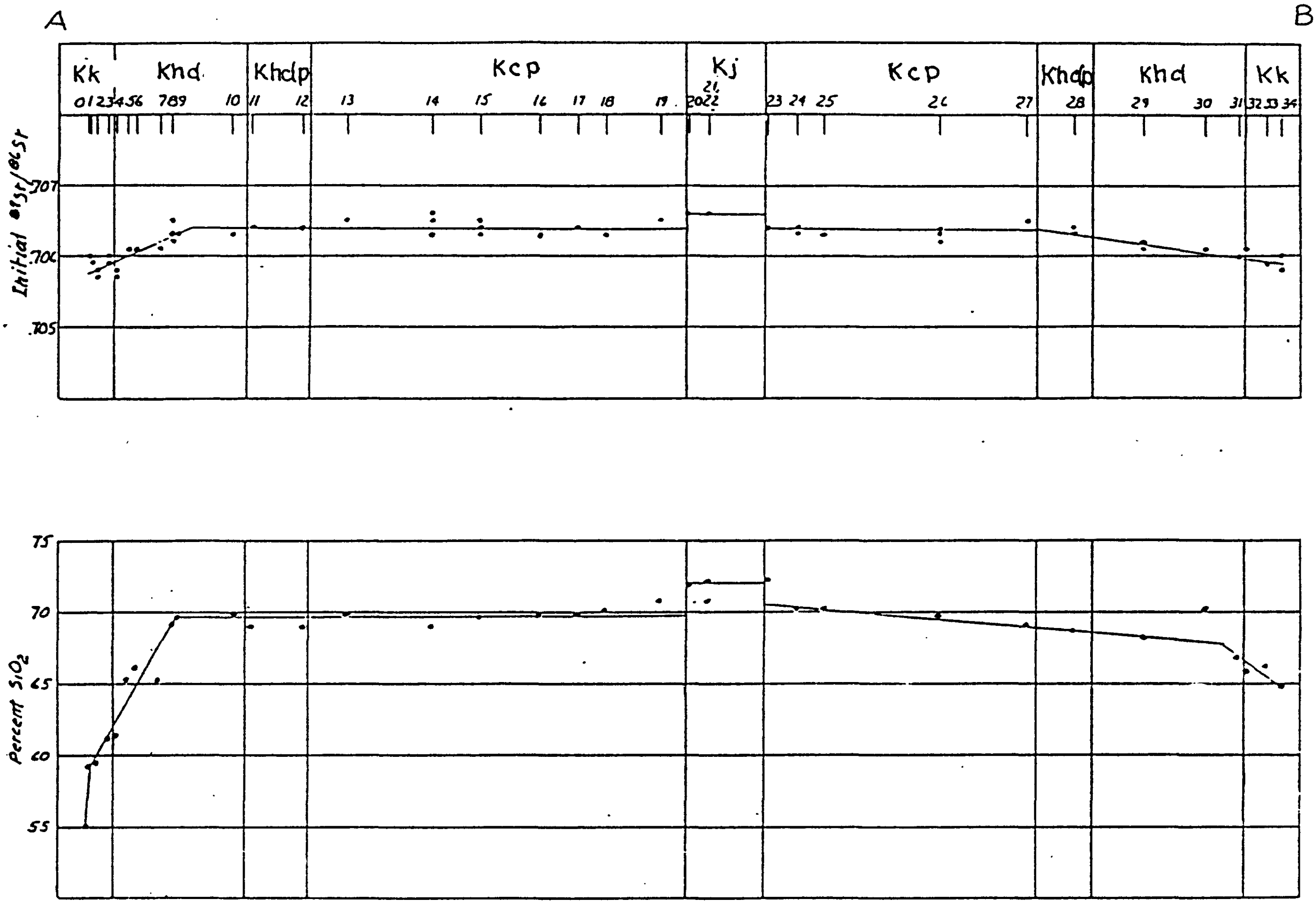


FIGURE 37.--Variation of initial $^{87}Sr/^{86}Sr$ (assuming an age of 87 Ma) and SiO_2 along traverse A-B, figure 33. Kk, granodiorite of Kuna Crest and other marginal granitoids; Khd , equigranular facies of the Half Dome Granodiorite; Khdp, porphyritic facies of the Half Dome Granodiorite; Kcp, Cathedral Peak Granodiorite; Kj, Johnson Granite Porphyry. Lines through points indicate general variation trends.

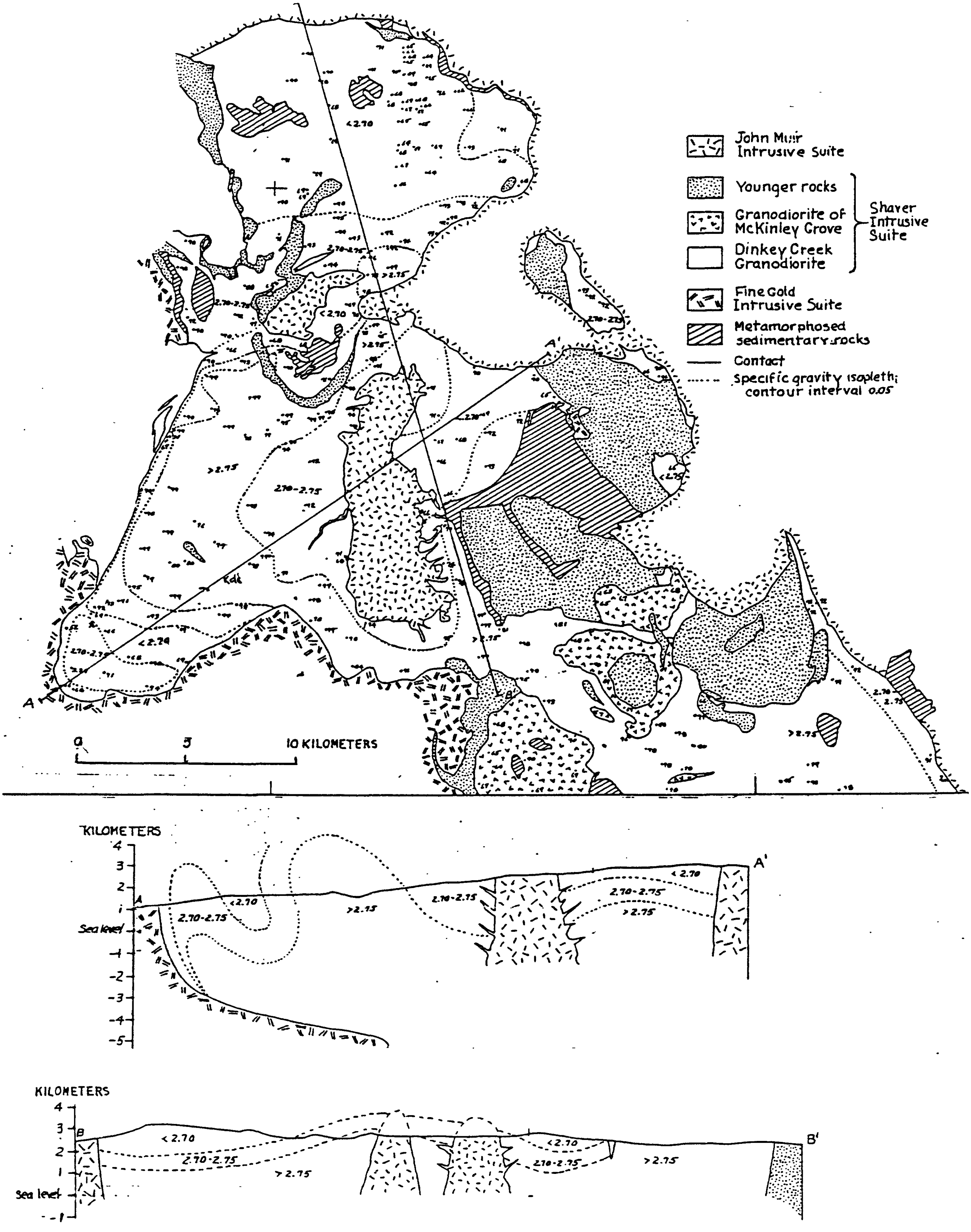


FIGURE 38.--Generalized geologic map of Dinkey Creek Granodiorite and associated rocks showing variations in specific gravity. Sample localities shown by dots. Cross sections are conjectural.

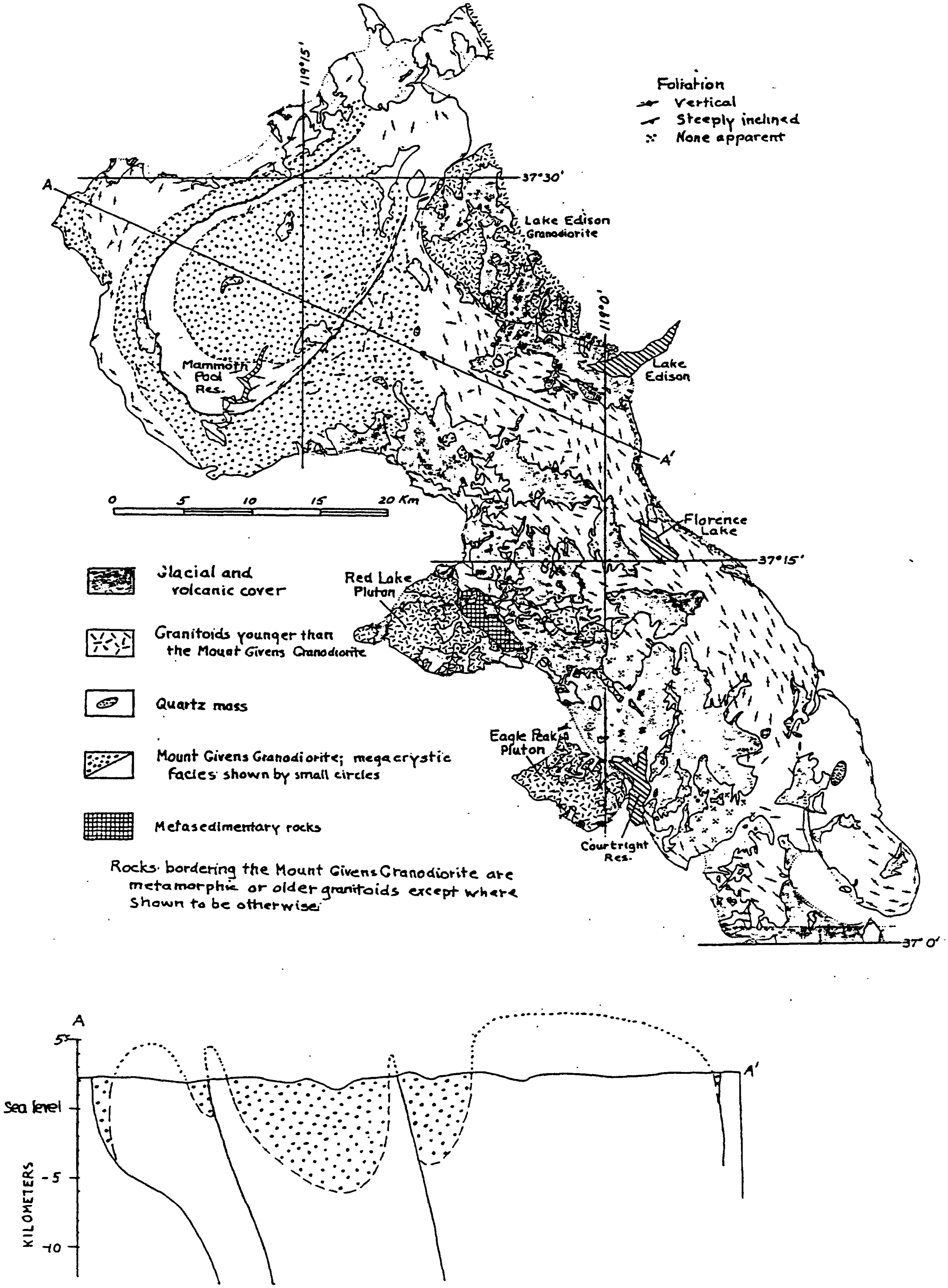


FIGURE 39.--Mount Givens Granodiorite showing foliation patterns. Cross section is conjectural.

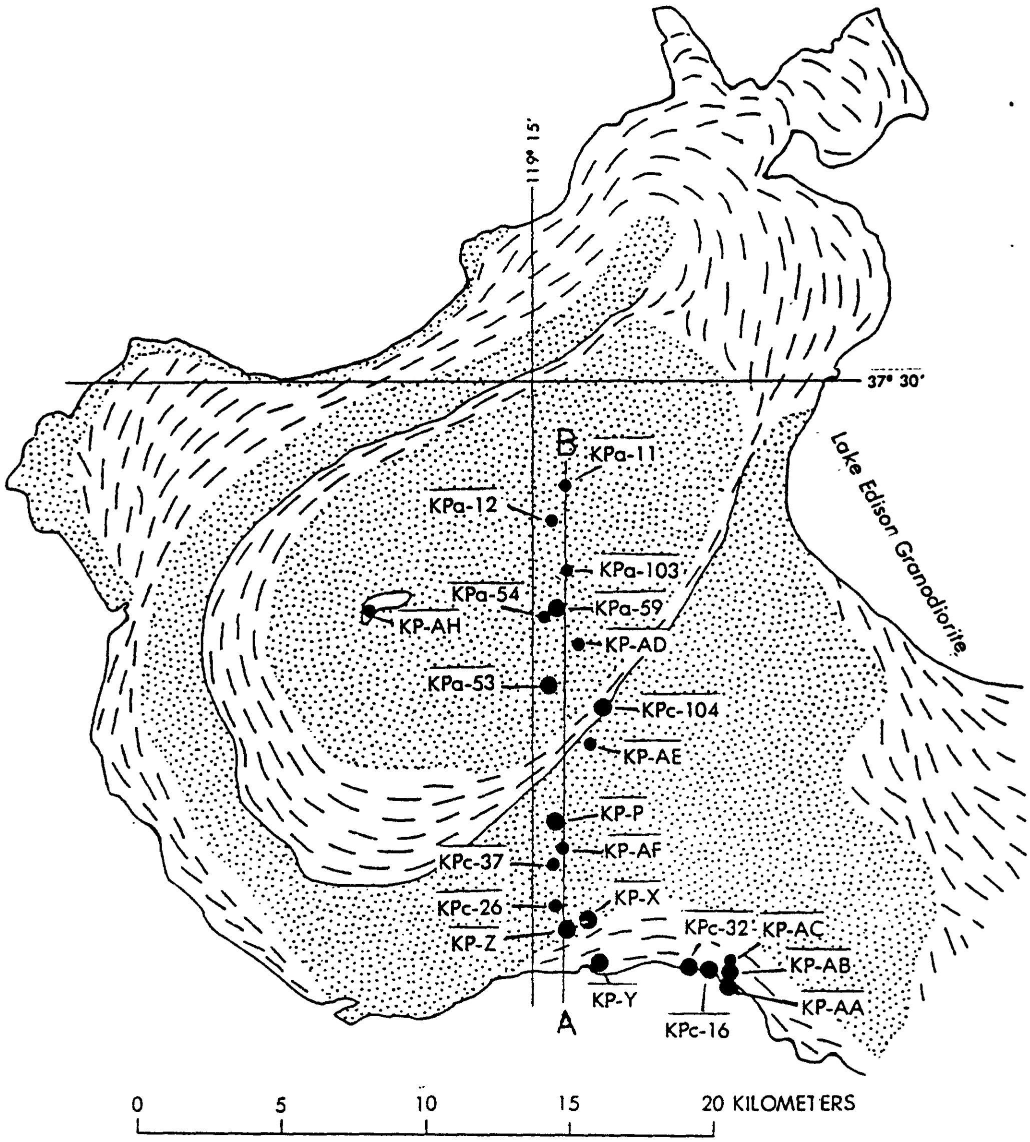


FIGURE 40.--Map of north end of Mount Givens Granodiorite showing traverse A-B and sample localities (dots). Dot labels referred to in figure 41 and text. Megacrystic facies is shown by stipple. Unpatterned area at sample locality AP-AH is the aplitic facies of the Mount Givens at Jackass Rock.

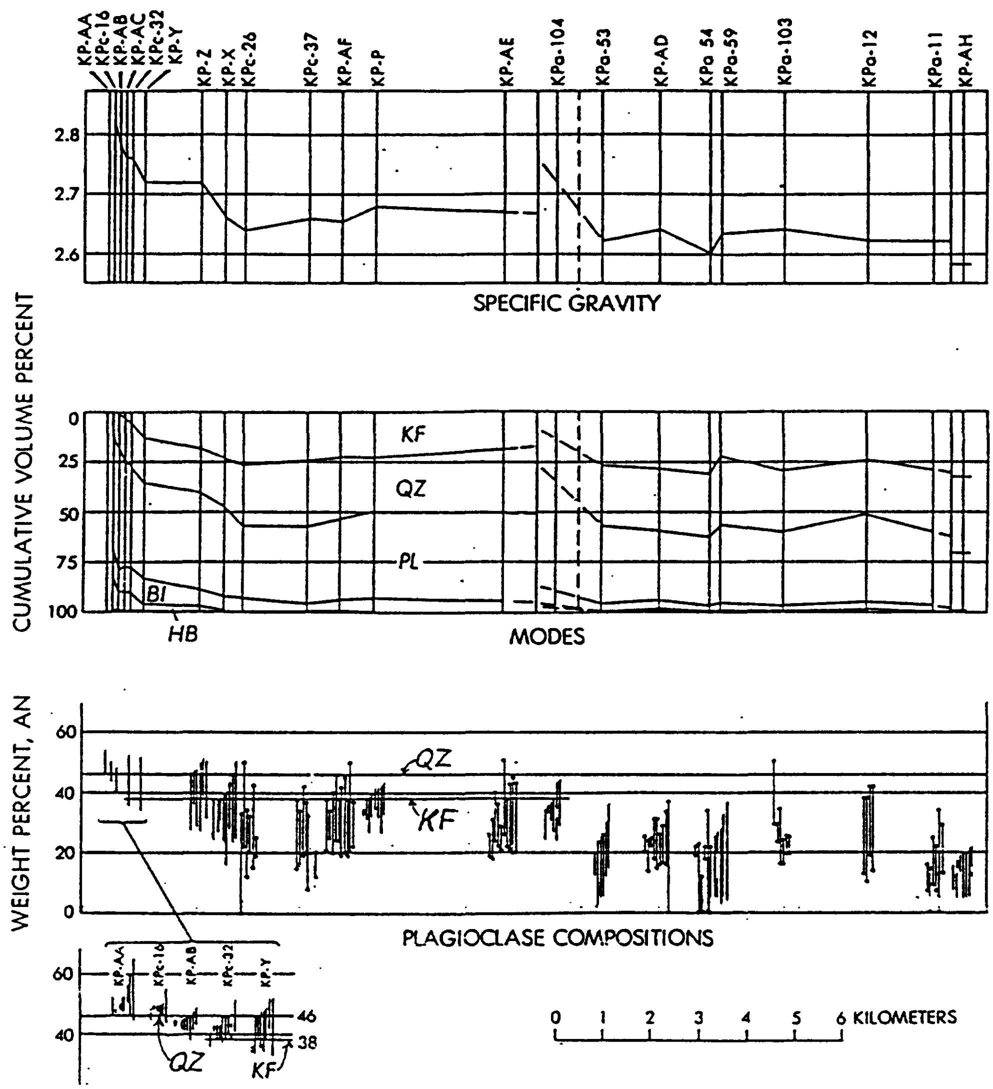


FIGURE 41.--Specific gravities, modes and plagioclase compositions of samples collected from Mount Givens Granodiorite along traverse A-B, figure 40. KF, alkali feldspar; QZ, quartz; PL, plagioclase; BI, biotite; HB, hornblende. Microprobe determinations of plagioclase are shown by heavier lines and optical determinations by lighter 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 the beginning of crystallization of quartz and alkali feldspar.

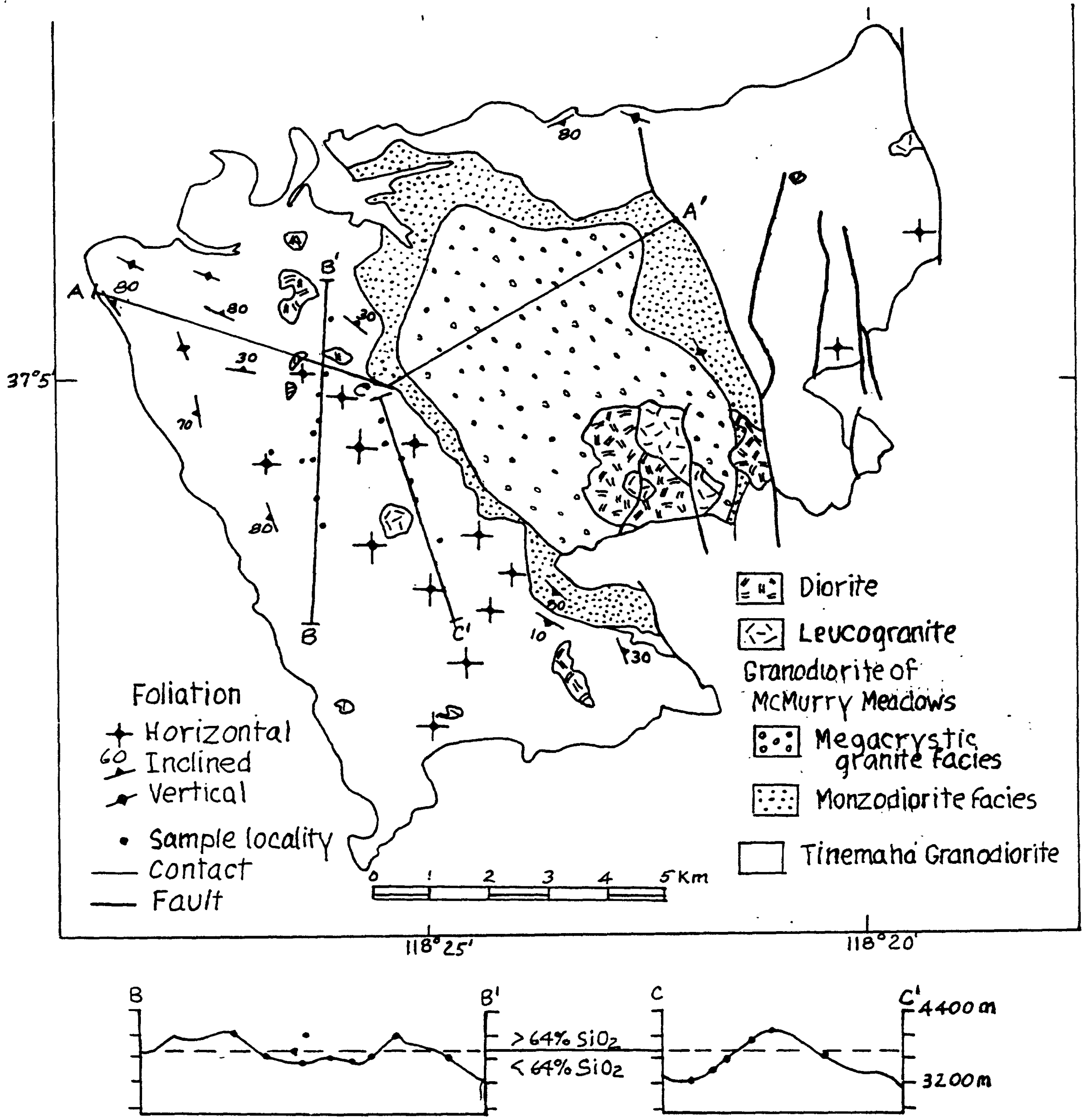


FIGURE 42.--Generalized geologic map of Tinemaha Granodiorite and granodiorite of McMurry Meadows showing locations of constant-altitude traverse A-A' (see figure 43) and of traverses B-B and C-C', which ascend to higher altitudes. Samples collected along traverses B-B' and C-C' are also shown on profiles. From Sawka (1985).

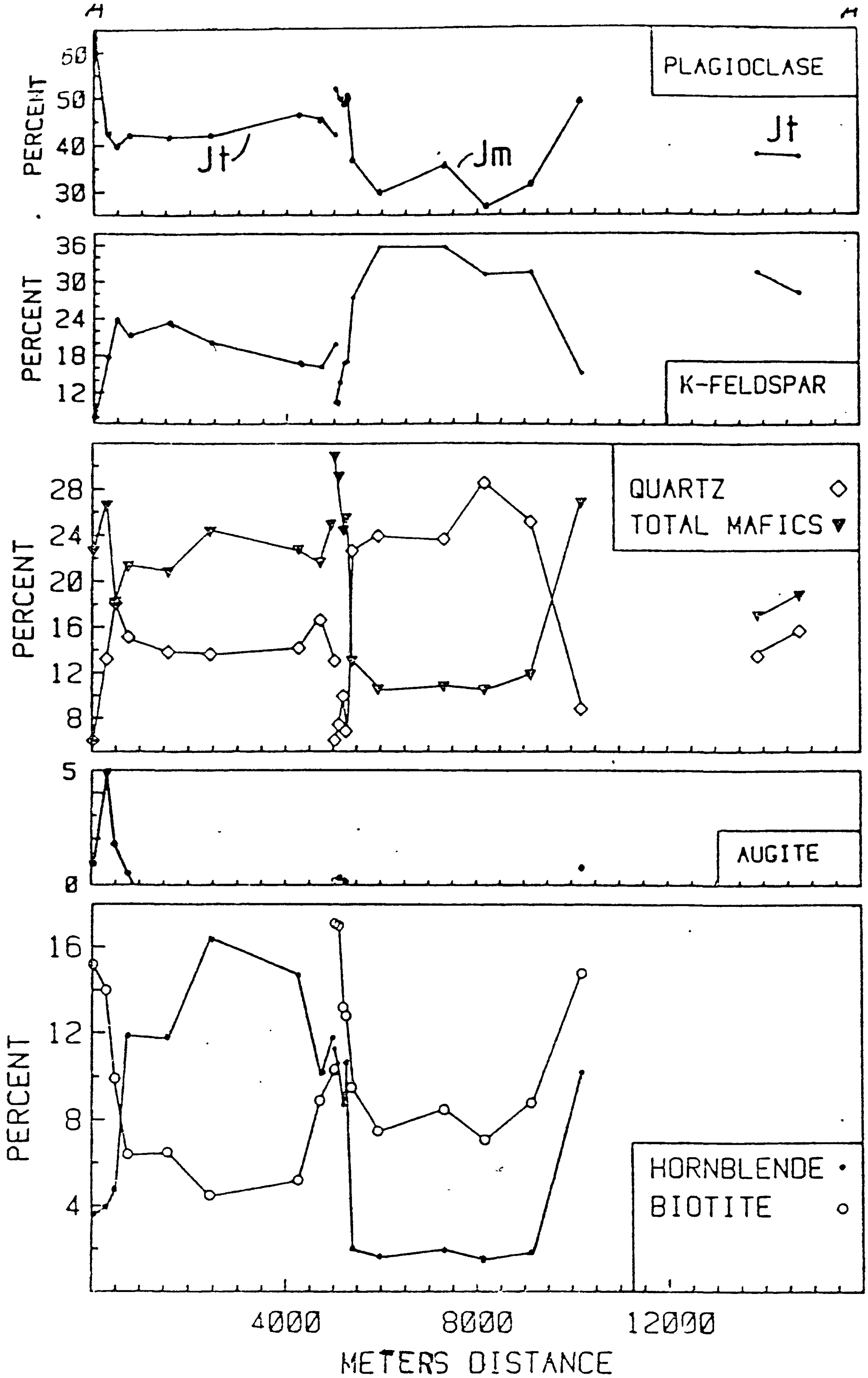


FIGURE 43.--Modal abundances of minerals in Tinemaha Granodiorite (Jt) and granodiorite of McMurry Meadows (Jm) along constant-altitude traverse A-A' of figure 42. From Sawka (1985).

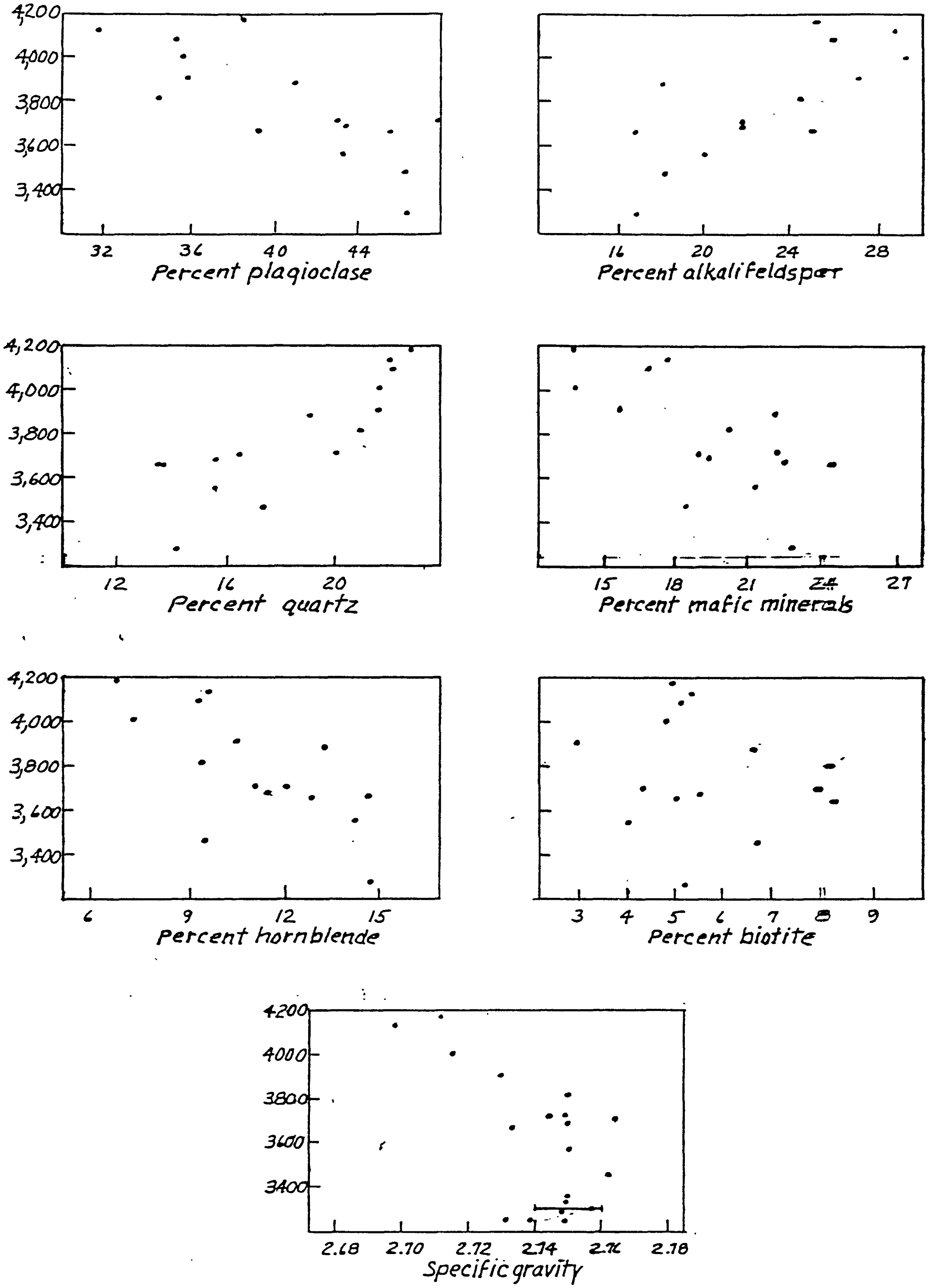


FIGURE 44.--Variation in modal abundances of minerals and specific gravity with altitude within interior of western lobe of Tinemaha Granodiorite. Bar shows range of specific gravity along equal-altitude traverse A-A' (figs. 42, 43). From Sawka (1985).

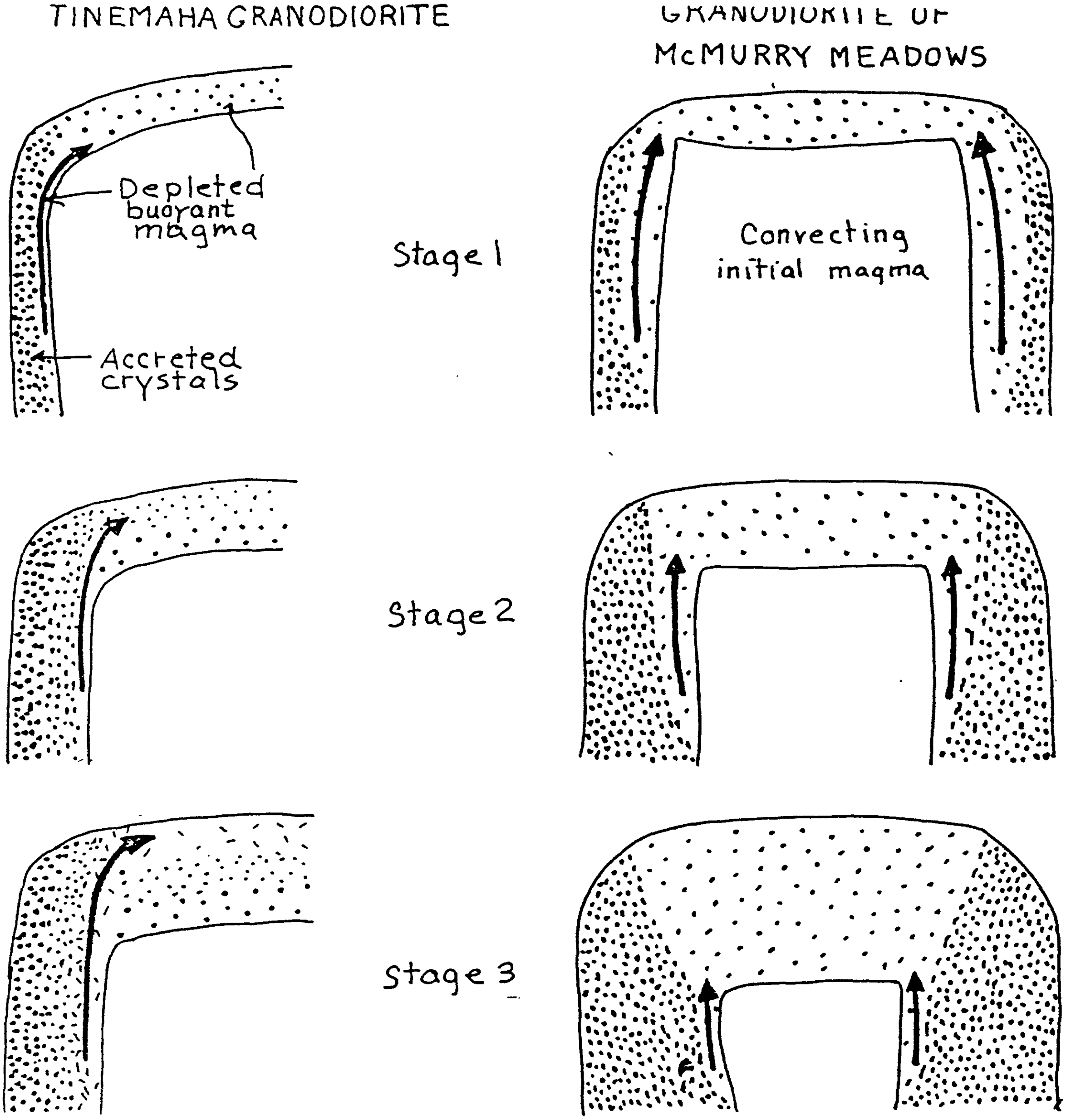


FIGURE 45.--Sidewall accretion models for western lobe of Tinemaha Granodiorite and granodiorite of McMurry Meadows. Arrows show direction of flow of depleted buoyant magma. From Sawka (1985).

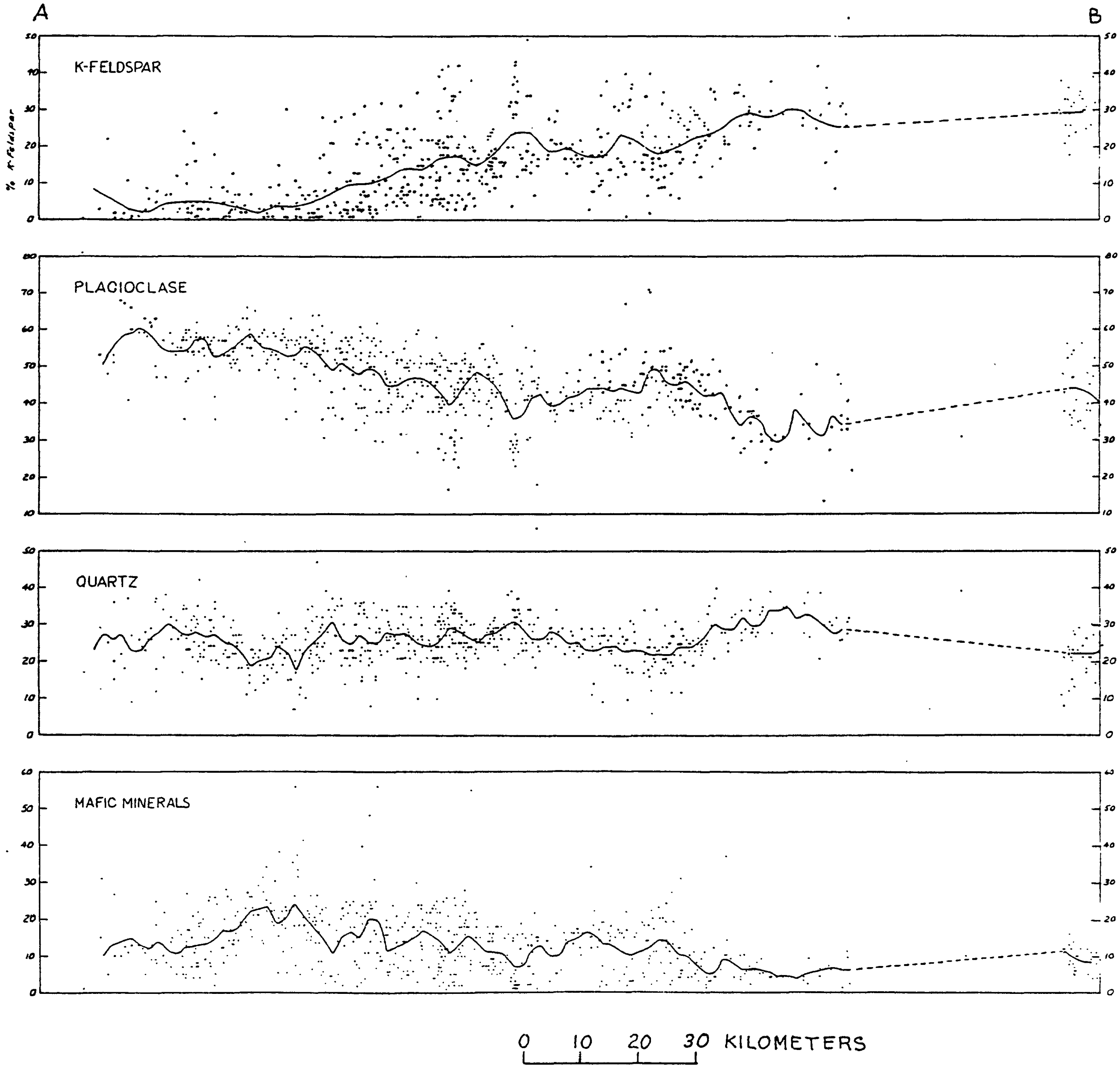
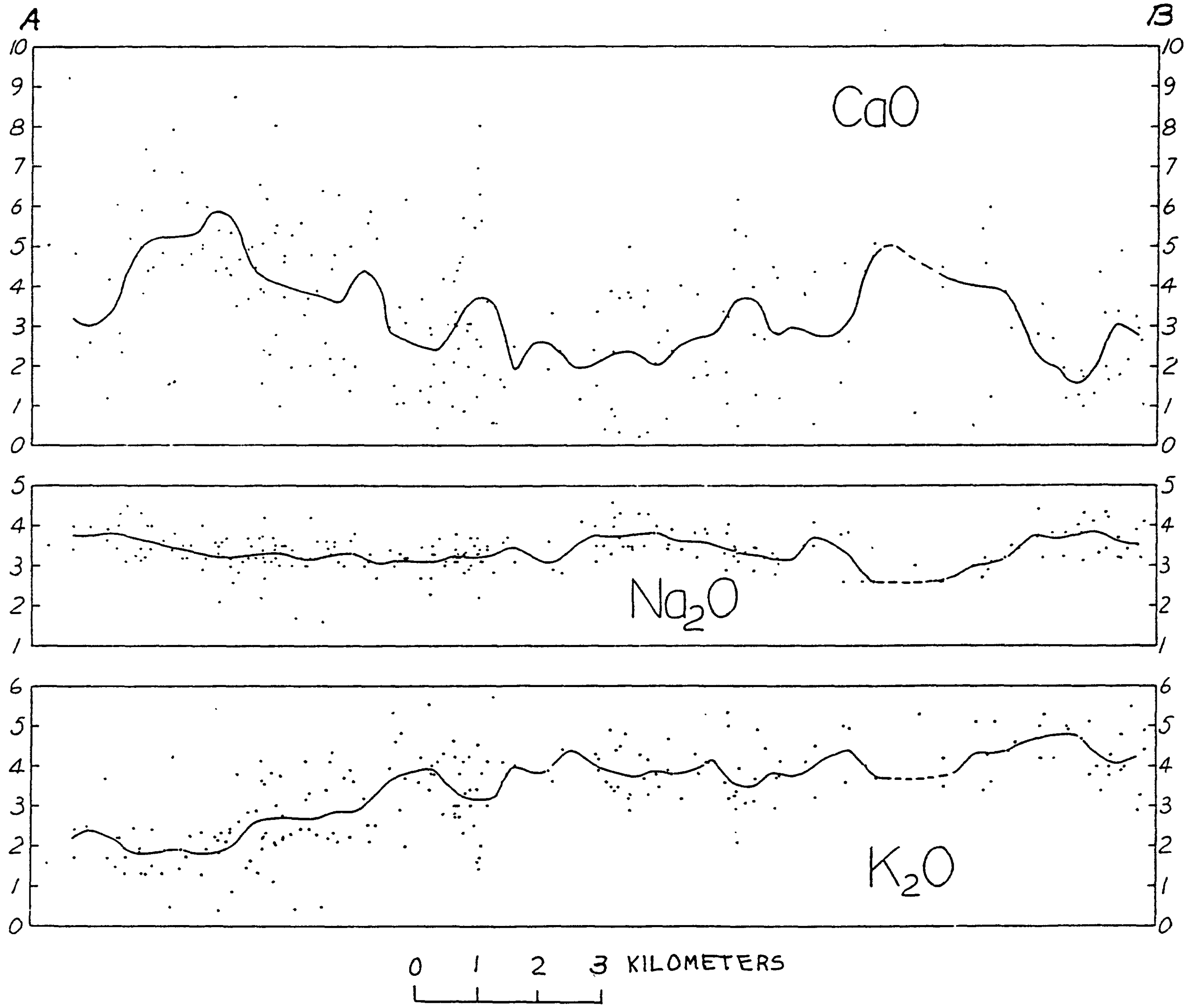


FIGURE 46.--Modal abundances of minerals in samples collected within 3 km of line A-B in figure 32. Moving-average curve constructed by averaging zones that overlap 50 percent. Curve dashed where alluvium conceals bedrocks, and no samples were collected.

Fig 47 con.



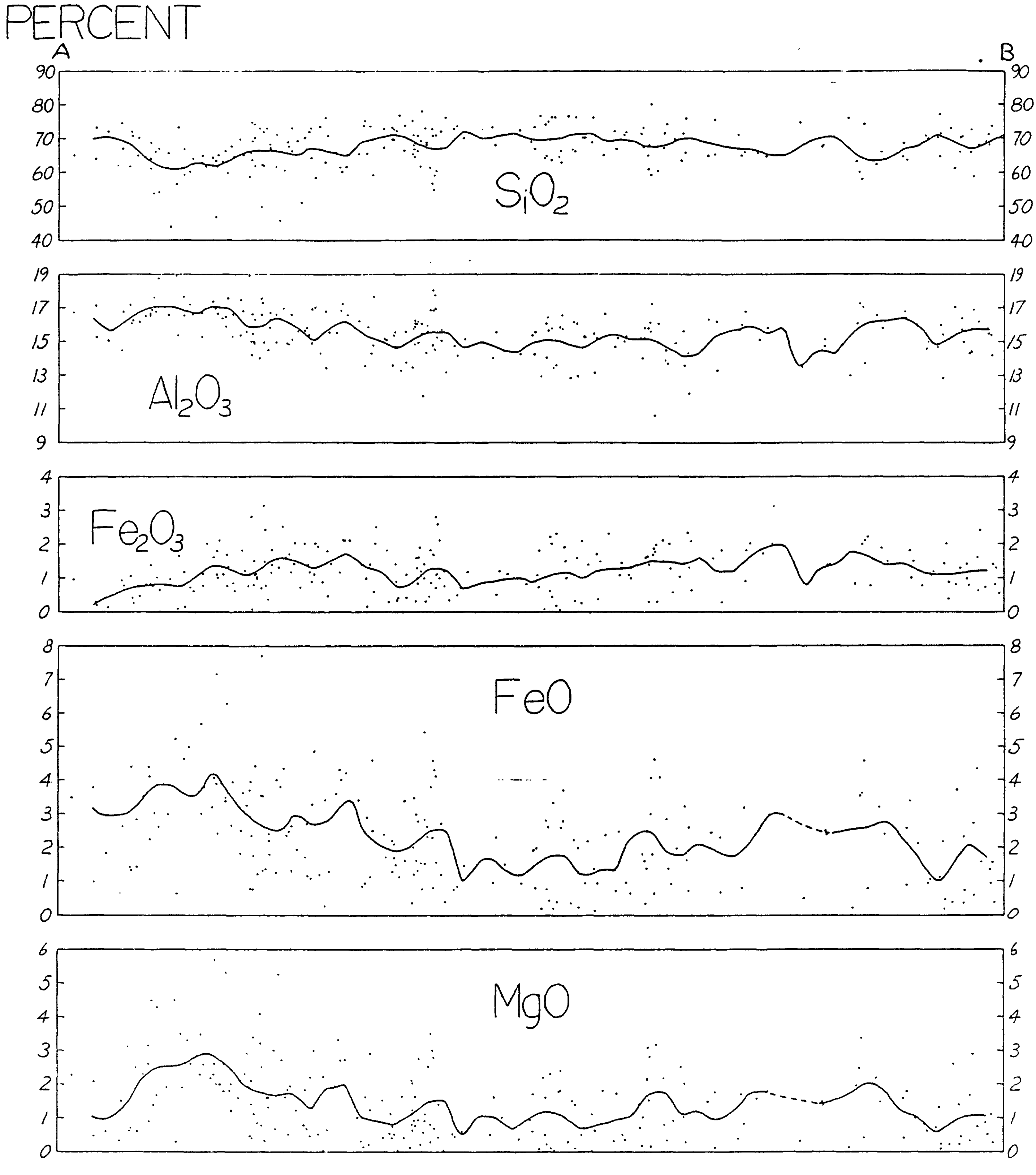


FIGURE 47.--Major oxide contents of samples projected on line A-B of of figure 32. Includes all samples collected within the Mariposa 1° by 2° quadrangle except those in the Yosemite area, which are distant from line A-B. Moving-average curves constructed by averaging zones that overlap 50 percent. Curve is dashed where few samples project on line A-B.

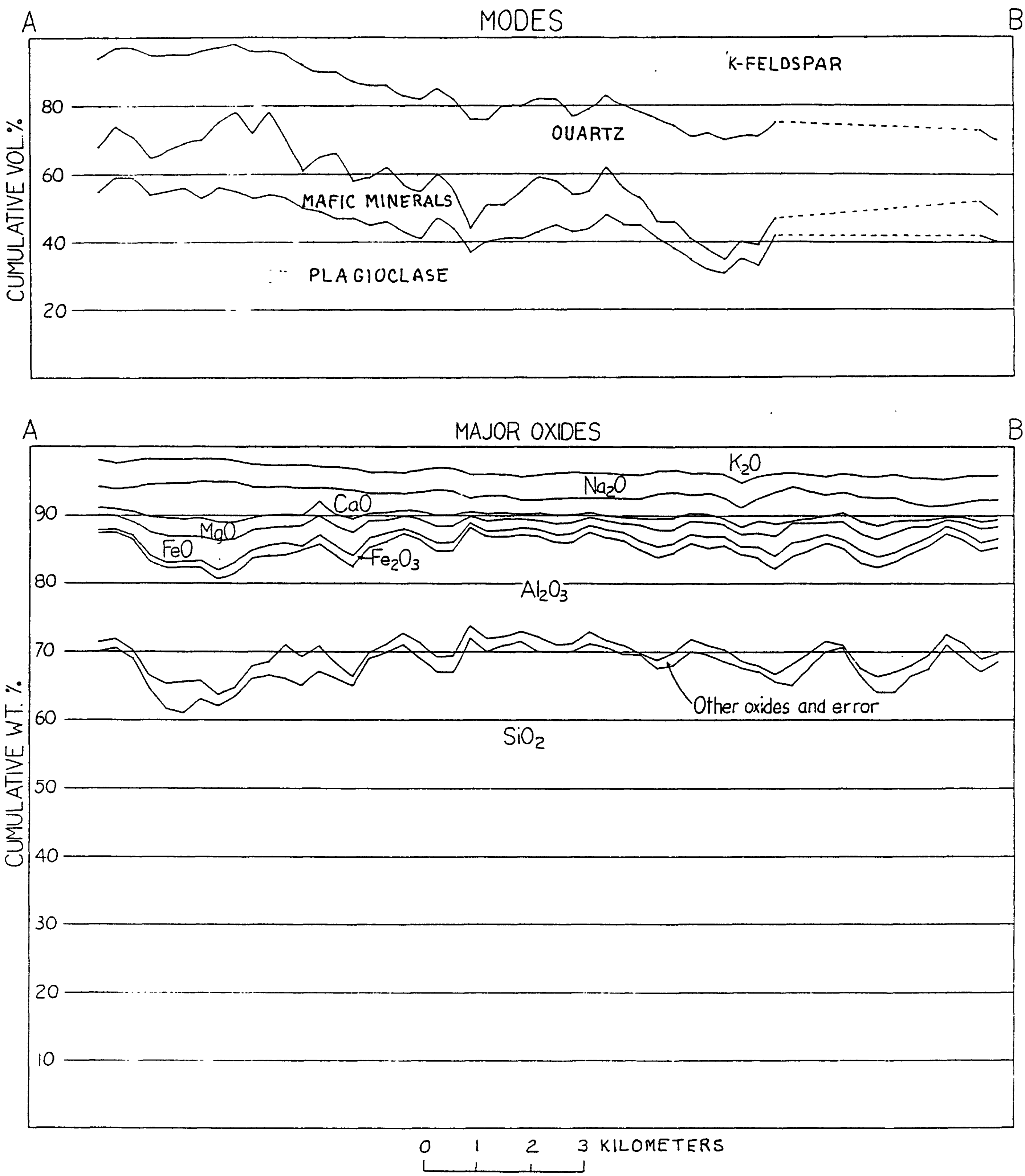


FIGURE 48.--Cumulative curves for modes and major oxides projected on line A-B in figure 32. Curves for modes are dashed where alluvium conceals bedrock, and no samples were collected.

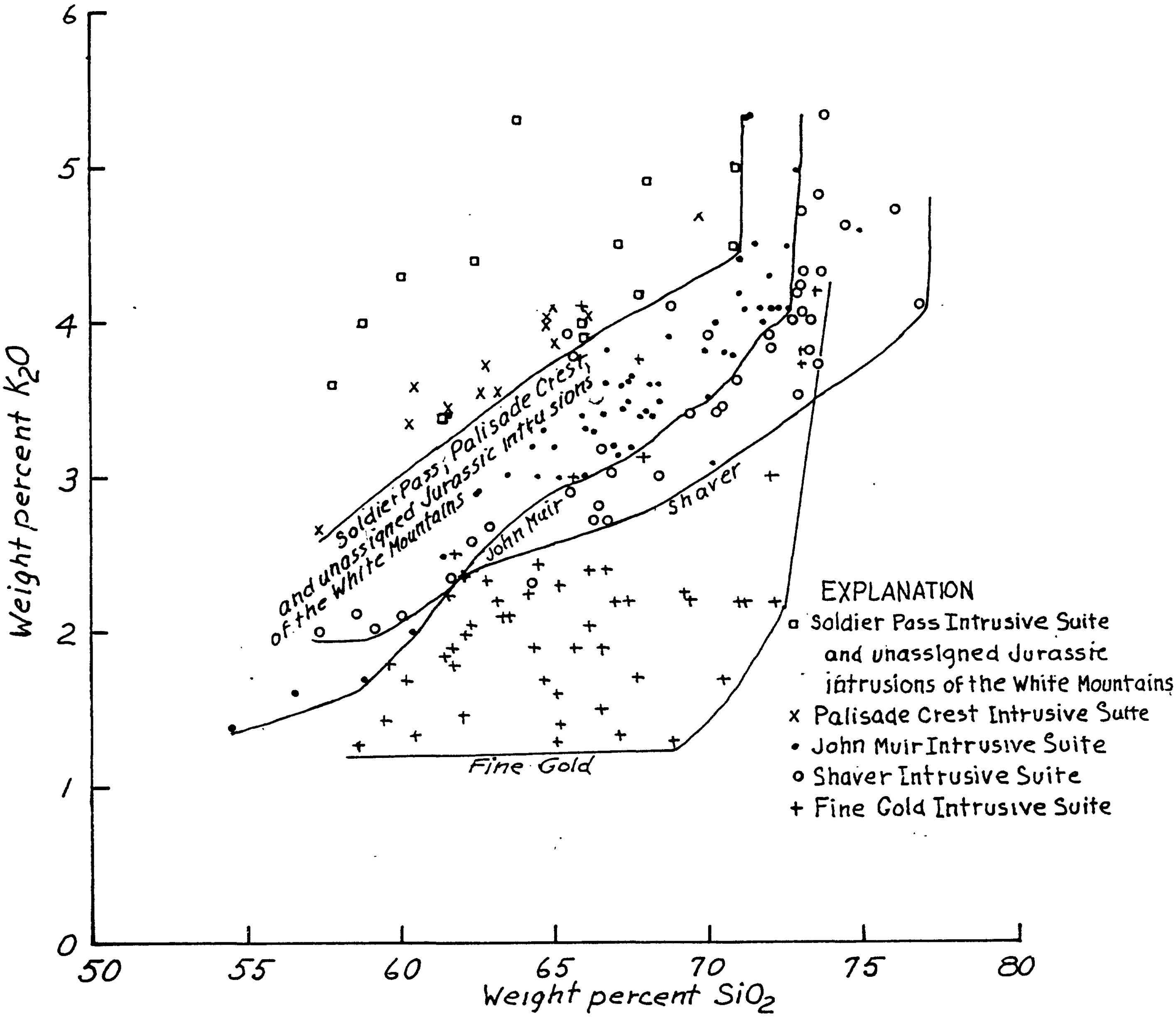


FIGURE 49.--Plots of K_2O against SiO_2 for representative intrusive suites. Solid lines are drawn at lower limit of weight-percent K_2O for each intrusive suite. Shows that K_2O/SiO_2 increases eastward in all suites, whereas SiO_2 generally ranges between ~50 and ~75 weight percent.

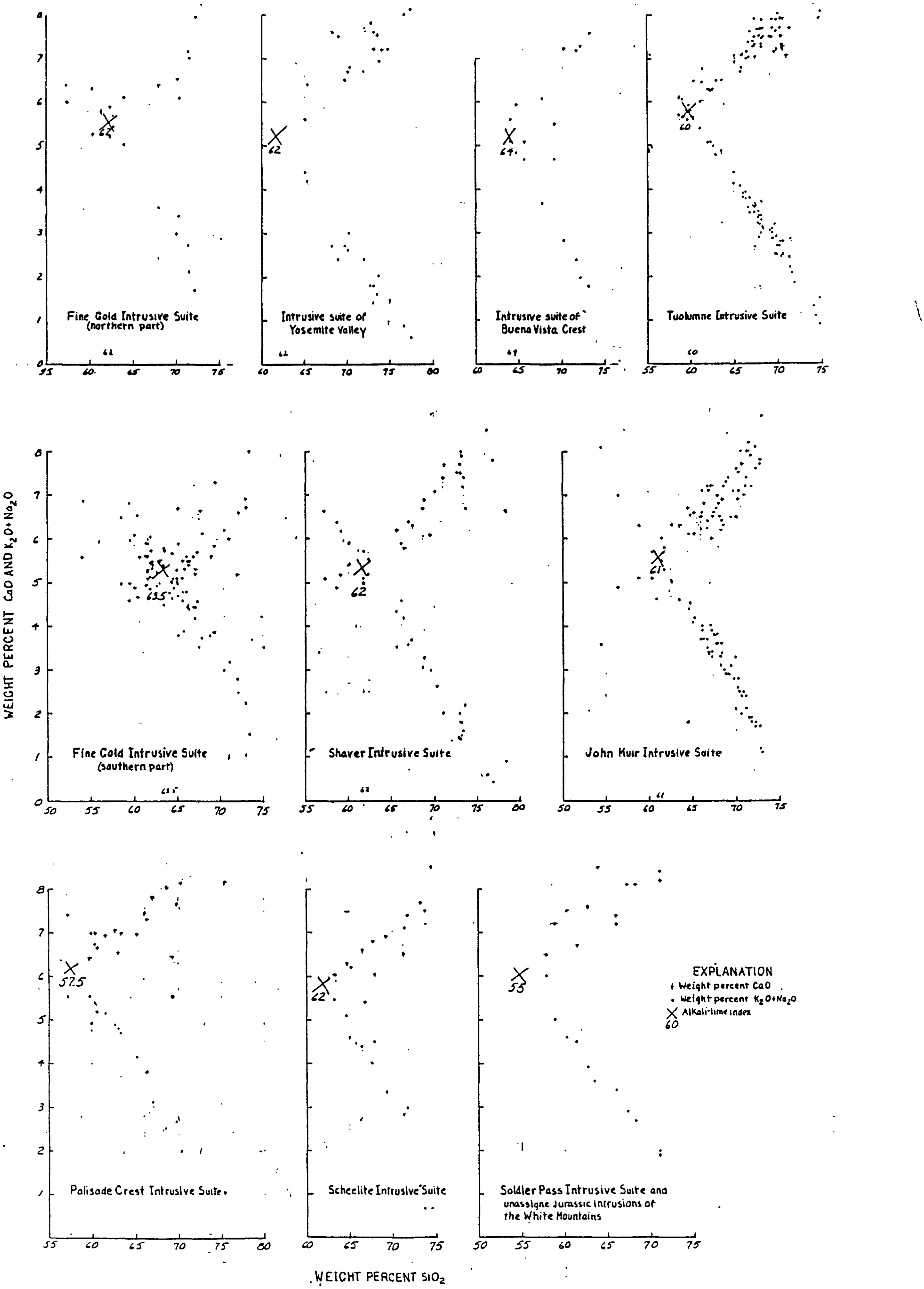


FIGURE 50.--Alkali-lime (Peacock) index for intrusive suites. Index is percent SiO_2 where weight percent CaO $\equiv \sum$ weight percent $K_2O + K_2O$.

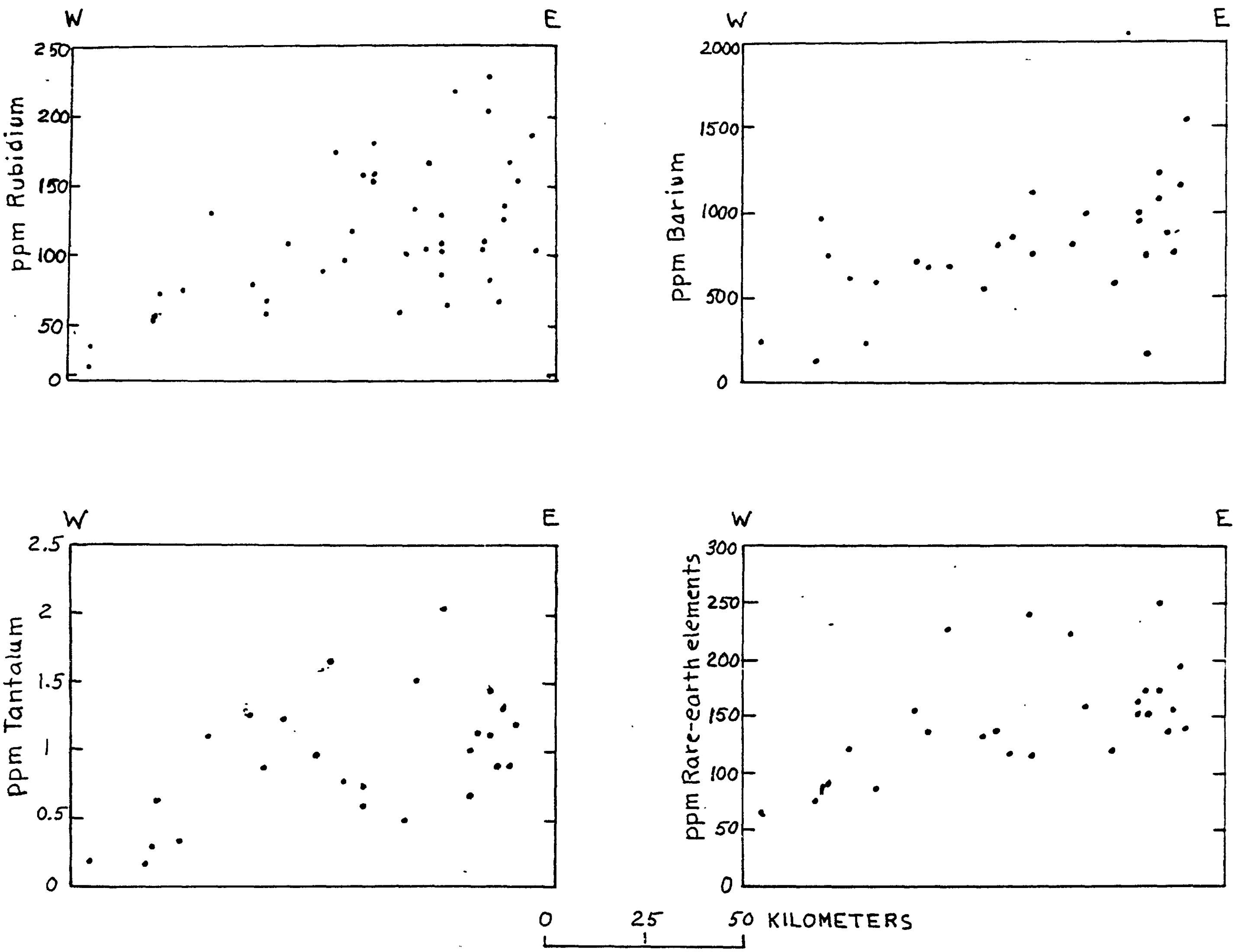


FIGURE 51.--Variation of selected elements east-west across Sierra Nevada. Rubidium from Dodge and others (1970) and barium, tantalum, and rare-earth elements from Dodge and others (1982).

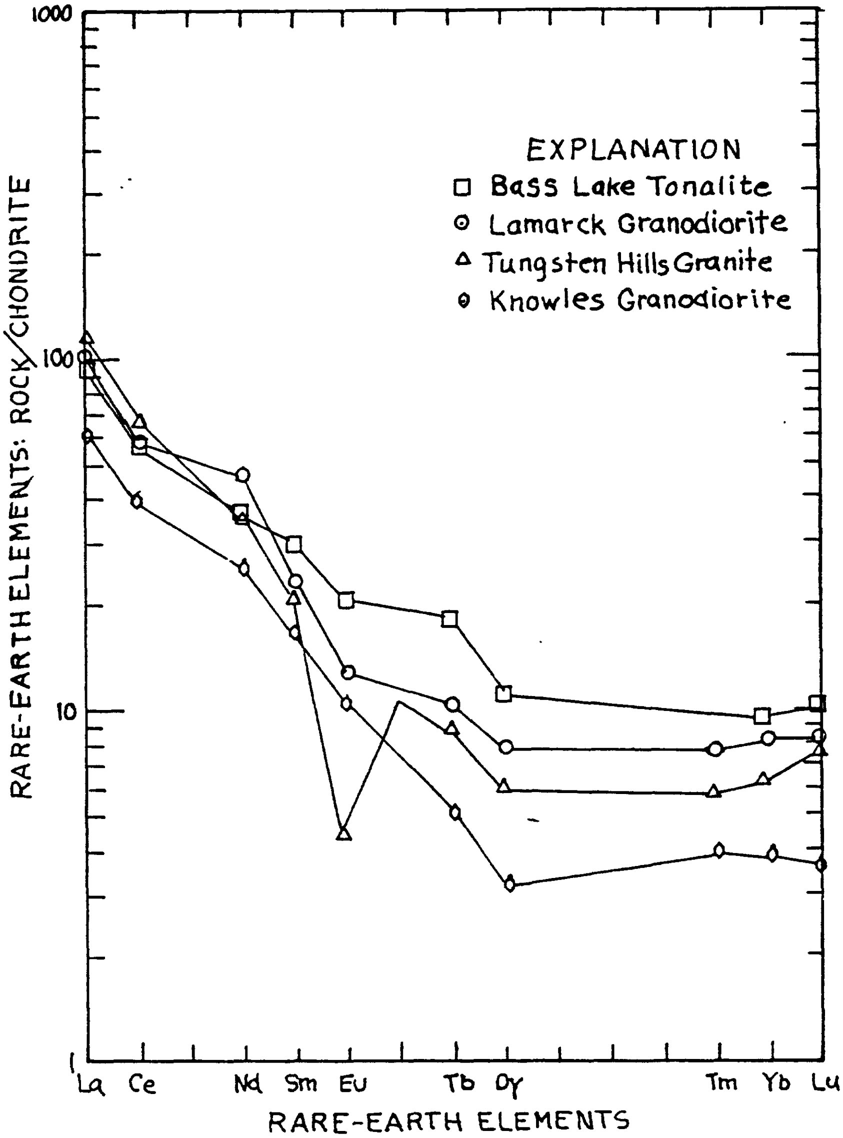


FIGURE 52.--Chondrite-normalized rare-earth element patterns for representative granitoids of central Sierra Nevada. From Dodge and others (1982).

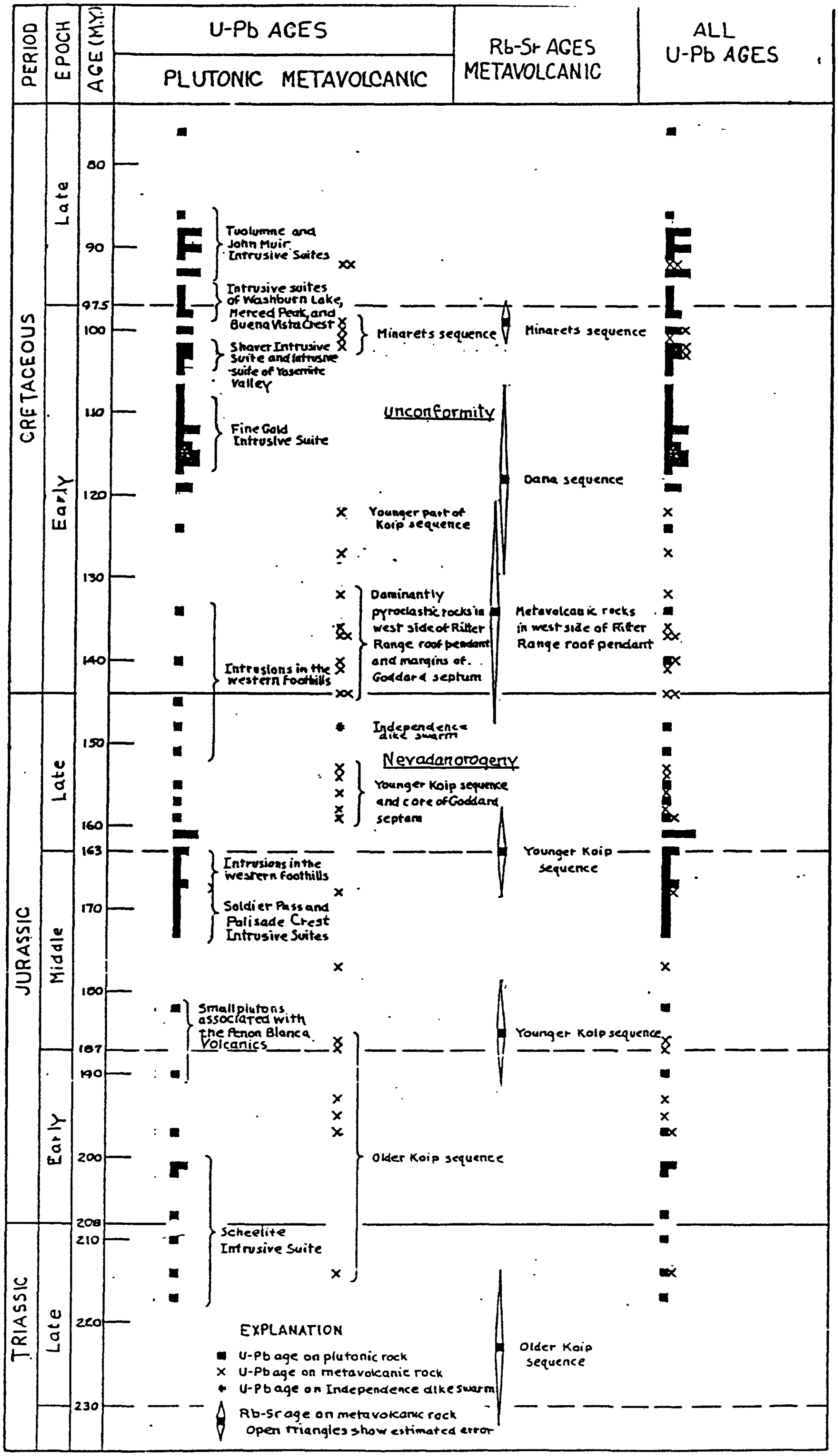


FIGURE 53.--Isotopic ages of volcanic and plutonic rocks within the Mariposa 1° by 2° quadrangle (pl. 1) plotted in accordance with Decade of North American Geology time scale (Palmer, 1983). Data from Sylvester and others (1978), Gillespie (1979), Stern and others (1981), Chen and Moore (1982), Kistler and Swanson (1981), Dodge and Calk (1986), Hanson and others (1987), and unpublished data of T.W. Stern, R.S. Fiske, and O.T. Tobisch (oral and written communications, 1981-1986) and R.W. Kistler (written communications, 1981-1986).

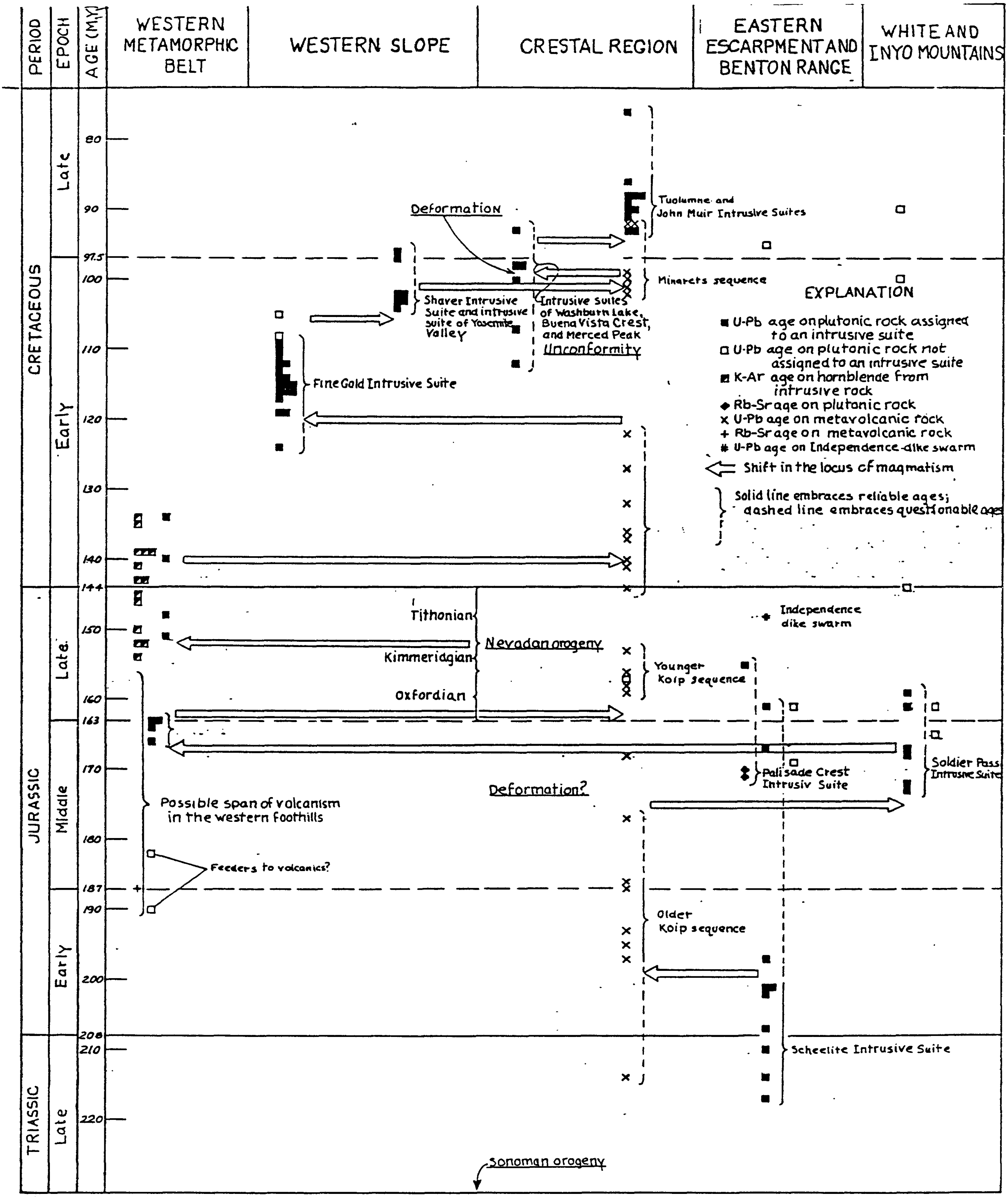


FIGURE 54.--Isotopic U-Pb ages plotted by zones across the Mariposa 1° by 2° quadrangle (pl. 1) to show loci of magmatism at different times. Data from Evernden and Kistler (1970), Sylvester and others (1978), Gillespie (1979), Stern and others (1981), Chen and Moore (1982), Dodge and Calk (1986), and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

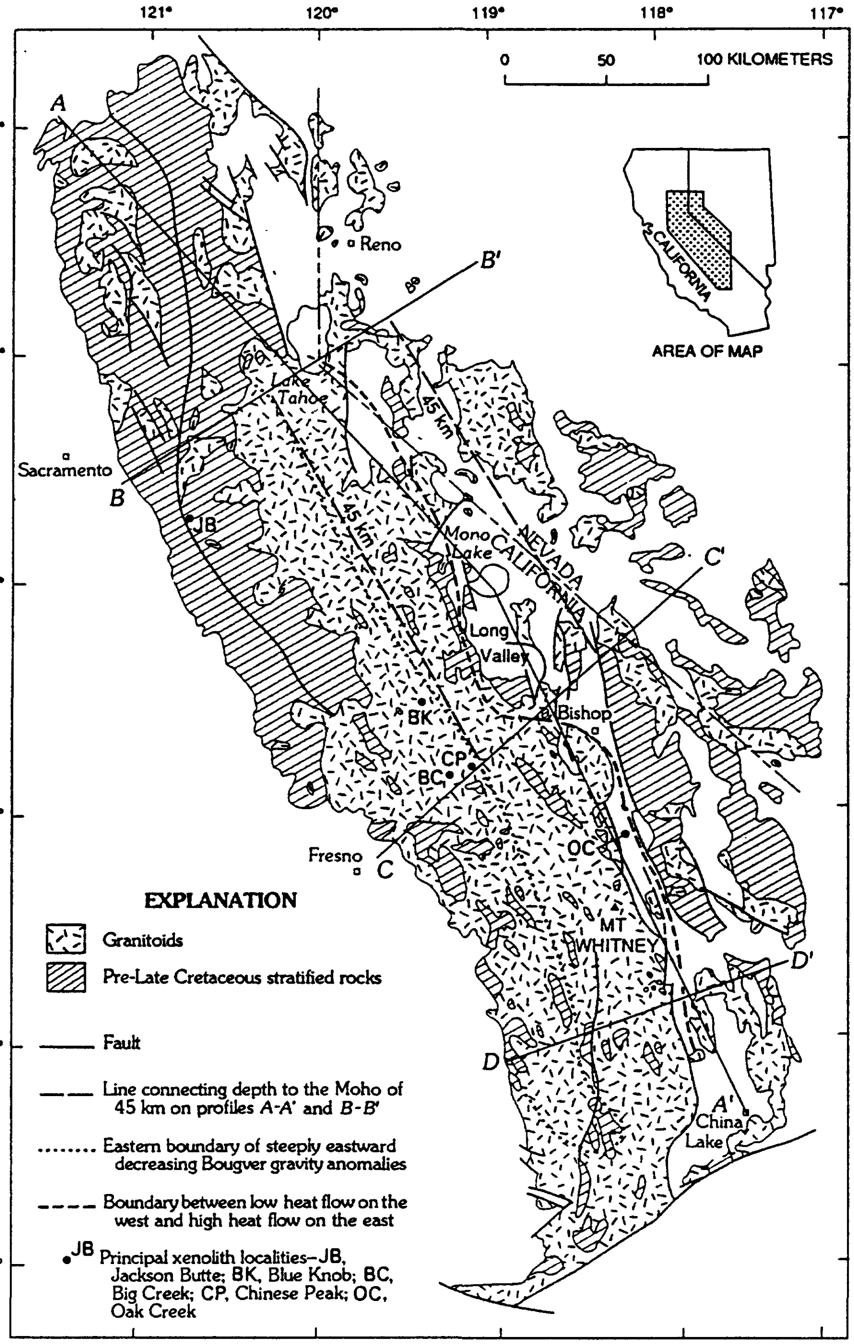


FIGURE 55.--Generalized geologic map of the Sierra Nevada showing locations of seismic-refraction sections A-A', B-B', and C-C' used in construction of figure 56, localities of xenoliths collected from Cenozoic volcanic rocks, and gravity anomaly and heat-flow boundaries. Seismic refraction profiles from Eaton (1966) and Bateman and Eaton (1967).

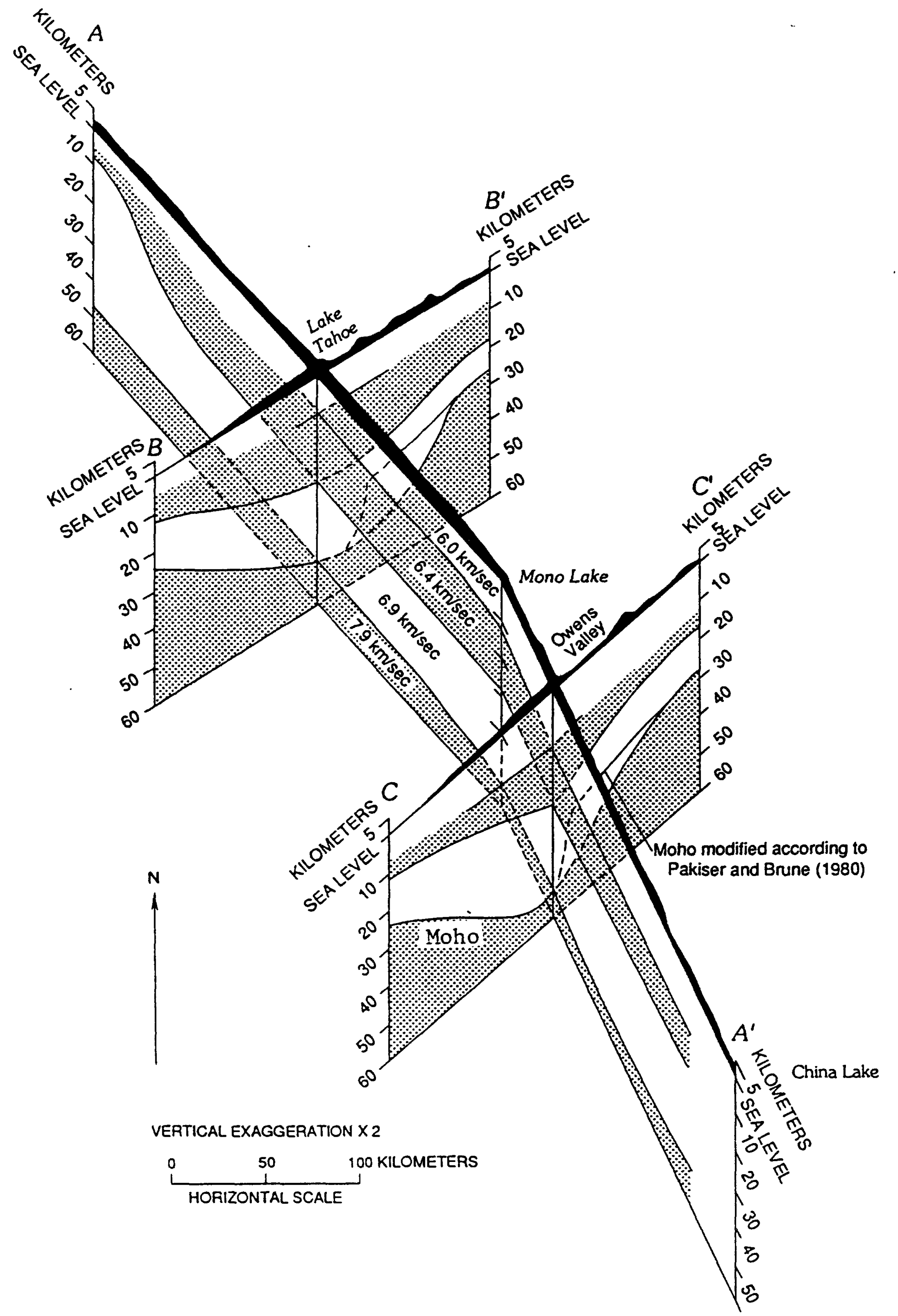


FIGURE 56.--Fence diagram showing seismic structure beneath the Sierra Nevada. Location of sections shown in figure 55.

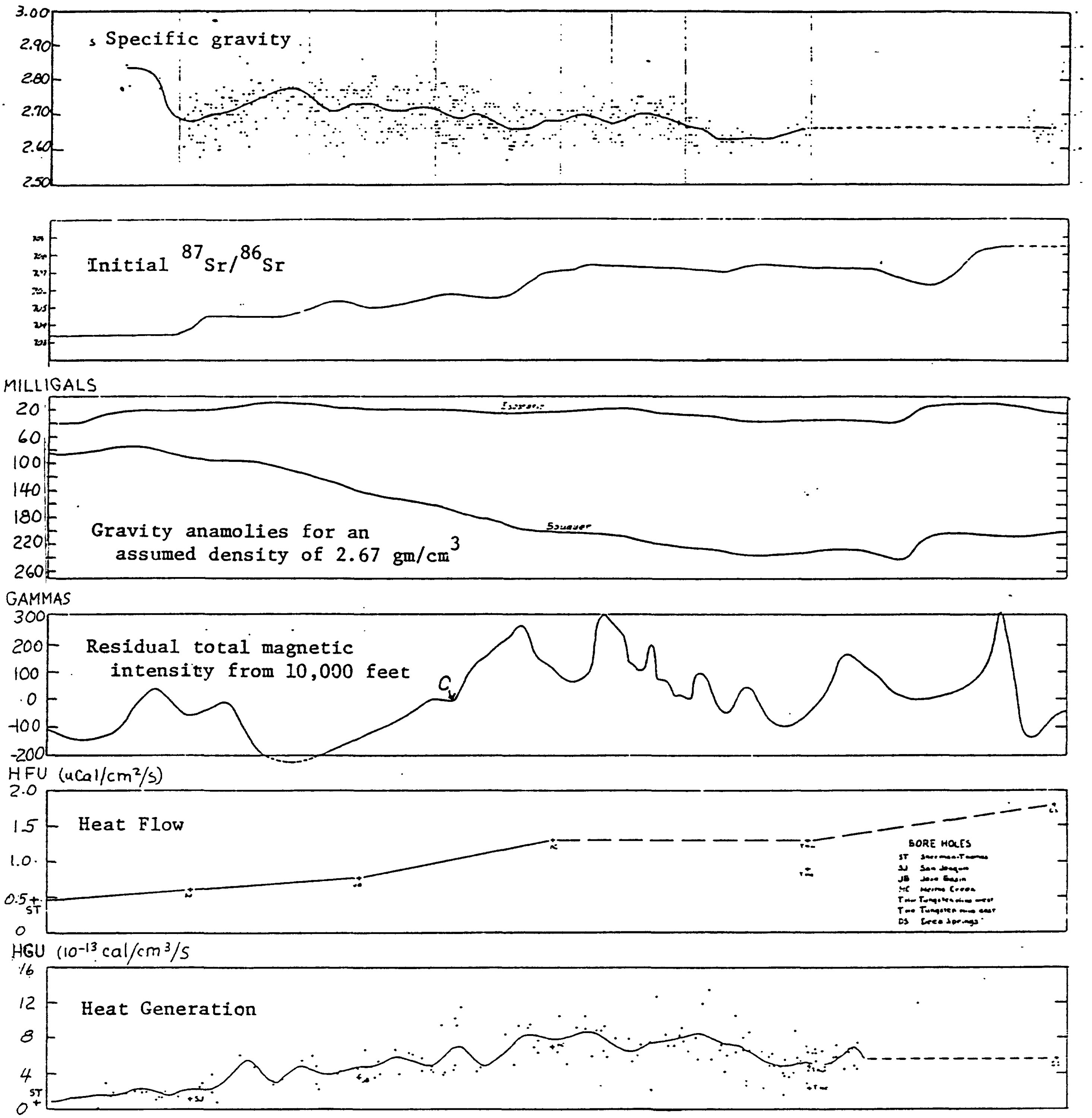


FIGURE 57.--Specific gravity, initial $^{87}Sr/^{86}Sr$, and geophysical profiles across the Sierra Nevada along line A-B, figure 32. Residual magnetic intensity profile is projected from a profile that trends N. 52° W. (Oliver, 1977) and crosses profile A-B near Red Mountain (C on the profile).

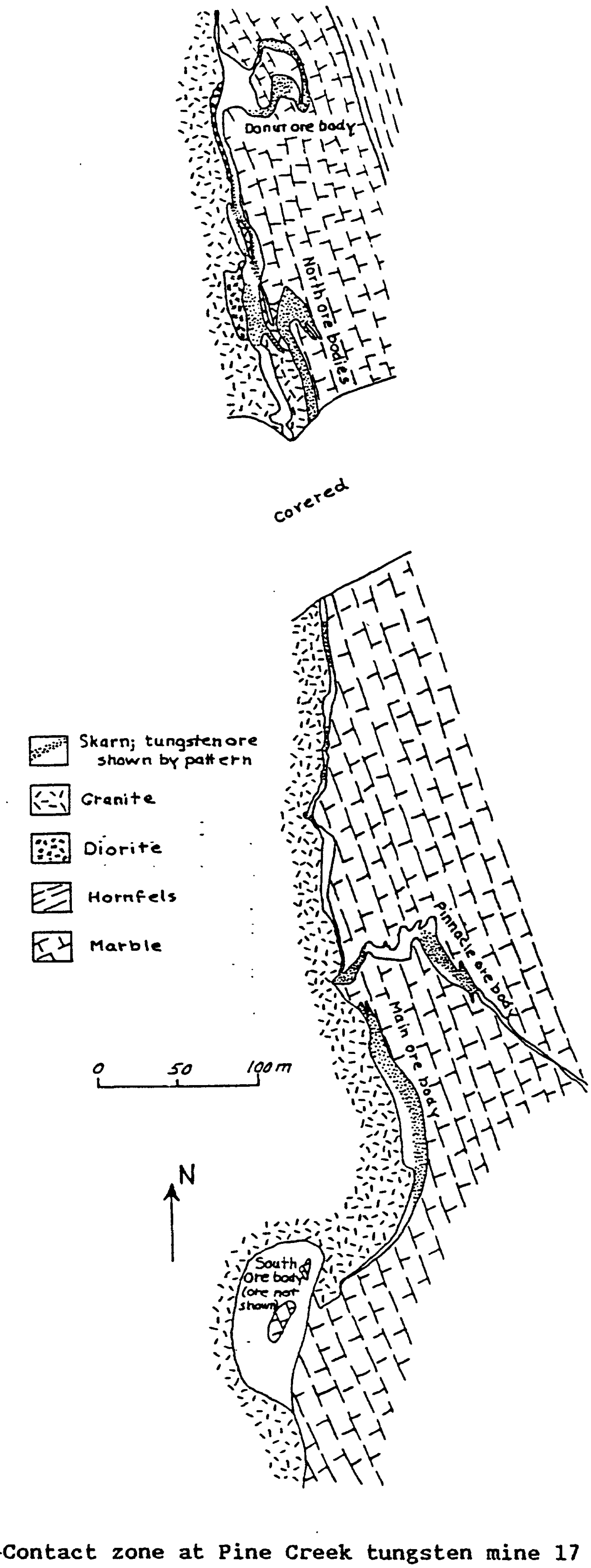


FIGURE 58.--Contact zone at Pine Creek tungsten mine 17 km west of Bishop on the north side of Pine Creek canyon. Modified from Lemmon (1941).

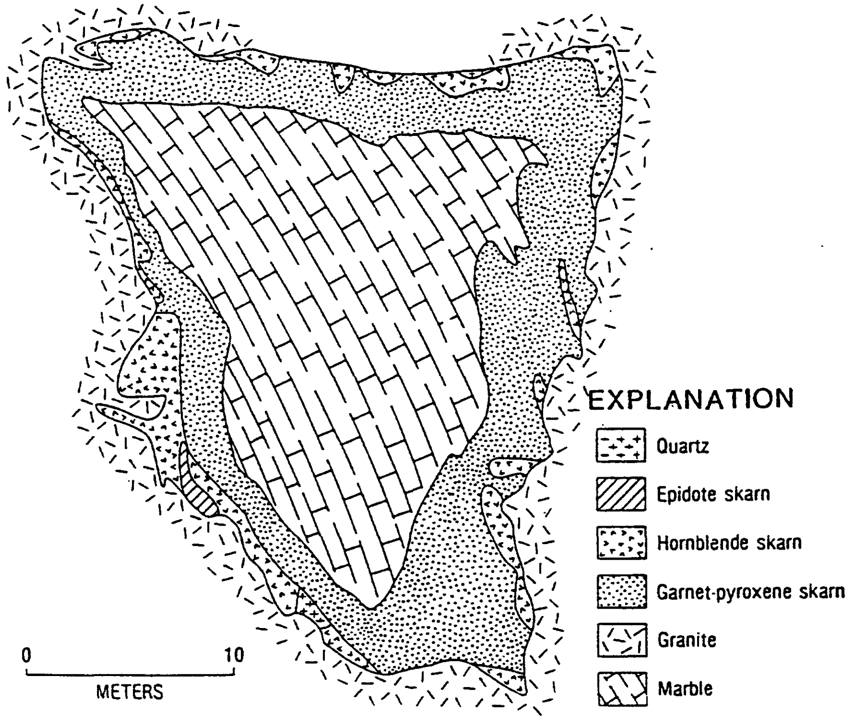


FIGURE 59.--Plan of Donut ore body, Pine Creek tungsten mine. See figure 58 for location. Modified from Gray and others (1968).

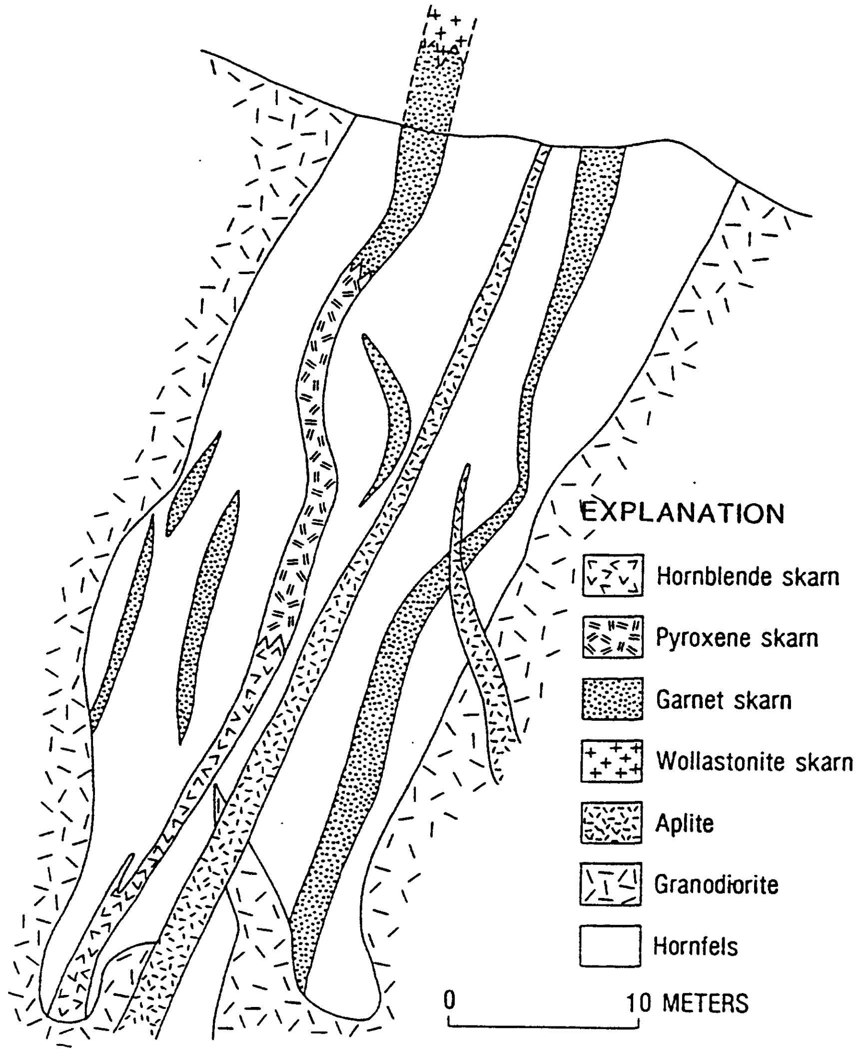


FIGURE 60.--Section across orebody no. 7 of Strawberry tungsten mine, about 90 km northest of Fresno, California near Clover Meadow in eastern Madera County. Modified from Nokleberg (1981).

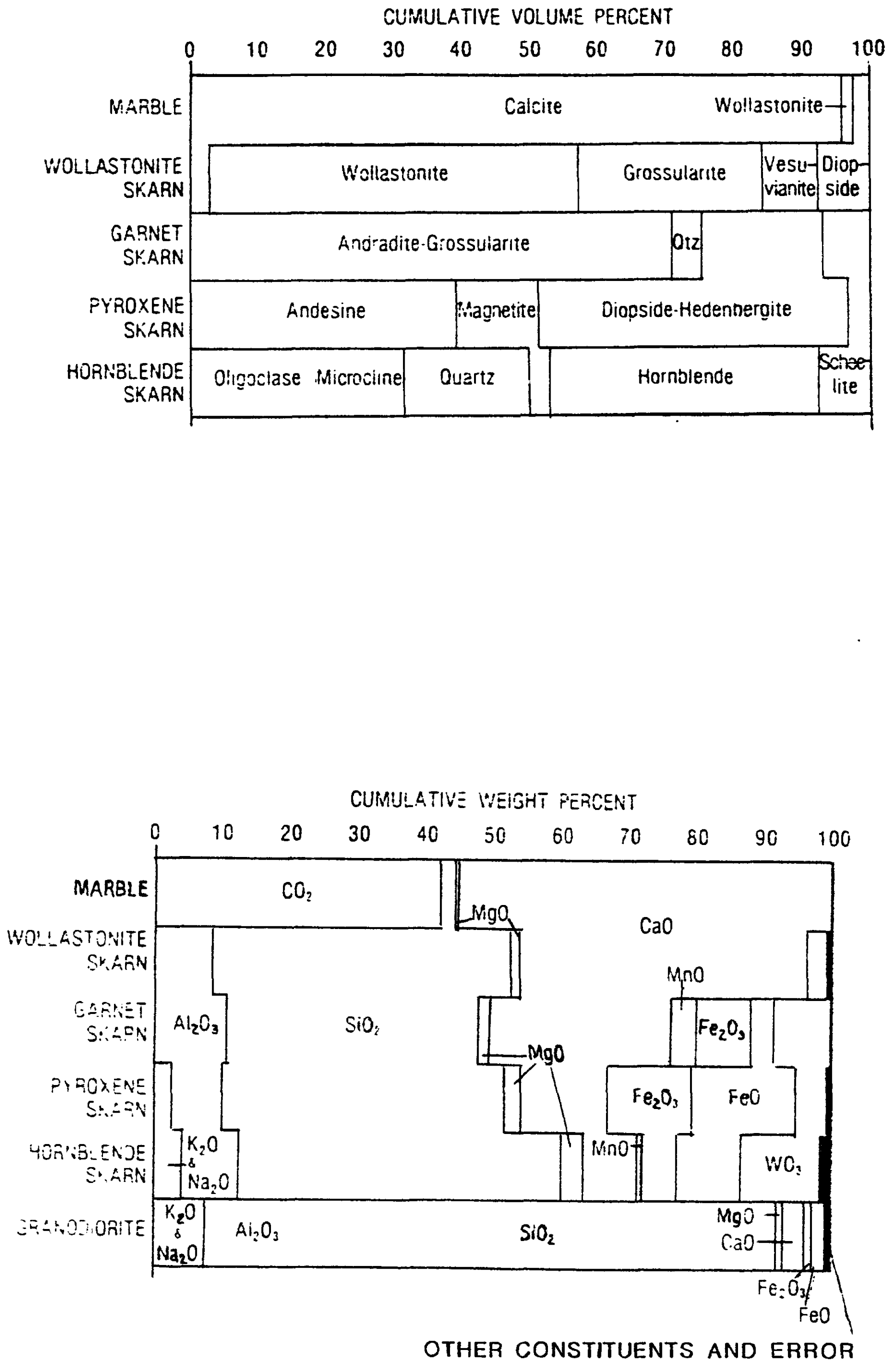


FIGURE 61.--Vertical modal and chemical changes in skarns of orebody 7 of Strawbery tungsten mine. 63A, modal changes. 63B, chemical changes. Based on data of Nokleberg (1981).

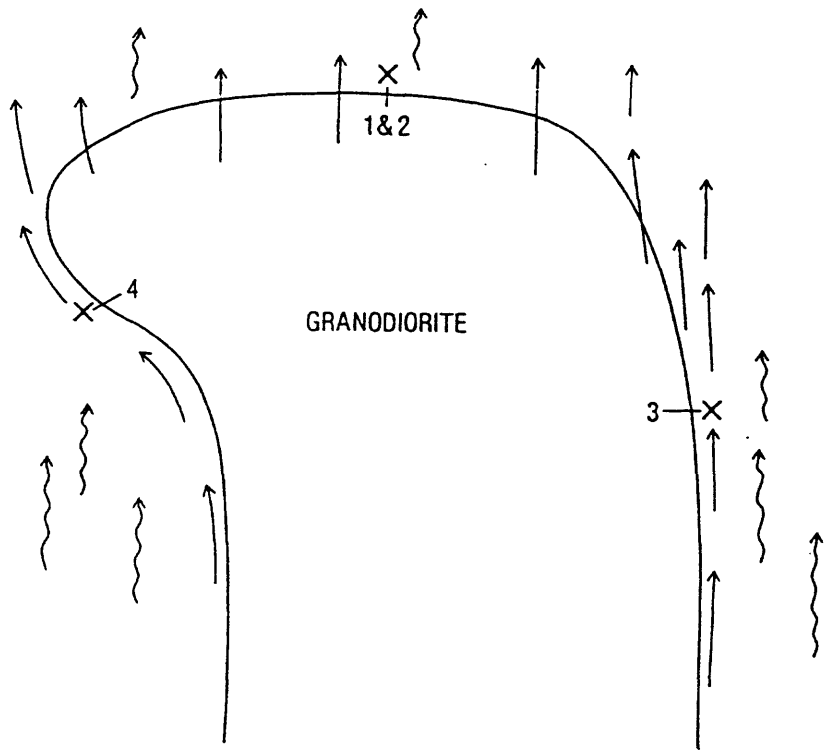


FIGURE 62.--Schematic section across Round Valley Peak Granodiorite showing inferred positions of skarn with reducing (1 and 2) and oxidizing trends (3 and 4). Arrows indicate flow of volatiles.

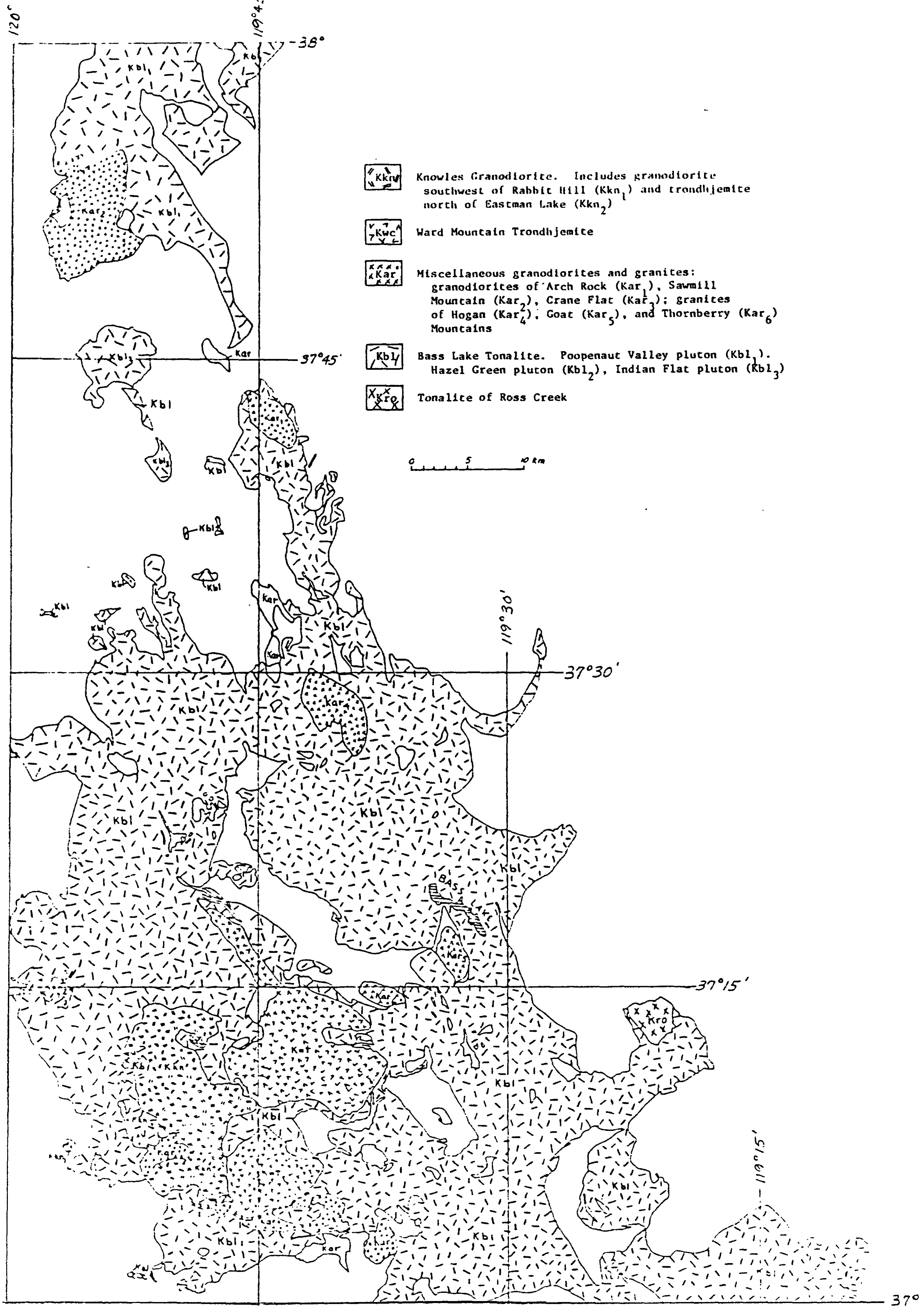


FIGURE 63.--Geologic index map of Fine Gold Intrusive Suite.

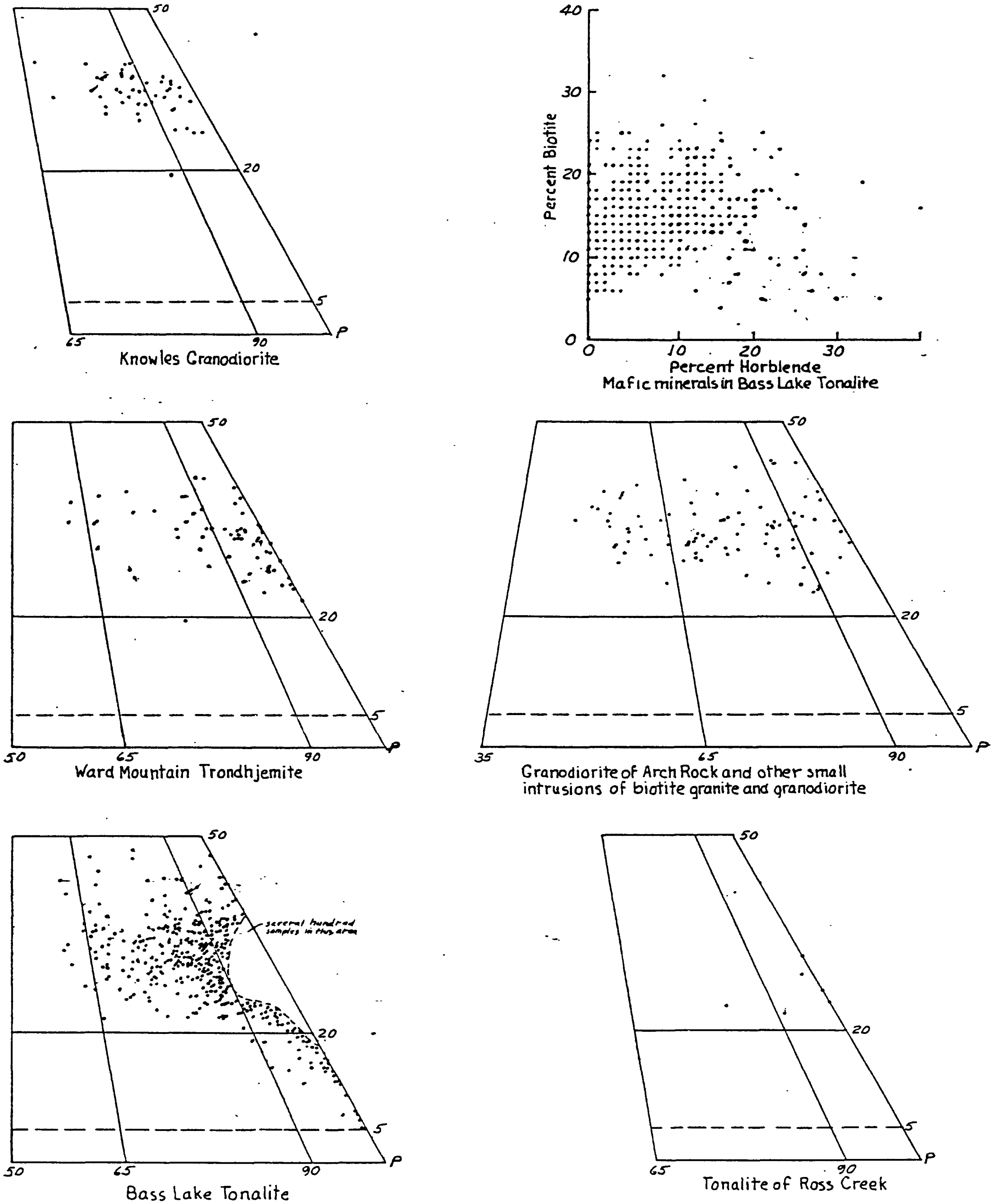


FIGURE 64.--Modal plots of Fine Gold Intrusive Suite. Includes plot showing amounts of biotite and hornblende in Bass Lake Tonalite.

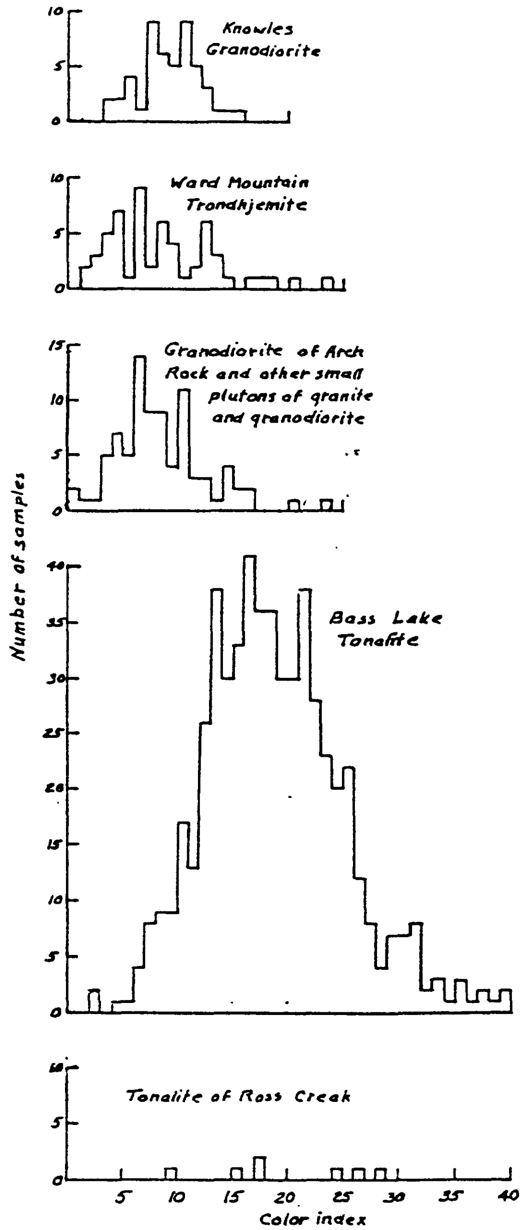


FIGURE 65.--Color index of Fine Gold Intrusive Suite.

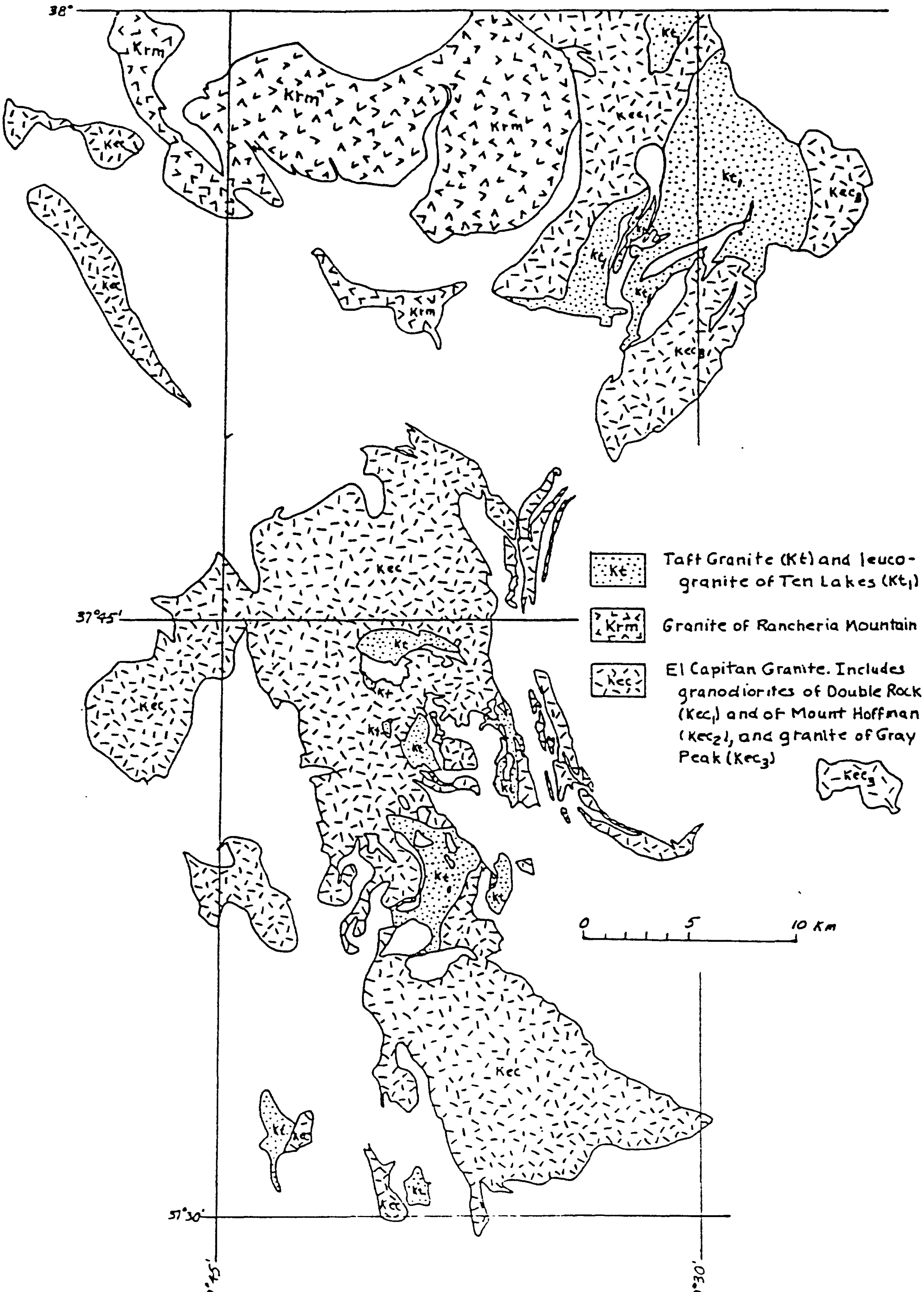


FIGURE 66.--Geologic index map of intrusive suite of Yosemite Valley.

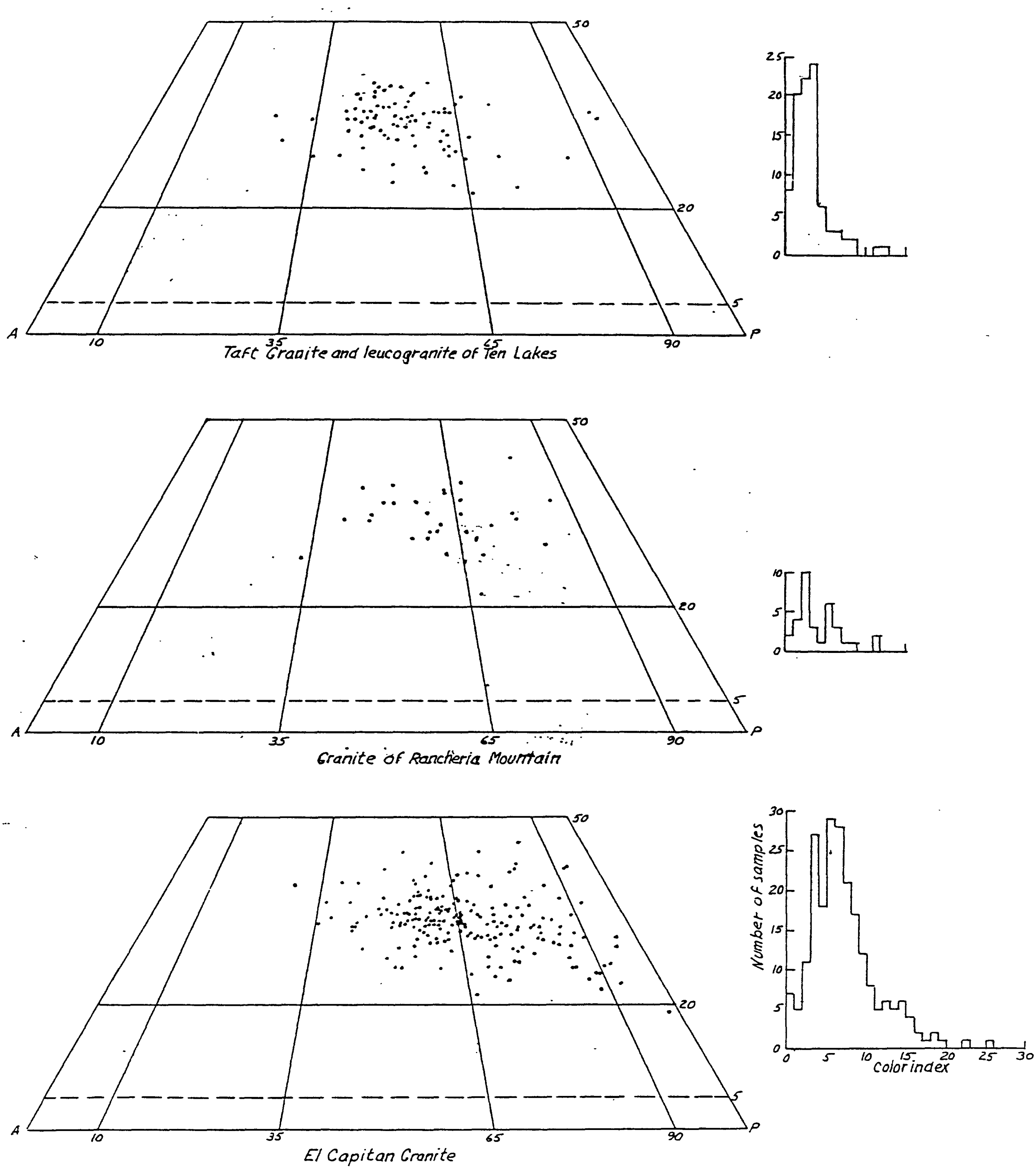


FIGURE 67.--Modes and color index of intrusive suite of Yosemite Valley.

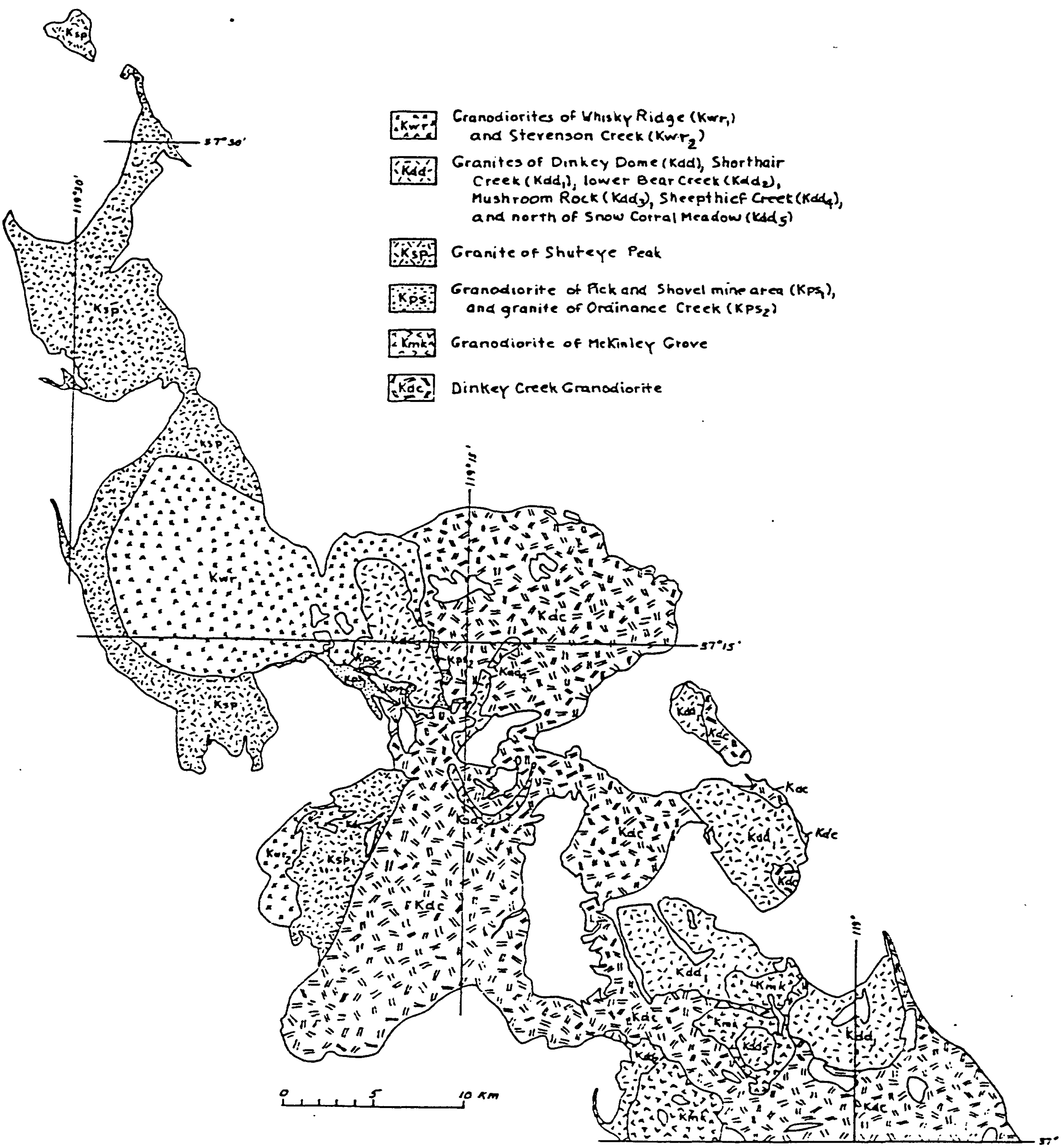


FIGURE 68.--Geologic index map of Shaver Intrusive Suite.

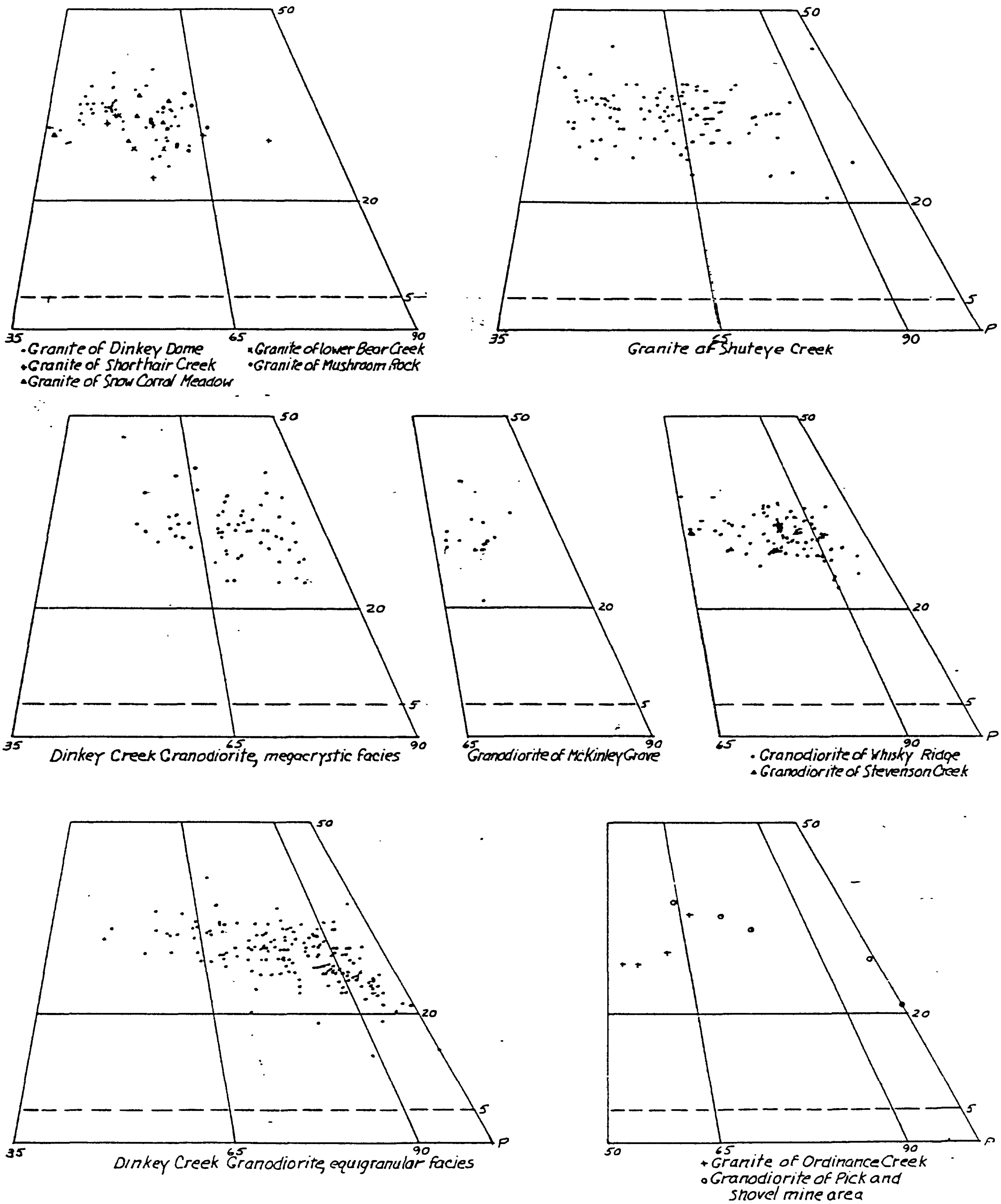


FIGURE 69.--Modes of Shaver Intrusive Suite.

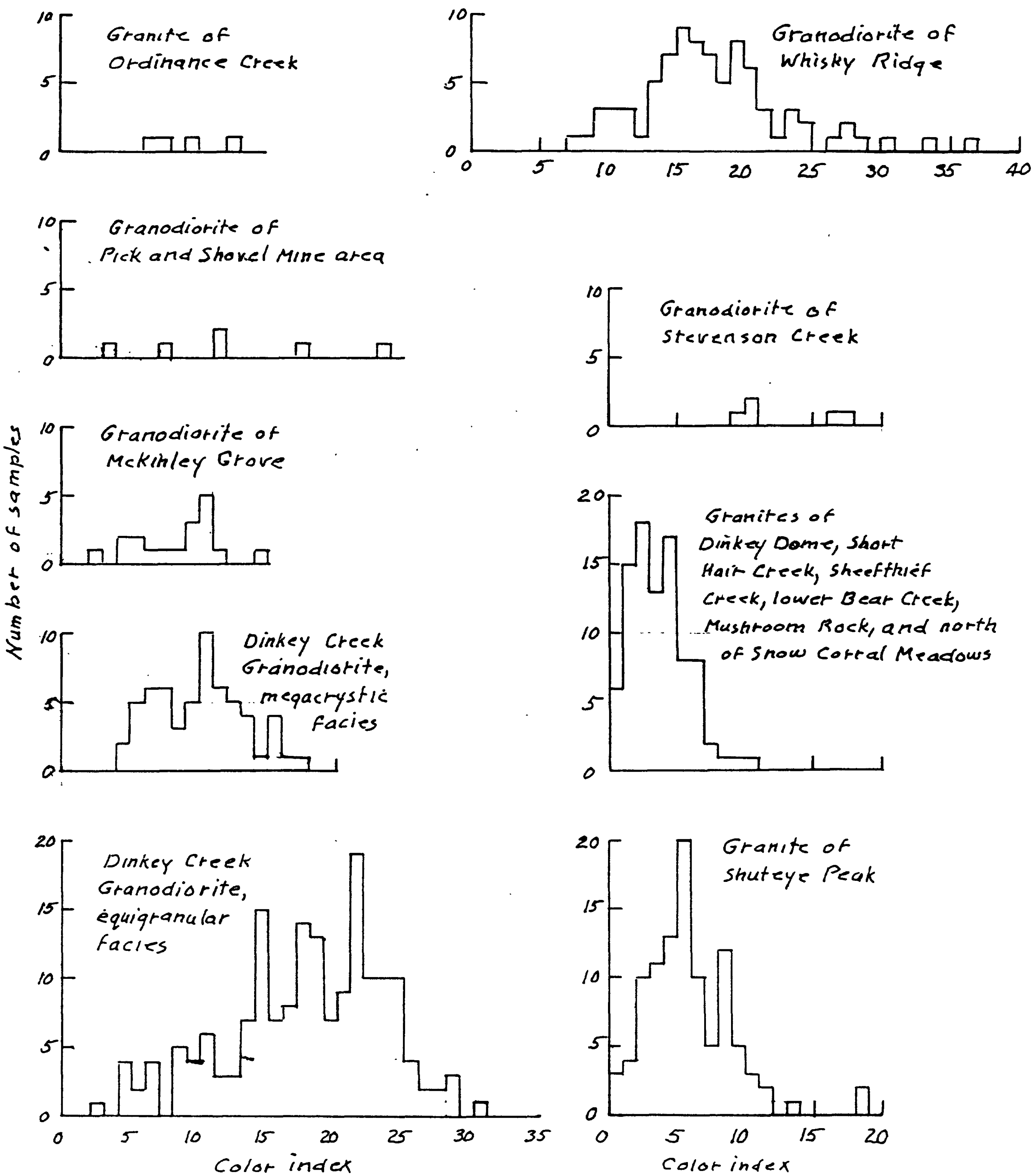


FIGURE 70.--Color index of Shaver Intrusive Suite.

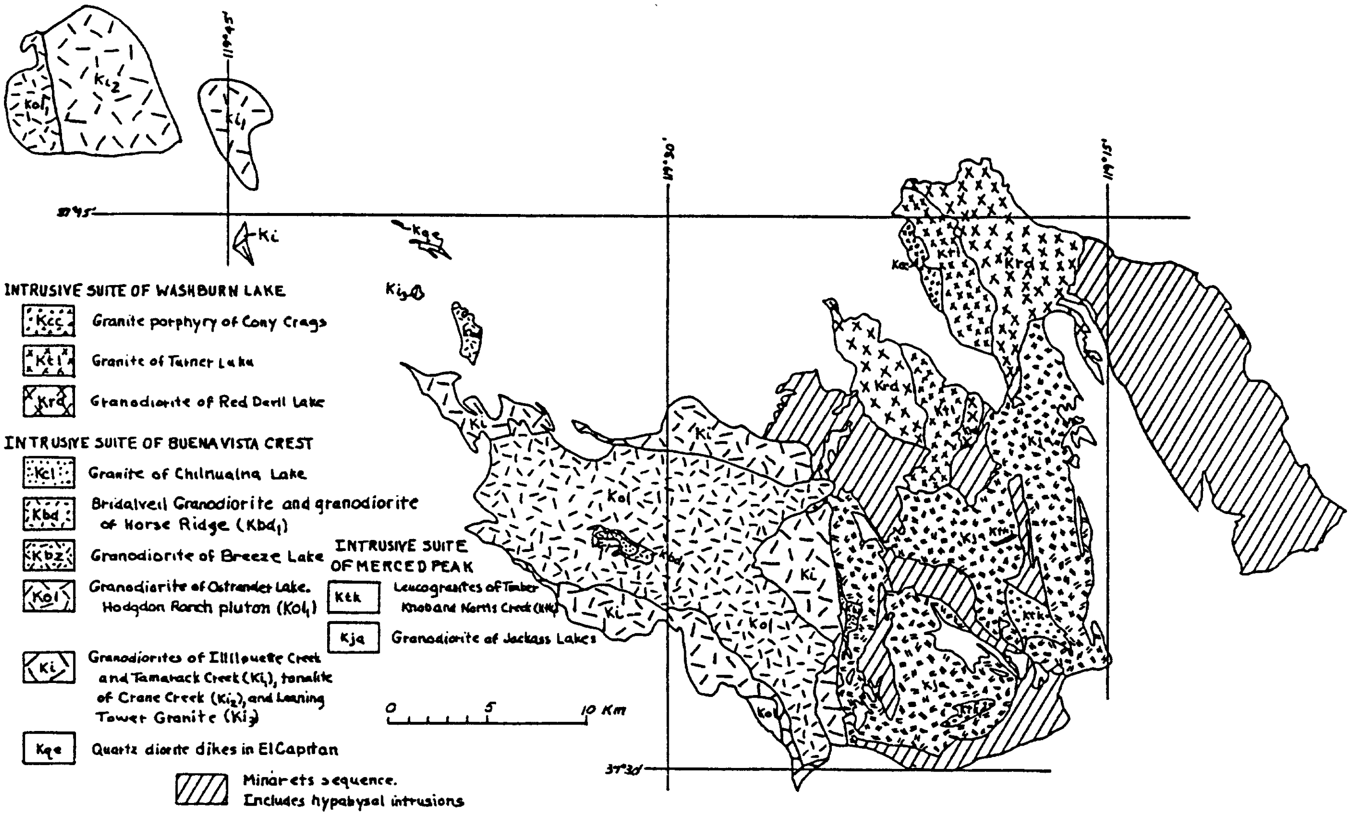


FIGURE 71.--Geologic index map of the intrusive suites of Washburn Lake, Merced Peak, and Buena Vista Crest.

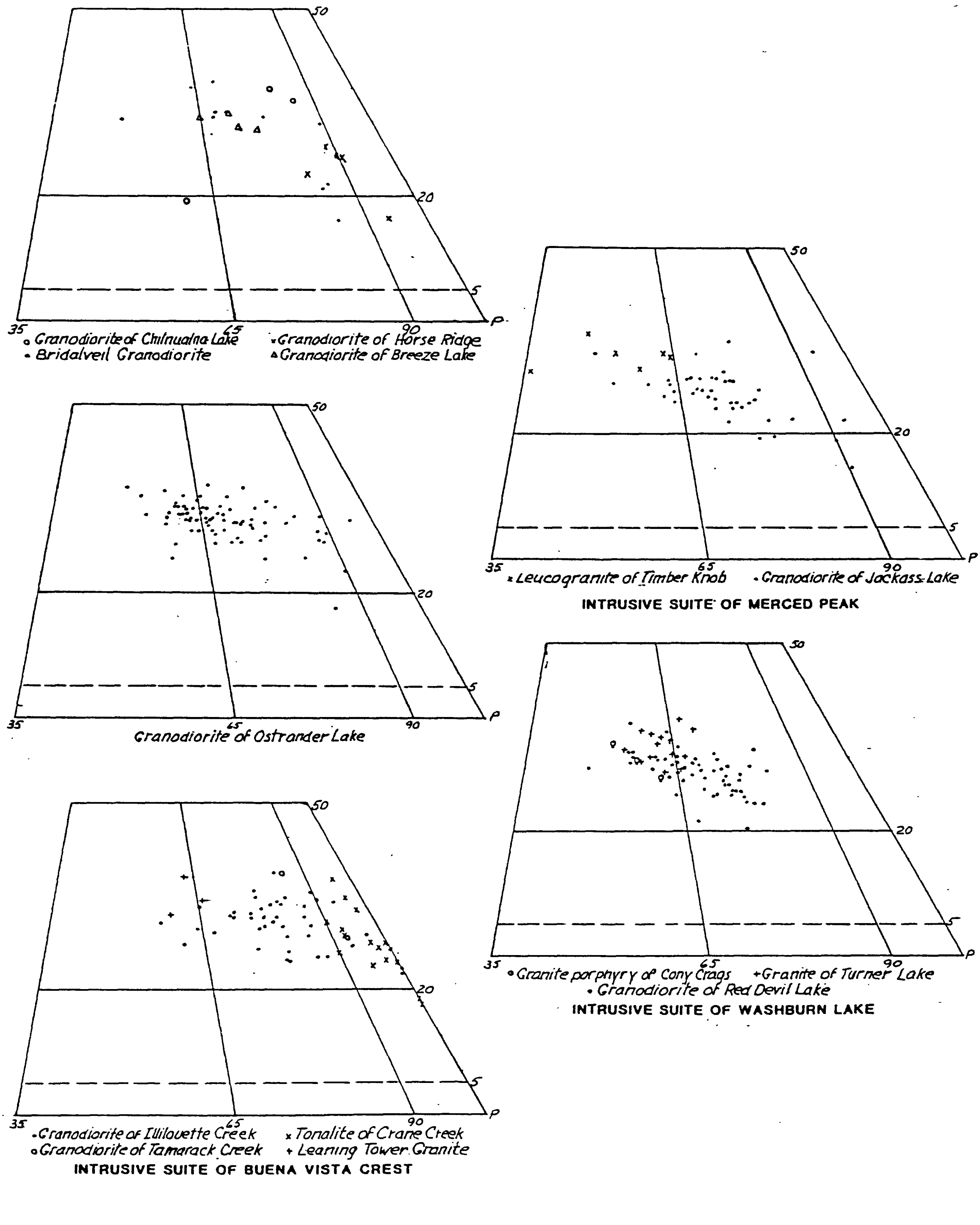
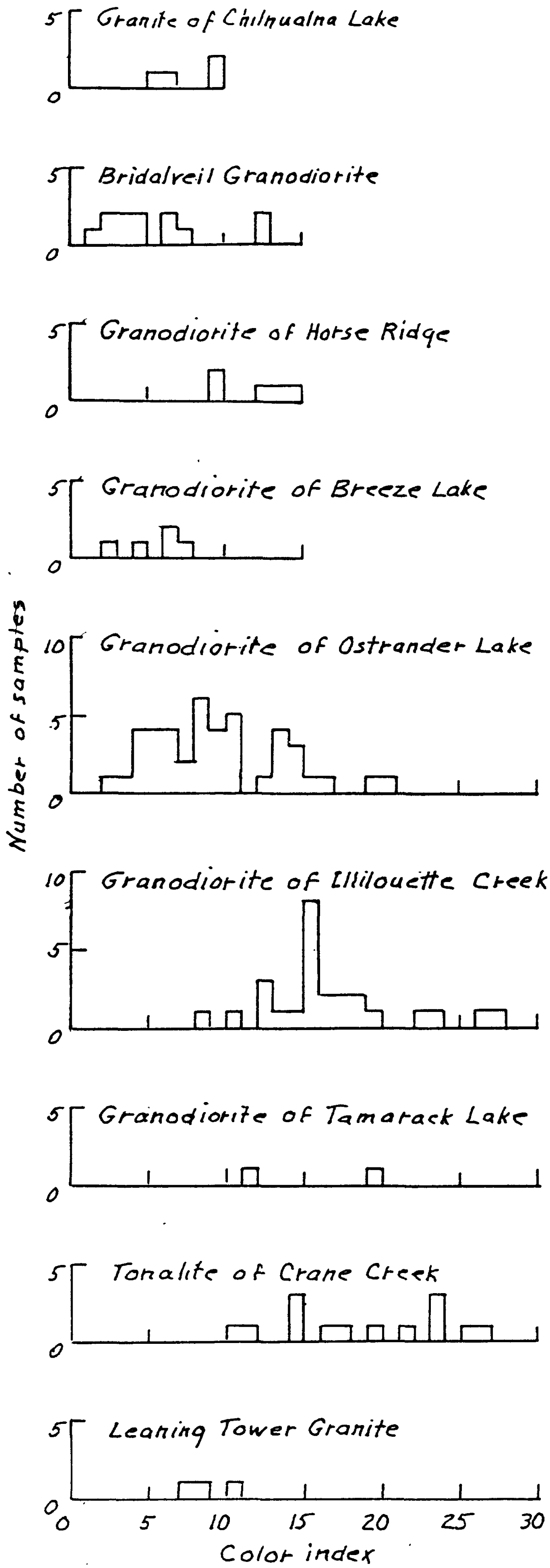


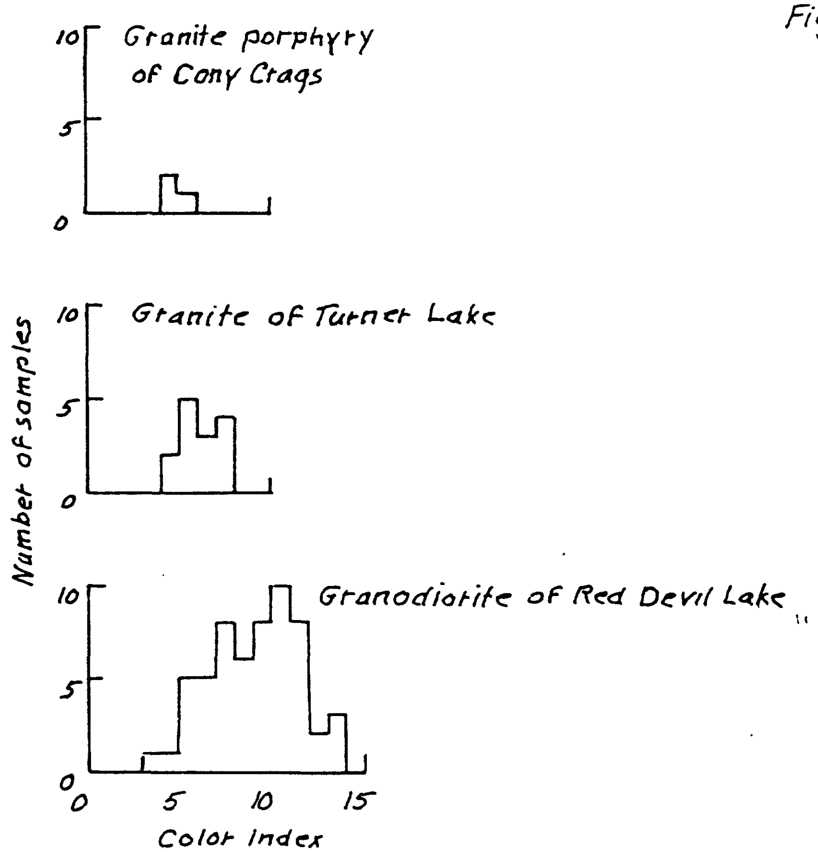
FIGURE 72.--Modes of intrusive suites of Washburn Lake, Merced Peak, and Buena Vista Crest.



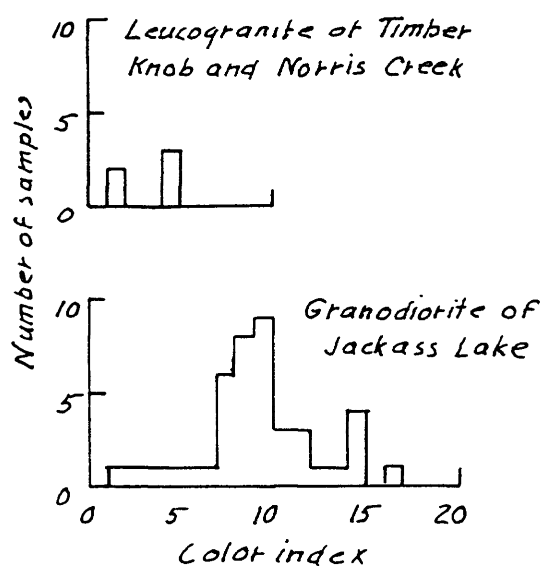
INTRUSIVE SUITE OF BUENA VISTA CREST

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

Fig. 73 con.



INTRUSIVE SUITE OF WASHBURN LAKE



INTRUSIVE SUITE OF MERCED PEAK

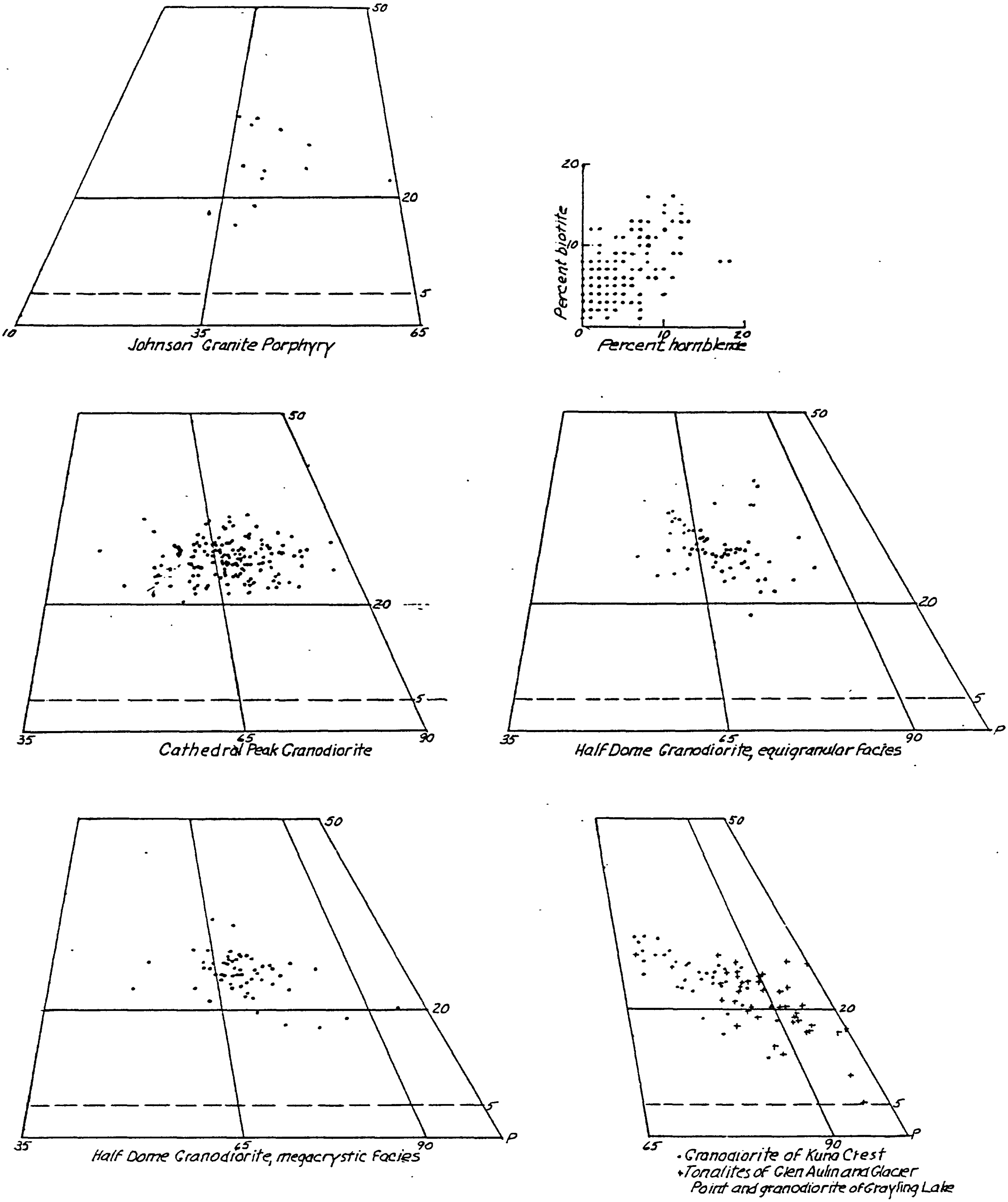


FIGURE 74.--Modes of Tuolumne Intrusive Suite. Includes plot showing amounts of hornblende and biotite in suite.

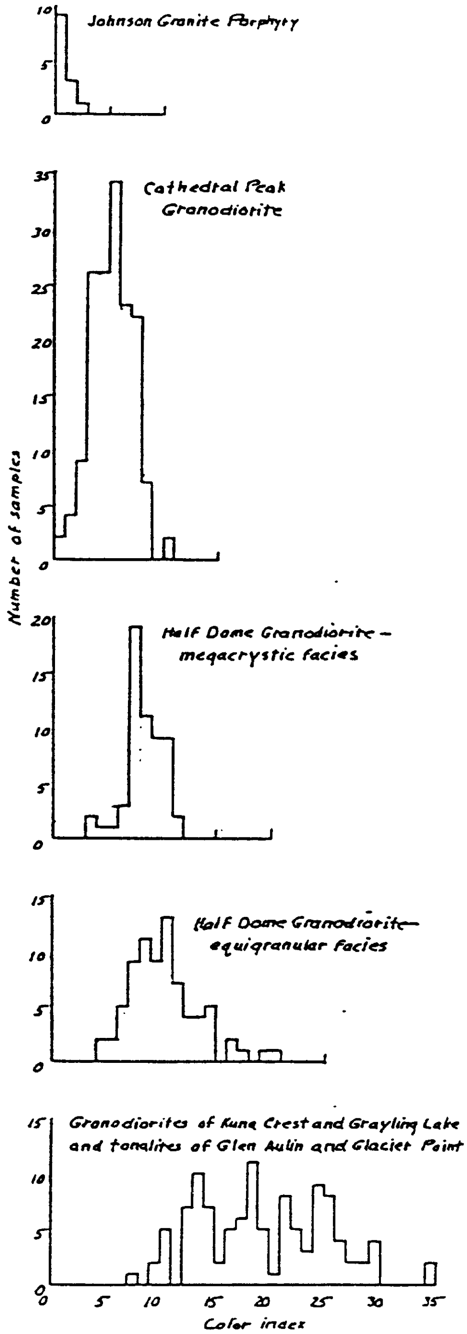


FIGURE 75.--Color index of Tuolumne Intrusive Suite.

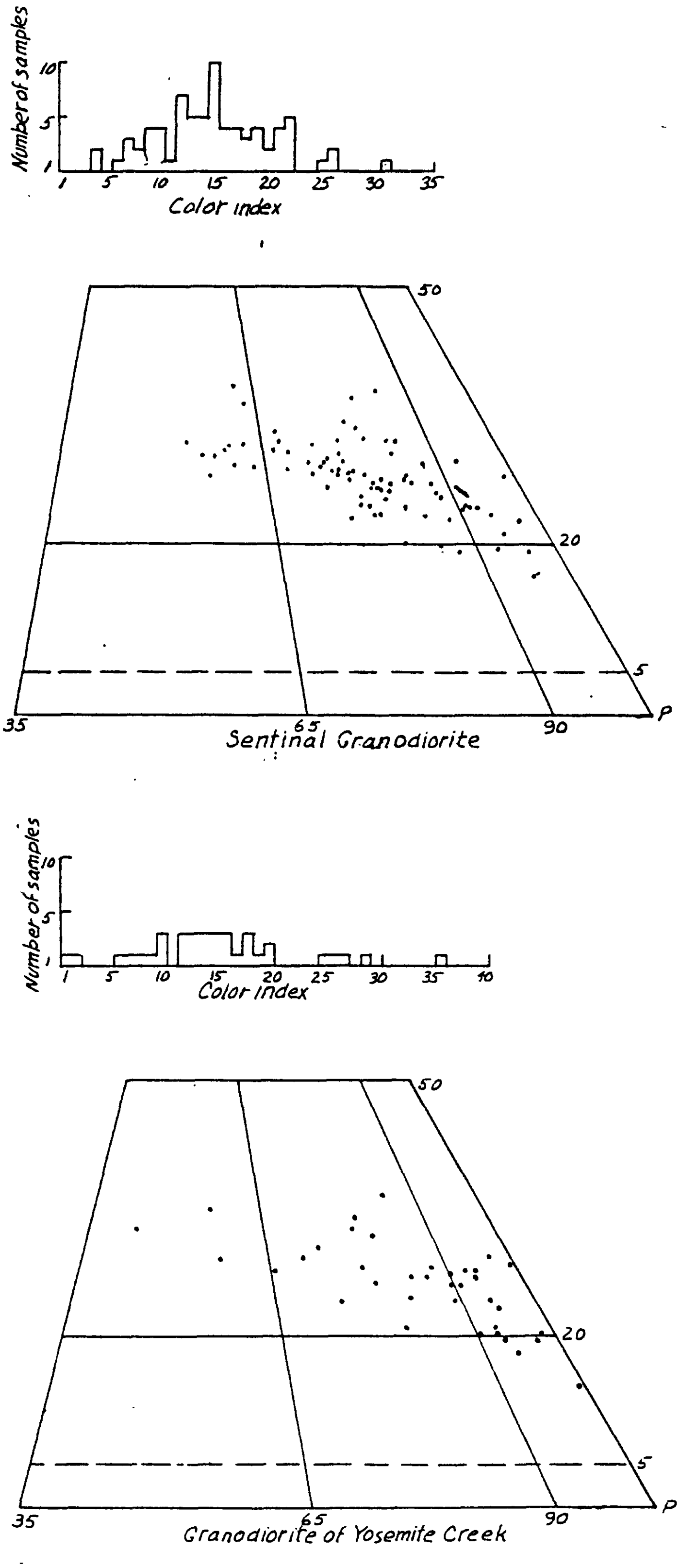


FIGURE 76.--Modes and color index of Sentinel Granodiorite and granodiorite of Yosemite Creek.

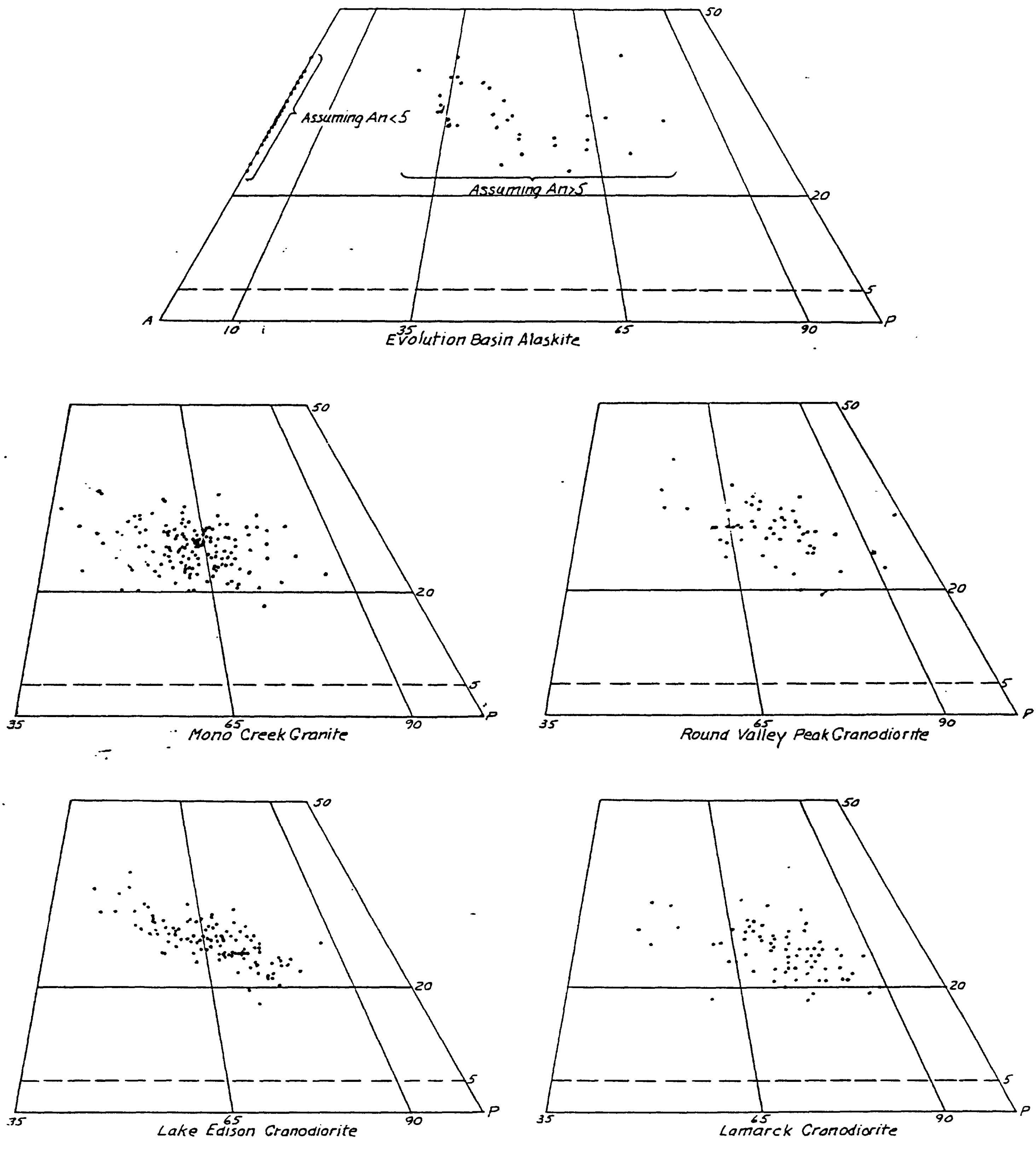


FIGURE 77.--Modes of intrusions of John Muir Intrusive Suite east of Mount Givens Granodiorite.

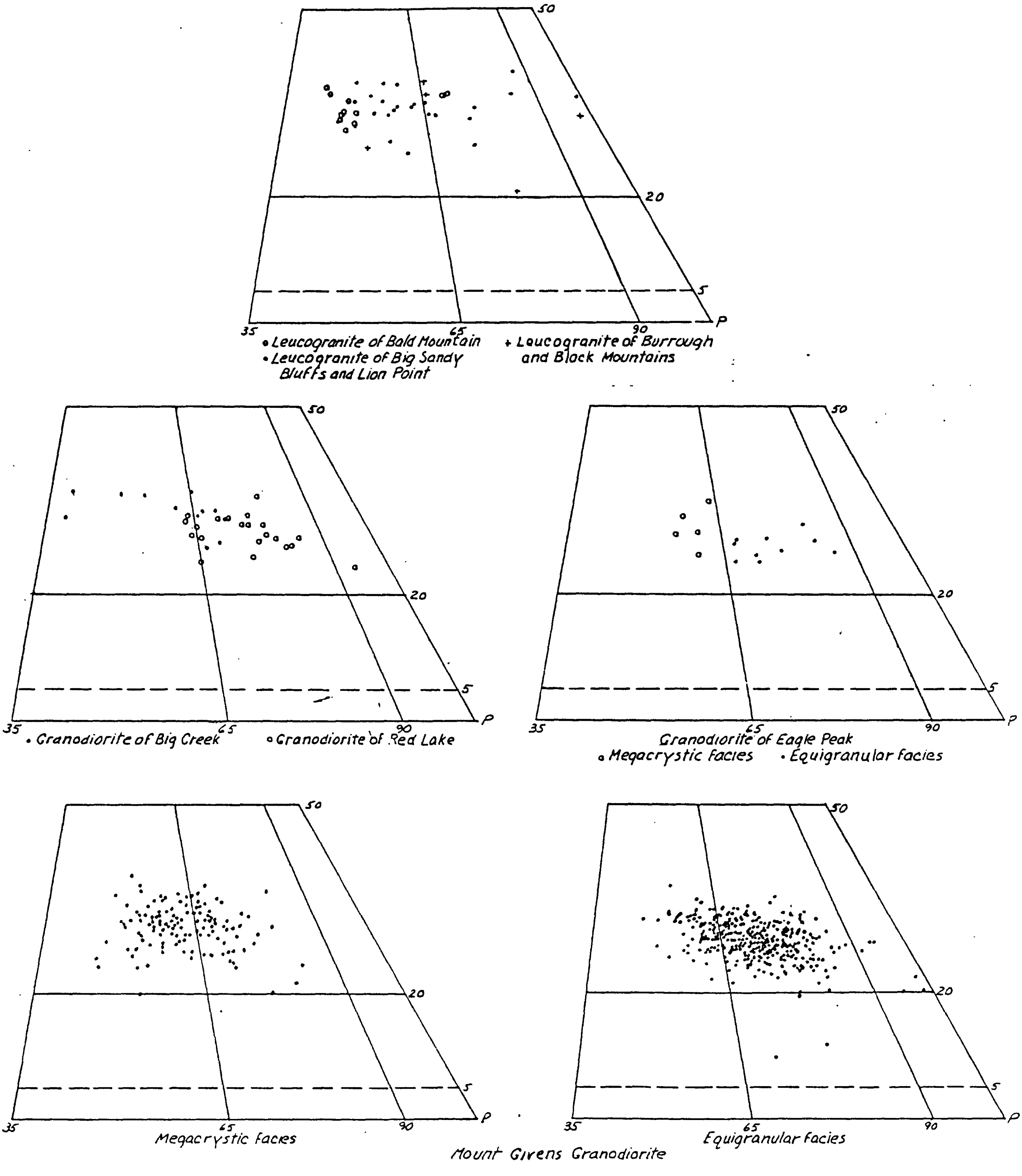


FIGURE 78.--Modes of Mount Givens Granodiorite and small intrusions of the John Muir Intrusive Suite west of the Mount Givens Granodiorite.

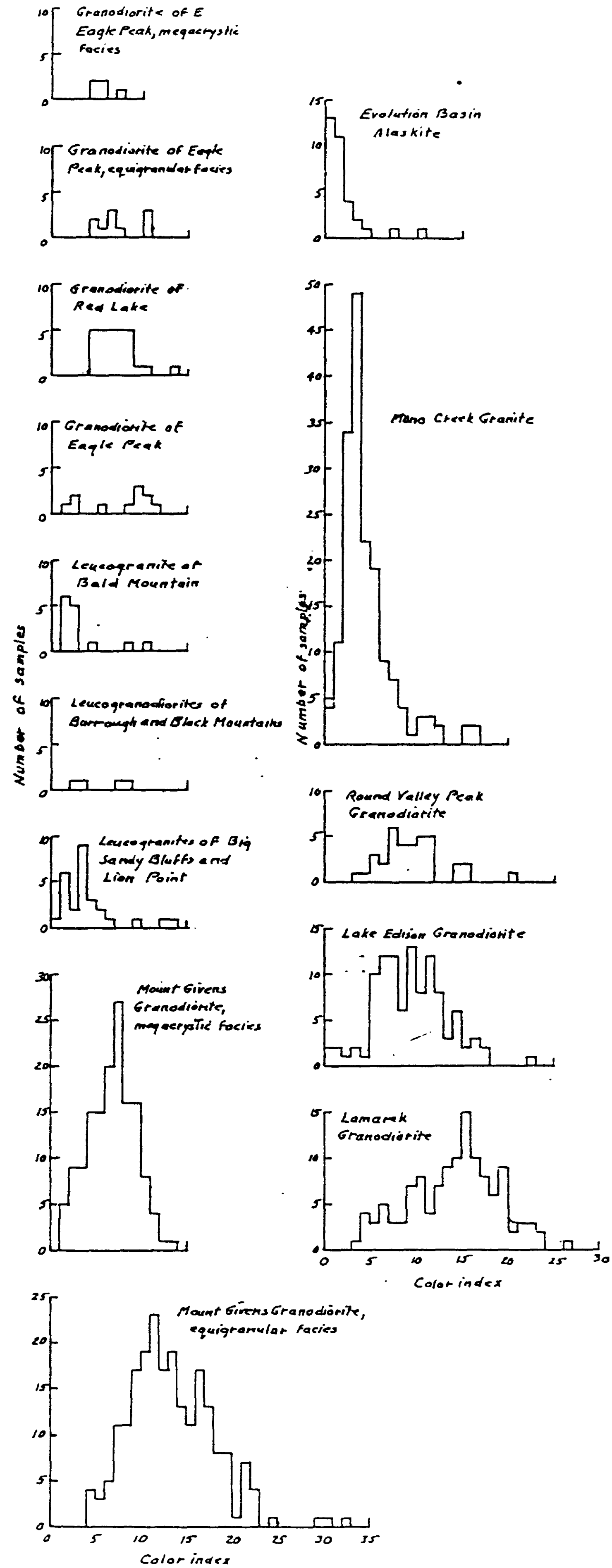


FIGURE 79--Color Index of John Muir Intrusive Suite.

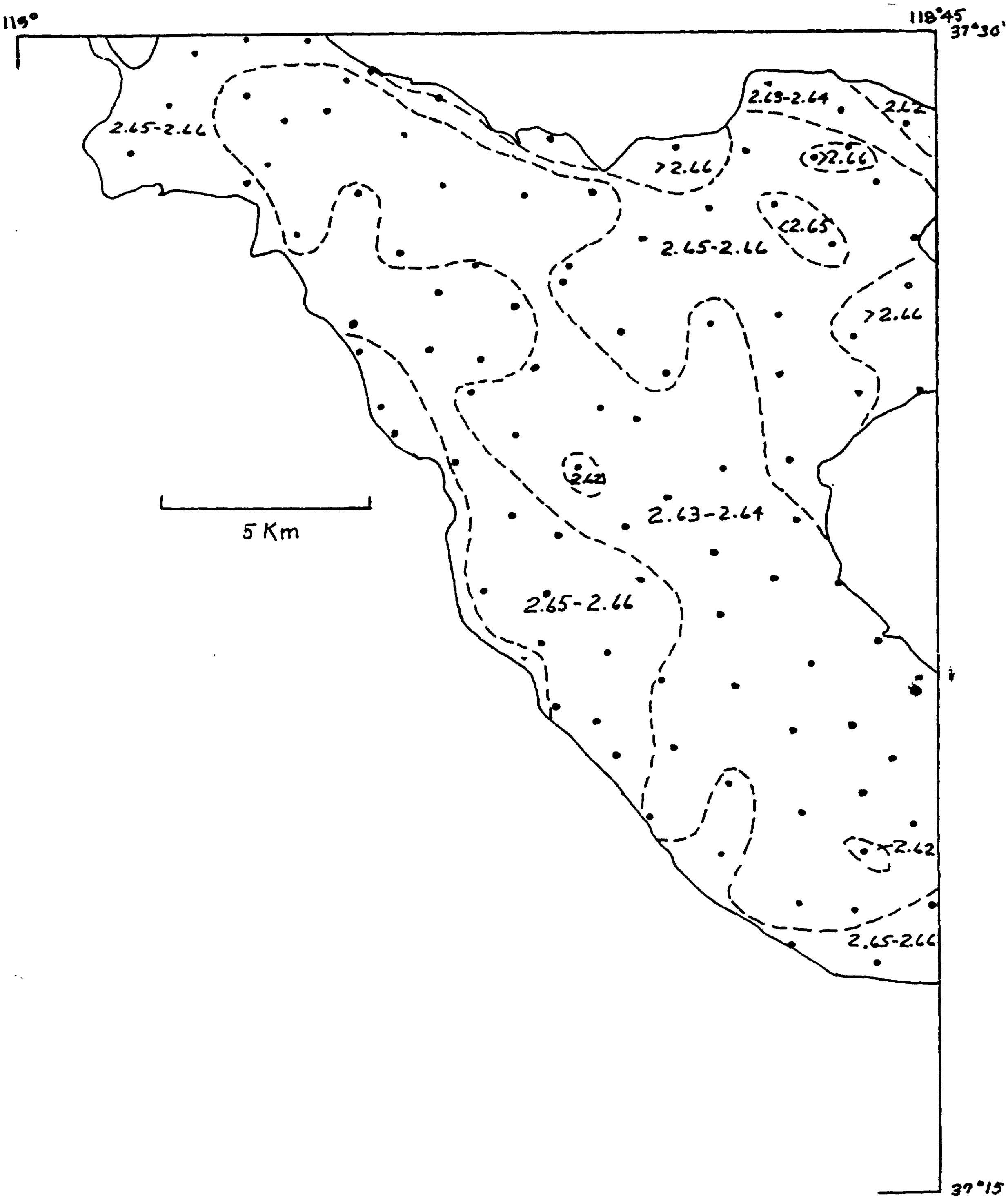


FIGURE 80.--Isopleth map showing variations of specific gravity within central and southern part of Mono Creek Granite. Modified from Lockwood (1975). Dots show sample localities. Dashed lines are isopleths and separate areas with different specific gravity.

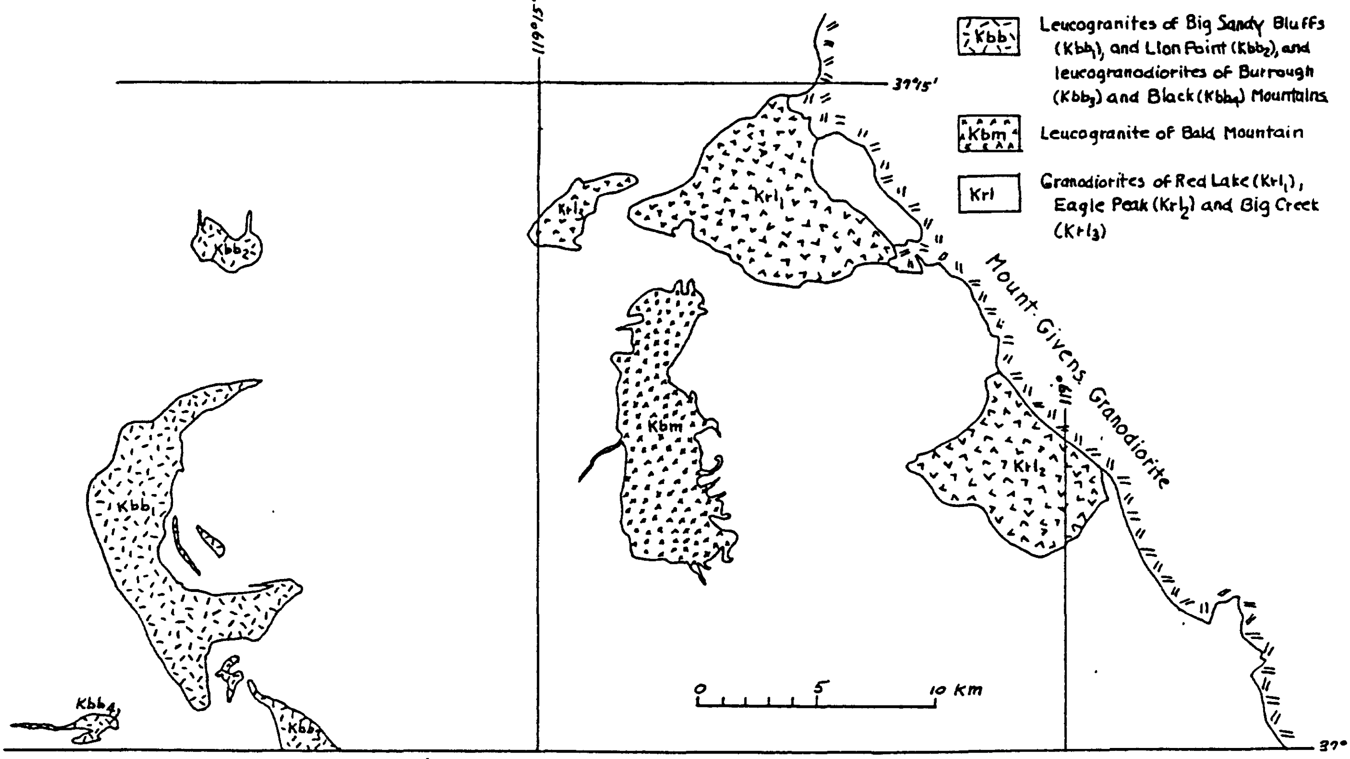


FIGURE 81.--Geologic index map of intrusions west of Mount Givens Granodiorite, which are tentatively assigned to John Muir Intrusive Suite.

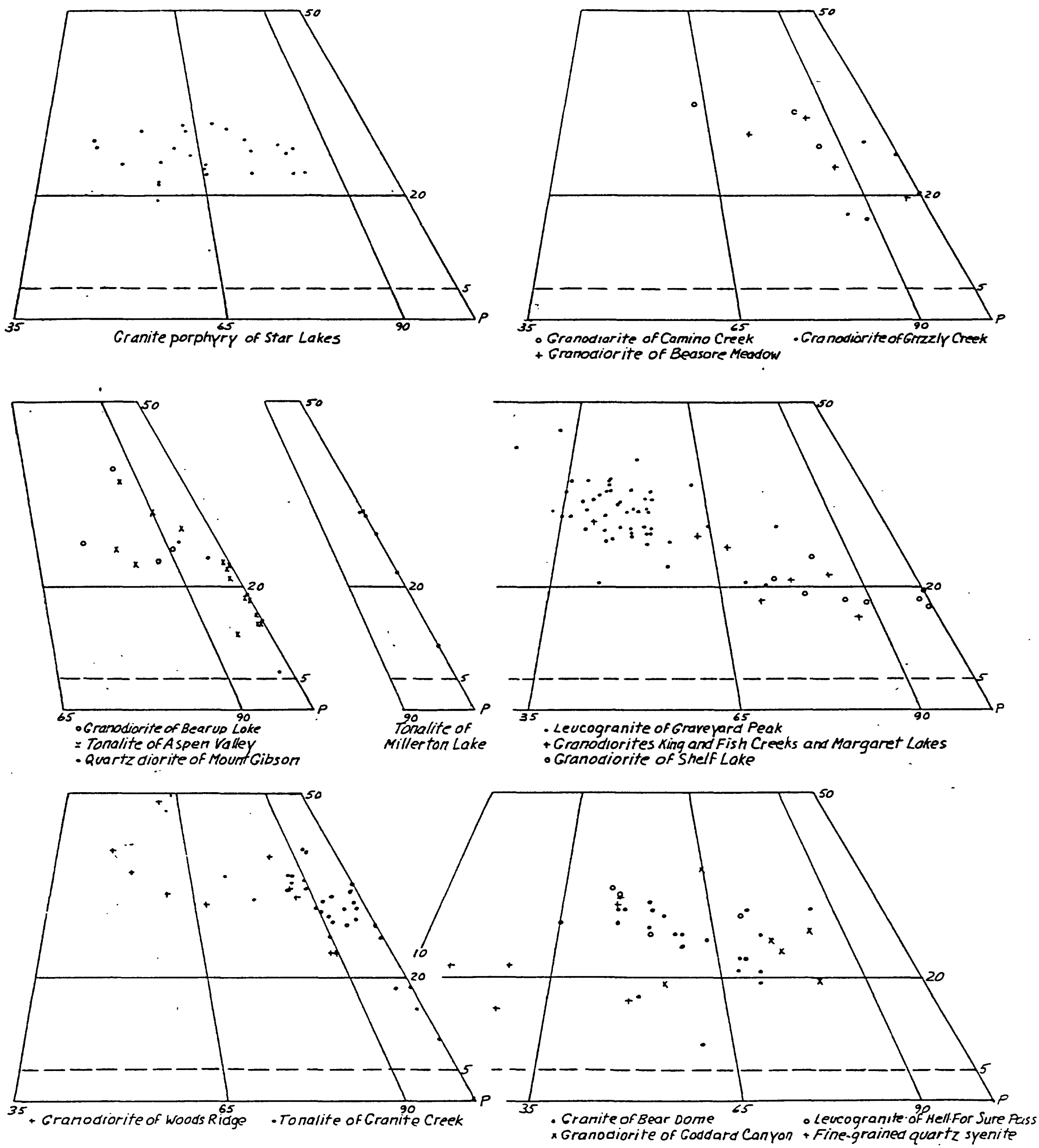


FIGURE 82.--Modes of unassigned intrusions in western and central Sierra Nevada.

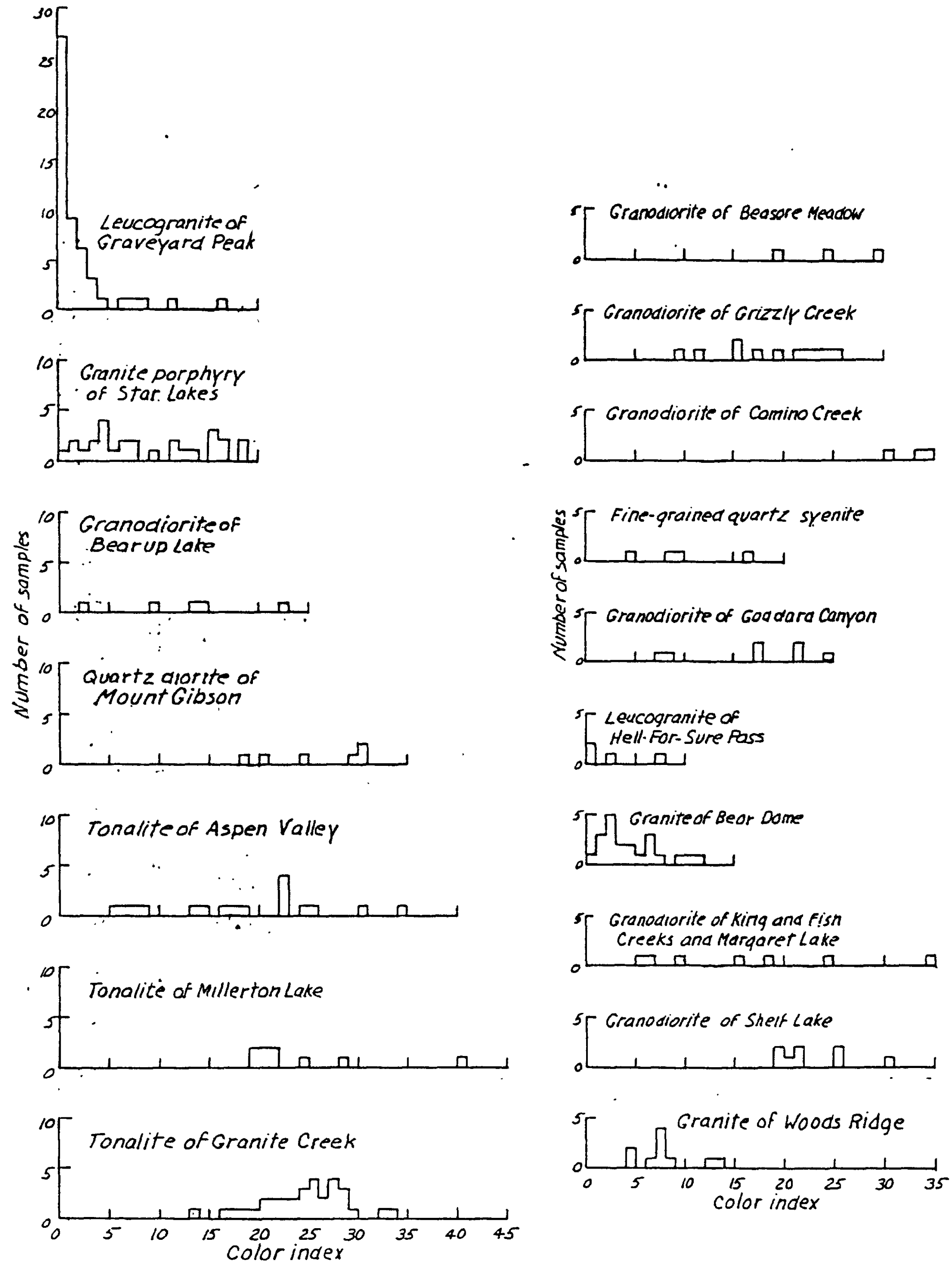


FIGURE 83.--Color index of unassigned intrusions in western and central Sierra Nevada.

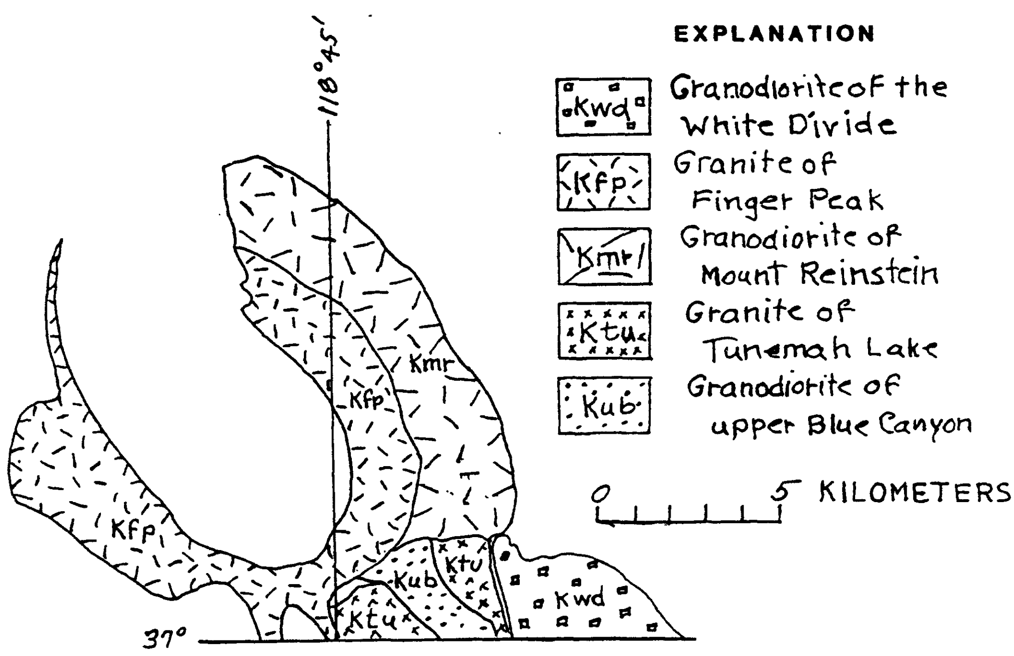


FIGURE 84.--Geologic index map of granitoids southwest of the Mount Goddard septum.

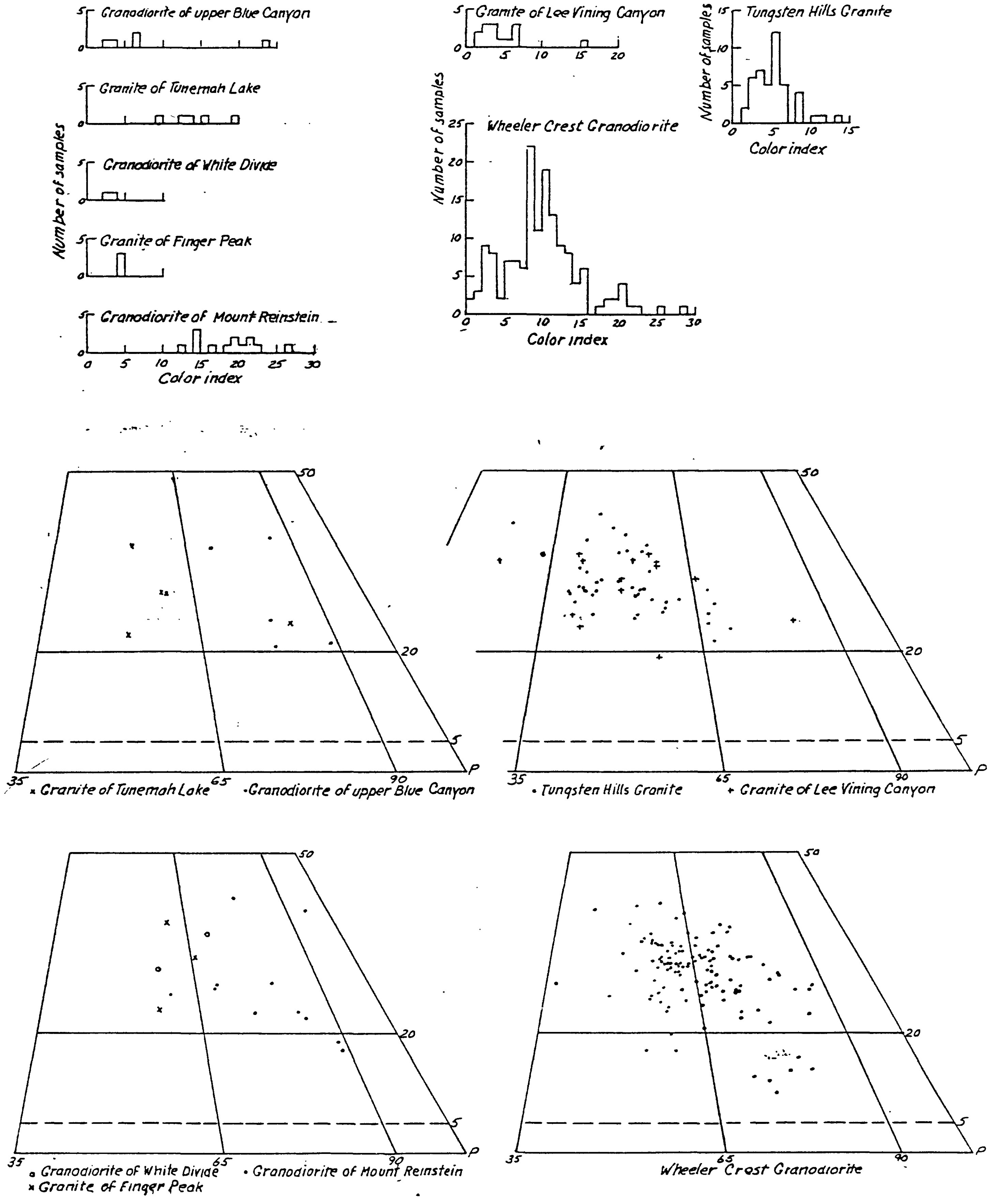


FIGURE 85.--Modes and color index of Scheelite Intrusive Suite and unassigned intrusions southwest of Goddard septum.

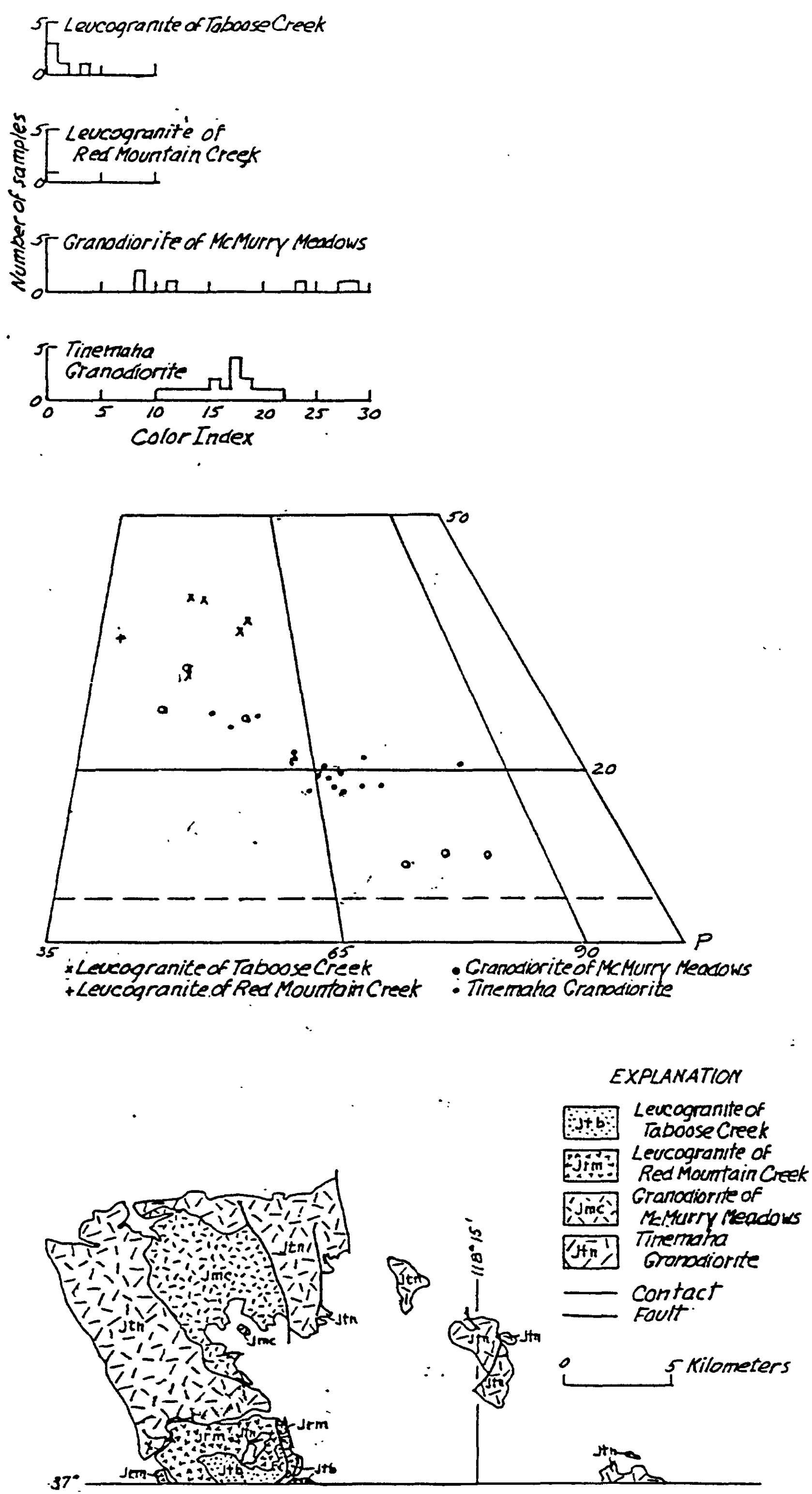


FIGURE 86.--Geologic index map, modes, and color index of Palisade Crest Intrusive Suite. Leucogranites of Red Mountain and Taboose Creek are tentatively assigned to suite.

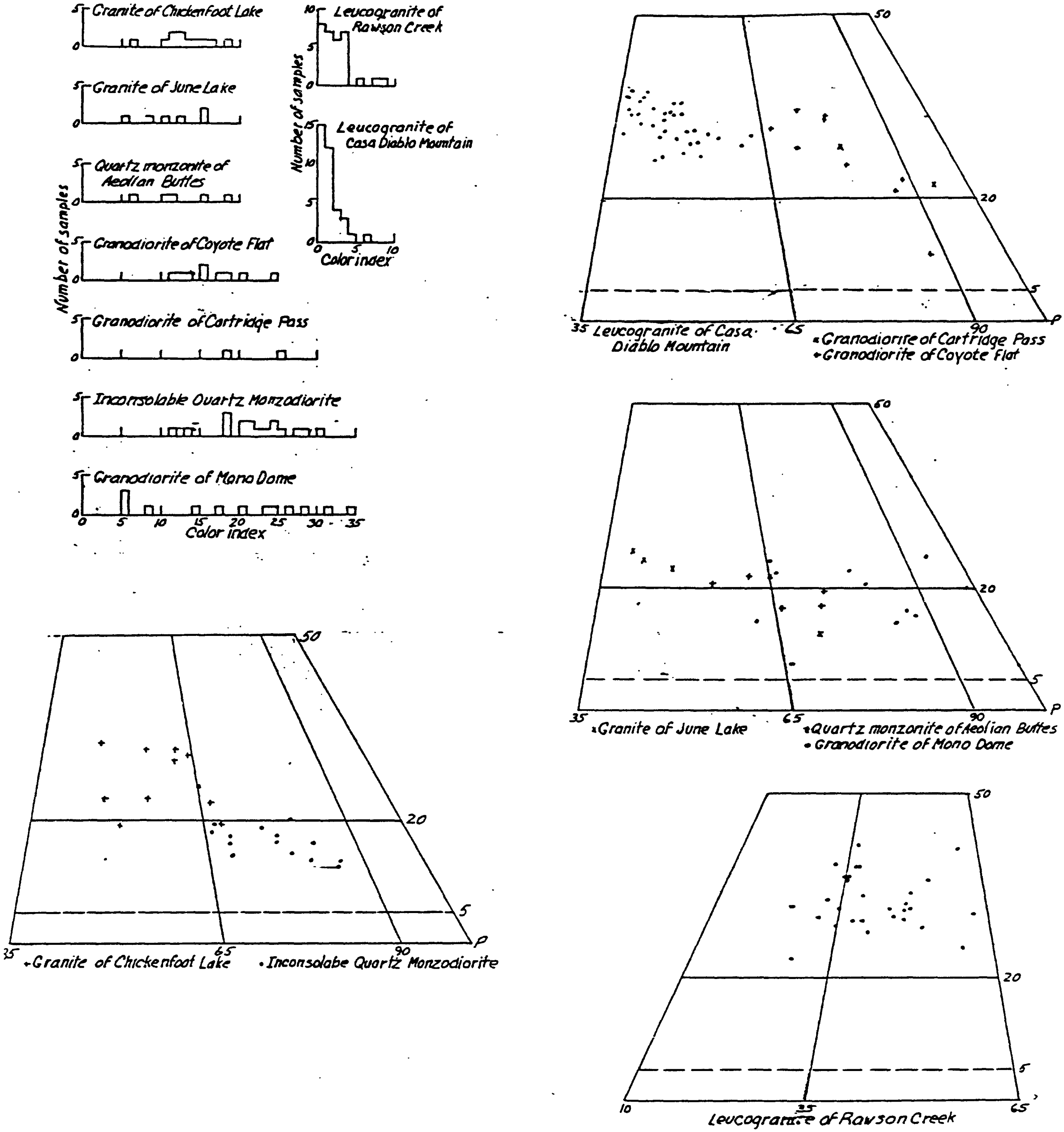


FIGURE 87.--Modes and color index of unassigned granitic rocks of eastern Sierra Nevada and Benton Range.

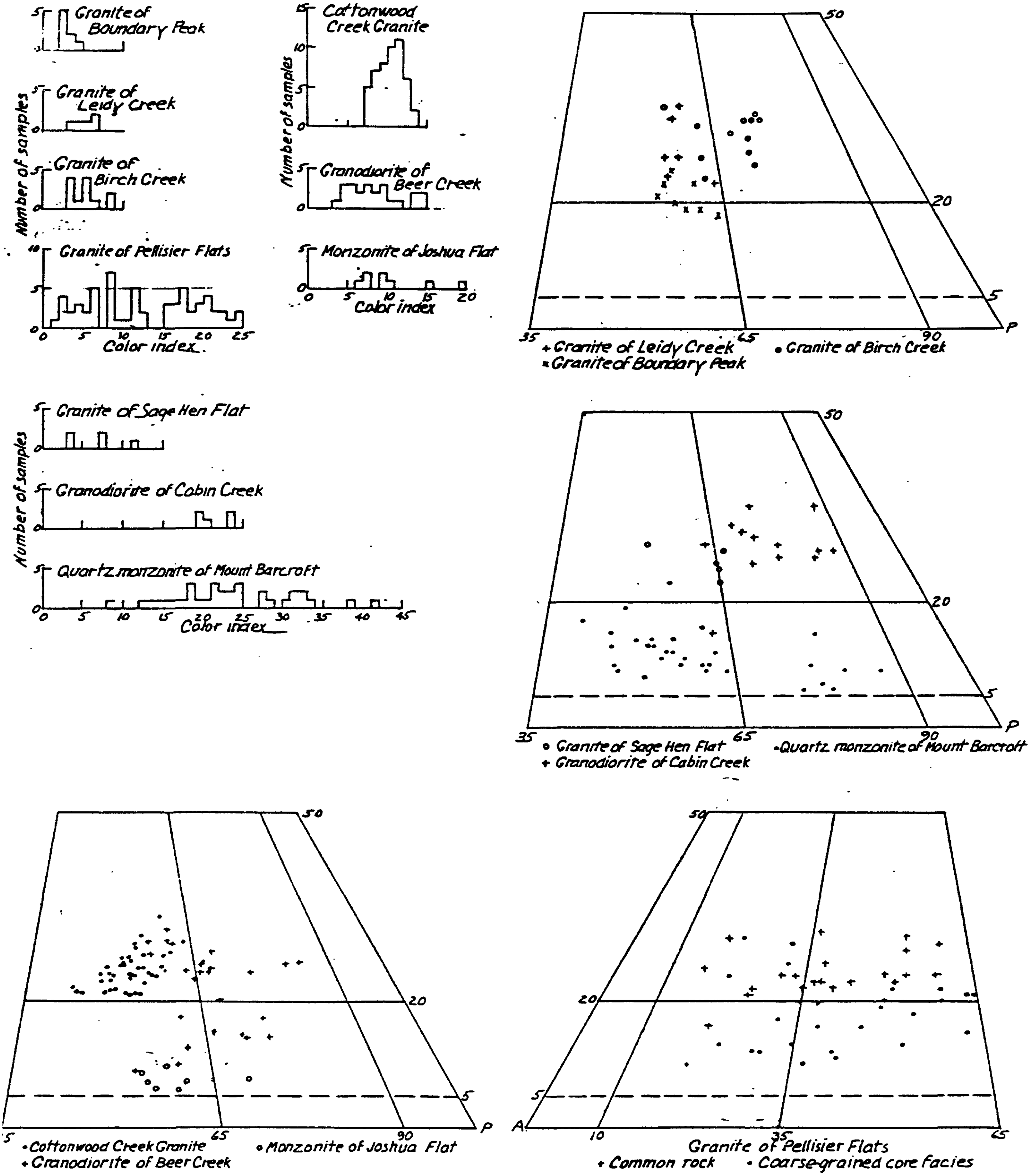


FIGURE 88.--Modes and color index of Soldier Pass Intrusive Suite and unassigned Jurassic and Cretaceous intrusive rocks of White and northern Inyo Mountains.