The Metamorphic and Plutonic Rocks of the Southernmost Sierra Nevada, California, and Their Tectonic Framework

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By DONALD C. ROSS

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THE METAMORPHIC AND PLUTONIC ROCKS OF THE SOUTHERNMOST SIERRA NEVADA, CALIFORNIA, AND THEIR TECTONIC FRAMEWORK

By DONALD C. ROSS

ABSTRACT

The south tip of the Sierra Nevada batholith terminates abruptly along the San Andreas fault and has an anomalous east-west-trending structural grain. Apparently, a part of the batholith has been torn away and displaced by San Andreas fault movement.

The metamorphic framework rocks of the southernmost Sierra Nevada can be divided into (1) a dark gneissic, amphibolitic, and granulitic sequence, in part hornblende rich, that appears to be of oceanic origin; and (2) a quartzofeldspathic sequence of schist, hornfels, impure to pure quartzite, and locally abundant calcareous rocks of continental origin. In addition, large exotic fault slivers of the Rand Schist are present, whose relation to the other metamorphic framework rocks is not yet fully understood.

The southernmost Sierra Nevada batholith is made up of numerous granitic plutons that range in composition from tonalite to granite and, on the basis of sparse radiometric data, appear to be largely, if not wholly, of Cretaceous age. These plutons consist of tonalite on the west and granodiorite and granite on the east. The contact between these two terranes marks the quartz diorite line of Moore (1959). The tonalite of Bear Valley Springs, the largest tonalite body on the west side of the batholith, is characterized by abundant hornblende and biotite in irregular shreds and aggregates, conspicuous foliation defined by abundant dark inclusions, and streaky schlieren and relict gneiss patches that suggest at least wallrock contamination and, possibly, some degree of reconstitution of the gneiss or gabbroic rocks. Most of the other bodies of tonalite and the bodies of granodiorite and granite show minor, if any, contamination effects and appear to have originated from the crystallization of largely uncontaminated silicate melts.

The Garlock fault marks a major structural break between the Sierra Nevada block and the Mojave block. Evidence of offset of (1) garnetiferous amphibolitic gneiss, (2) fault horses composed of the Rand Schist, and (3) the granodiorites of Lebec and Gato-Montes and associated granite plutons suggests 50 km of left-lateral offset at the west end of the fault zone.

Since 1932, when relatively comprehensive instrumentation began to record seismic activity in southern California, 21 earthquakes $M \geqslant 5$ have occurred within the study area. Most of these shocks were associated with the White Wolf fault, which broke in July 1952, resulting in an $M = 7.7$ earthquake attended by considerable property damage and some loss of life. By contrast, the most recent fault movements along the San Andreas fault in this region were in 1857. Although the Garlock and Sierra Nevada faults do not appear to have broken during historical time, they exhibit features that indicate probable Holocene activity.

The most geologically significant feature of the area is a hornblende-rich, gneissic to granoblastic metamorphic terrane of amphibolite to granulite grade that, along with associated diorite to tonalite magmatic rocks, dominates the southernmost Sierra Nevada. The rocks probably reflect a significantly deeper crustal level than the typical Sierran granodiorite and granite plutons to the north, on the basis of (1) index minerals (hypersthene, coarse red garnets, and brown hornblende), (2) textures (considerable ambivalence as to whether individual samples are metamorphic or magmatic), (3) metamorphic grade (at least local granulite facies), and (4) the presence of migmatite and evidence of local melting and mobilization. These rocks may be exposures of the upper part of the root zone and metamorphic substrate of the Sierra Nevada batholith. Xenoliths of gneiss, amphibolite, and granulite from subbatholithic levels, which have been transported upward and preserved in volcanic rocks in the central Sierra Nevada, are similar to some exposed rocks of the southernmost Sierra Nevada.

Hypersthene-bearing granulite and tonalite, as well as distinctive granofels of mid-Cretaceous age, are also exposed in the western part of the Santa Lucia Range (some 300 km to the northwest across the San Andreas fault). The similarity of these rocks to some of the metamorphic and magmatic rocks in the southernmost Sierra Nevada suggests that the two areas record similar metamorphic conditions and crustal depth. The rarity of mid-Cretaceous hypersthene granulite makes correlation of the Santa Lucia Range and the southernmost Sierra Nevada seem attractive. Nevertheless, possibly significant petrographic and rock-distribution differences between the two areas (particularly the relative abundance and distribution of carbonate rocks and amphibolite) dictate caution in suggesting that the two terranes were once contiguous.

INTRODUCTION

The southernmost Sierra Nevada has an anomalous trend in comparison with the rest of the Sierra Nevada block. The generally north-southward structural trends in the northern and central parts of the study area (pl. 1) contrast with the northwestward grain of the Sierran block farther to the north. The radically different east-westward trend of the Sierran tail offers even more contrast. This sharply hooked tail (fig. 1) of the arcuate Sierra Nevada block (Jennings, 1977) suggested to some workers that the south end of the Sierra Nevada has been dragged into its present curve as a result of shifts along northwest-trending fault zones (Locke and others, 1940). Mayo (1947) later expressed doubts that units as huge as the south end of the Sierra Nevada were ever dragged around. Instead, available data suggested to him that the curved structure had grown on essentially its present plan. Studies of regional variations of initial Sr-isotopic composition in the Sierran block and surrounding areas led Kistler and Peterman (1978, fig. 3) to suggest that the Sierran tail is an original feature of an ancient continental margin and adjacent oceanic terrane. Nevertheless, the truncation of the Sierran tail by the San Andreas fault, with its associated

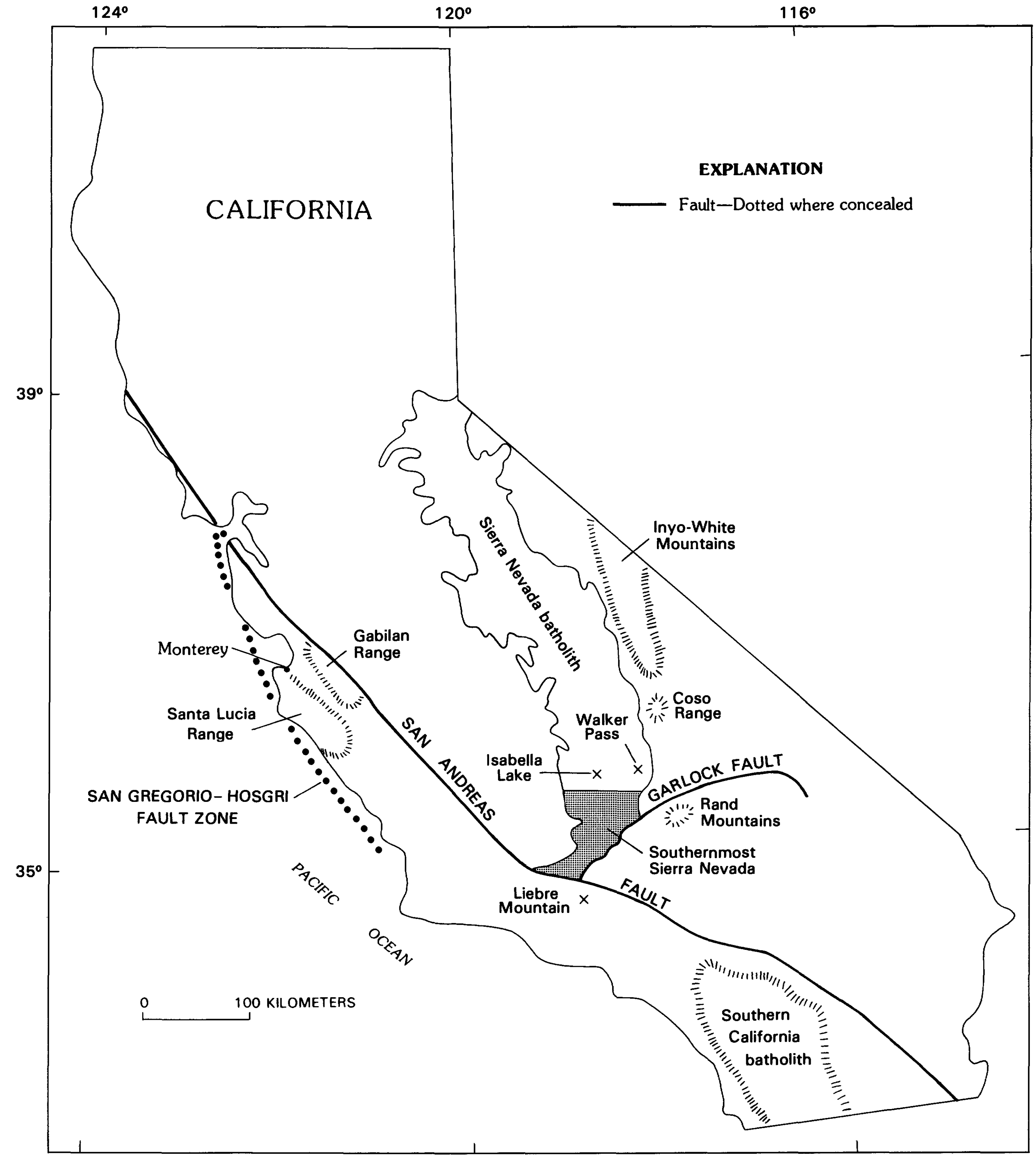


FIGURE 1.—California, showing location and setting of the southernmost Sierra Nevada.

right-lateral offset measured in hundreds of kilometers, has suggested to many previous workers, as well as to me, a possible drag effect as a result of movement on the San Andreas fault.

Probably the most noteworthy tectonic feature of the southernmost Sierra Nevada is its abrupt termination against the Garlock and San Andreas faults. However, correlation of basement units and structures across the fault zone (Smith, 1962; Smith and Ketner, 1970; Davis and Burchfiel, 1973) suggests that the block may be continuous southward across the Garlock fault, with displacements of only a few tens of kilometers. Regional continuity is also suggested by a north-southerly trending belt of Cordilleran miogeoclinal Paleozoic rocks, extending along the east side of the Sierra Nevada batholith from Nevada, that also are patchily preserved in the central Mojave Desert and in the San Bernardino Mountains (Stewart and Poole, 1975).

Nevertheless, the composition of the granitic rocks of the western Mojave wedge—between the Garlock and San Andreas faults—suggests problems with the correlations noted above. The quartz diorite line of Moore (1959), which passes through the length of the Sierra Nevada batholith, separating dominantly granite and granodiorite plutons on the east from dominantly tonalite plutons on the west, is not found in the Mojave block. If the Mojave block is a southward continuation of the Sierra Nevada and if the quartz diorite line has been offset by movement on the Garlock fault, one would expect to find tonalitic rocks in the western Mojave Desert. Yet the entire western Mojave wedge has only the granitic rocks that are typical of the terrane east of the quartz diorite line. Clearly, this discrepancy must be accounted for in any regional reconstruction. Kistler and Peterman (1978) suggested, again on the basis of Sr-isotopic data, that the western Mojave wedge is an exotic mass from northwestern Nevada which has accreted to the eastern Mojave. At present, insufficient petrographic and chemical data are available to confirm or deny this intriguing thesis. The truncation of the Sierran block by the San Andreas fault is much more severe, and displacements of at least several hundred kilometers across this fault are virtually universally accepted. To the best of my knowledge, the whereabouts of the lost pieces of the Sierran tail is still speculative.

A major purpose of this report is to provide a detailed description of the basement rocks of the southernmost Sierra Nevada so as to place constraints on the origins of terranes across the Garlock and San Andreas faults suggested to have originally been parts of the Sierra Nevada block. Although plate 1 shows generalized Cenozoic units that overlie and conceal the basement, my study was restricted to the pre-Cenozoic metamorphic and plutonic rocks. This study relied heavily on sampling of the metamorphic and plutonic rocks for thin-section studies, modal analyses, and chemical analyses (pl. 2). The study involved about 150 man-days of fieldwork in the 3,900-km^2 area; about 800 thin sections were examined, and nearly 700 modal analyses were determined.

This study makes use of previous mapping where completed and integrates that earlier work with new geologic mapping in unmapped areas at a scale of 1:125,000. The geology from previous maps had to be somewhat modified to make a generally consistent map of the basement rocks, and compromises were made. Given the limited time spent in the field, selective traverses were required, and so some areas were not visited. Many unresolved problems remain concerning the east-west-trending Sierran tail west of Tejon Creek (fig. 2), and followup detailed work is particularly needed there. I have tried to provide not a definitive report but a base on which to build.

ACKNOWLEDGMENTS

The fieldwork was made possible, and much eased, by the friendly cooperation of ranchers and landowners, too numerous to mention, who allowed me access to private land; special thanks go to Duncan V. Patty of the Tejon Ranch and Mr. and Mrs. Jack Porch of the San Emigdio Ranch for their help and advice.

On many isolated traverses, my point of contact with the outside world in case of trouble was the Kern County Fire Department; I thank the men of the Lebec and Tehachapi Stations.

For enduring the tedium of slabbing, staining, and point-counting nearly 700 granitic specimens for modal analysis, I especially thank Hector Villalobos, Jenny Metz, and Terry Kaplan.

In the course of this study, I have leaned heavily (and with appreciation) on the work of other geologists cited on the index map of plate 1 and in the section below entitled "References Cited"—especially T.W. Dibblee, Jr., whose published and unpublished mapping covered almost the entire area. Although he and I did not always agree on the name that should be applied to a particular basement-rock unit, I could almost always count on his mapping to show faithfully the distribution of the various units.

METAMORPHIC ROCKS

METASEDIMENTARY ROCKS OF THE KEENE AREA

SETTING

The name "Kernville Series" (Miller, 1931) is widely used for metasedimentary rocks in the southernmost Sierra Nevada (Dibblee and Chesterman, 1953; Dibblee

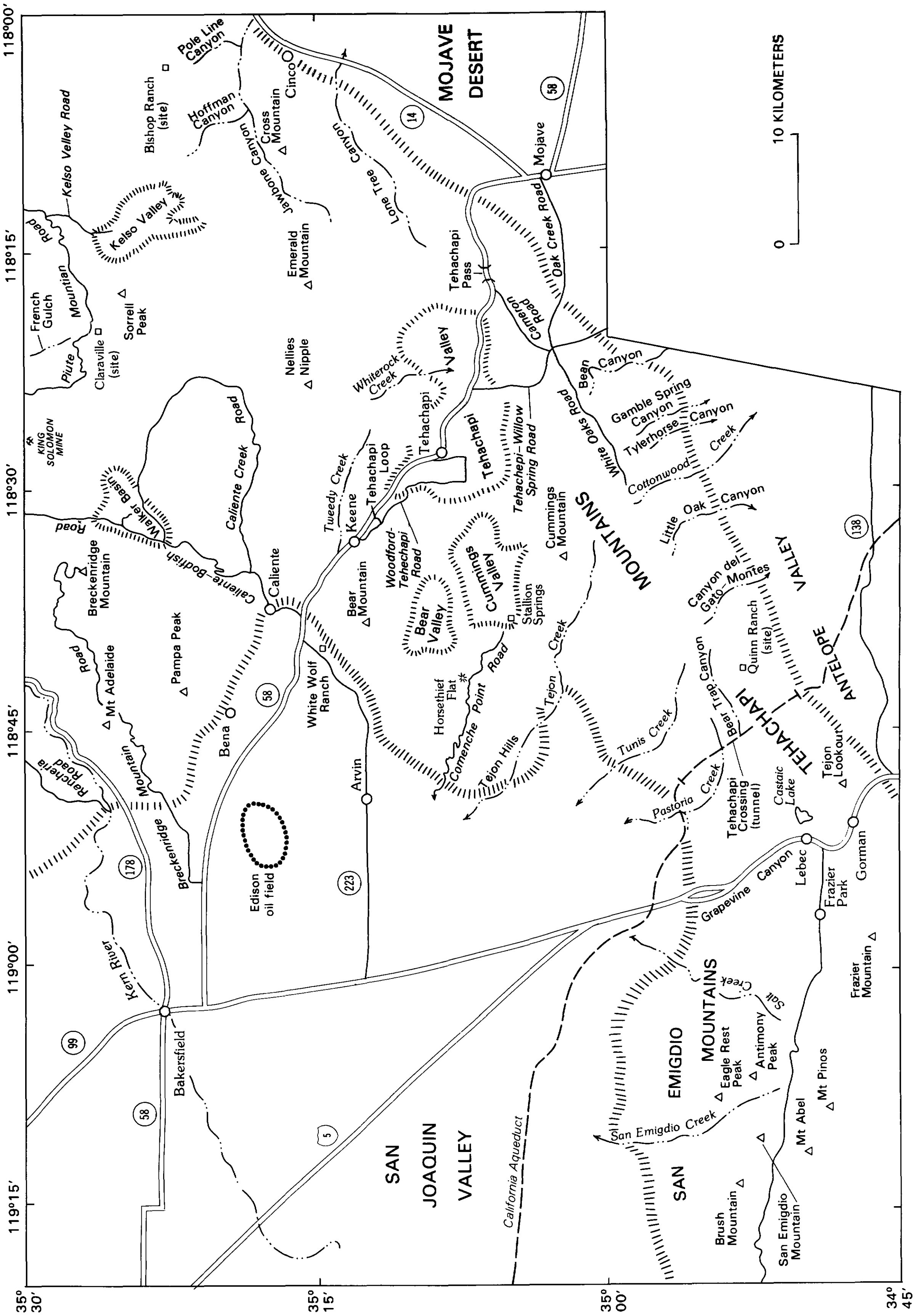


FIGURE 2.—Selected physical and cultural features in the southernmost Sierra Nevada.

and Warne, 1970; Dibblee and Louke, 1970). The Kernville Series is divisible into several metamorphic units in its type area north of the study area; furthermore, the term "series" is no longer used in this manner in modern stratigraphic nomenclature. Therefore, I favor using a local name (Keene) for these, at present, sparsely studied metasedimentary rocks south of lat 35°30′ N. and north of the Garlock fault. The town of Keene (approx 10 km northwest of Tehachapi) is located in the midst of several bodies of these metasedimentary rocks.

The metasedimentary rocks of the Keene area form a broad belt of discrete, rather large pendants or screens that generally trend north-south. The approximate total area of outcrop is 185 km^2. These pendants are strongly elongate north-southerly, generally dip steeply, and form an open S-shaped outcrop belt from the north end of the study area to the Garlock fault. One major exception to this pattern is a dome of metamorphic rocks with gentle dips in the southeastern part of the Emerald Mountain quadrangle. The metasedimentary rocks of the Keene area do not appear to extend westward of Tejon Creek, and they are present north and west of the White Wolf and Breckenridge faults only in two small exposures west and southwest of the Walker Basin. To the east, the belt extends to the southwestern part of the Cross Mountain quadrangle.

Saleeby and others (1978) considered these rocks to be part of their Triassic and Jurassic Kings sequence, which encompasses roof pendants in the western and central Sierra Nevada for a strike length of more than 200 km. They expanded the area of the Kings sequence, which was originally limited by Bateman and Clark (1974) to the area between Dinkey Creek and Mineral King.

No fossil evidence for the age of these rocks was found within the study area, although Saleeby and others (1978) noted that the bivalves found southeast of Lake Isabella are Late Triassic to Early Jurassic in age. These fossils were collected about 15 km north of the study area, from a calc-hornfels associated with quartzite, marble, quartz-mica schist, and andalusite-biotite schist—rock types typical of the Keene area.

ROCK TYPES AND PETROGRAPHY

Marble, pure to impure quartzite, and mica schist that is commonly coarse grained and contains sillimanite and garnet are the most common rock types in the study area. Quartzofeldspathic hornfels to schist and calc-hornfels are widespread, though less common. In my sampling and petrographic study, I observed no major differences between the various pendants. Probably, there is more marble in the broad, eastern belt of pendants that extends through the Tehachapi quadrangle into the Emerald Mountain quadrangle, and more relatively pure quartzite in the thinner western belt of pendants that trends north-south through the Cummings Mountain quadrangle up into the Breckenridge Mountain quadrangle. Although these differences might reflect original stratigraphic distinctions, the present dismemberment and contortion of the section preclude any firm statement.

The marble is fine to very coarse grained and generally white to gray, and contains trains and stringers of graphite flakes and other metamorphic minerals that probably reflect original bedding. Dolomite is present, but calcite marble predominates. Associated calc-hornfels layers range from trains of calc-silicate grains in relatively pure marble, through thin calc-silicate layers in marble, to calcareous metashale and quartzite. Pale-green clinopyroxene is by far the most common calc-silicate mineral. Some thin layers rich in plagioclase and dark hornblende could be metavolcanic rocks, but the close association of clinopyroxene, epidote, and thin marble layers points to a sedimentary origin. Garnet-clinopyroxene-epidote tactite also occurs along and near some marble contacts with granitic rocks.

Quartzite, ranging from a virtually pure mat of quartz crystals to layers that contain 5 to 25 percent of impurities, is conspicuous and mappable in some of the pendants. In addition, much more is present in layers and lenses too small to show at the scale of the map (pl. 1). The purest quartzite contains very scattered crystals of clinopyroxene, feldspar, or mica that probably reflect original argillaceous and calcareous cement. Many layers seem to contain about 5 to 15 percent of these impurities as scattered crystals and trains and small lenses that mimic original bedding. From these quartz-rich rocks there are gradations to quartz hornfels containing 40 to 50 percent impurities. The quartz-rich rocks commonly also contain garnet, tourmaline, and graphite. Some rounded zircon crystals are probably original detrital grains. Locally, there are layers rich in apatite crystals, which may reflect original phosphatic beds.

One of the more distinctive rock types is a coarsely crystalline mica schist. This schist is characterized by abundant red to brown biotite, muscovite, and sillimanite; the sillimanite occurs both in fibrous bundles and coarse prismatic crystals. Red garnets are also common in these rocks. Large, irregular andalusite crystals are present in some sillimanite-bearing rocks. Although most of the rocks are dominated by mica, others contain abundant K-feldspar, lesser plagioclase, and quartz. In some specimens, original clastic quartz grains are preserved where they have been cushioned by metamorphic mica. Tourmaline and graphite are present locally. These dark, red- to brown-weathering rocks produce distinctive shiny, coarse-grained, mica-rich outcrops.

METASEDIMENTARY ROCKS OF SALT CREEK

SETTING

Numerous pendants and inclusions of metasedimentary rocks occur along the entire outcrop length of the granodiorite of Lebec, from about 40 km westward from the Garlock fault into the outcrop area of the granite of Brush Mountain and to the San Andreas fault. The largest body, an elongate mass near the center of this belt, is cut by Salt Creek and covers an area of about 10 km^2. Most of the bodies are small, however, and in aggregate these metamorphic rocks cover no more than 25 km^2. The strike of these bodies is grossly east-westerly; generally they dip moderately to steeply, but with many contortions both in strike and dip. Some of the masses appear to be slivers along the faulted north margin of the granodiorite of Lebec, but most are completely surrounded by granitic rocks.

ROCK TYPES AND PETROGRAPHY

Although considerable variation exists in these metamorphic rocks, they can generally be lumped into three rock types: (1) marble, largely calcitic, but in part dolomitic; (2) pure to impure quartzite, containing 70 to 95 percent quartz; and (3) micaceous schist that commonly contains sillimanite. Calc-hornfels and quartzofeldspathic hornfels are less common, but widespread.

Marble makes up most or all of several small masses in the central and eastern parts of the metamorphic belt, but it is more commonly interbedded with other rock types to the west.

Quartzite can be delineated locally, but generally quartzite and schist are so interlayered that they are not mappable separately at the scale of plate 1. The protoliths of this section probably were dominantly limestone, quartz sandstone, argillaceous and calcareous sandstone, and shale that was richly argillaceous. These lithologies suggest a continental sedimentary section that was deposited near a stable platform, probably in a miogeoclinal environment. The rock types of Salt Creek resemble those of the Keene area.

Marble is the most conspicuous rock type. It forms highly visible white outcrops that have been quarried to some extent. Some color layering, possibly reflecting bedding, is present, and some layers are speckled with shiny black graphite flakes. Locally, the marble is ribbon rock, consisting of alternating thin layers of calcite and calc-hornfels made up of quartz, clinopyroxene, epidote, and feldspar. Some of the masses mapped as marble on plate 1 include calc-hornfels and siliceous layers. Marble interbeds also occur locally within other rock types.

The rock type most characteristic of the metamorphic belt is red- to brown-weathering, relatively pure quartzite. Rarely the dominant rock type in any one pendant or inclusion, it is widespread throughout this belt. The nearly pure quartzite is a granoblastic mat of 90 to 95 percent quartz, with scattered flakes of mica and feldspar; the feldspar probably reflects reconstituted argillaceous cement. These rocks range in composition from nearly pure quartzite to argillaceous and feldspathic quartzite containing more than 25 percent cement. Clinopyroxene, tremolite, and epidote in some of these rocks probably represent original calcareous cement. Tourmaline, garnet, and sillimanite also are found locally. In some specimens, original quartz grains, from 0.4 to 1 mm long, are preserved where they are protected by nests of reconstituted cement. Thus, the pure quartzite is an end member in a sequence that grades with the addition of mica, feldspar, and other minerals into quartz hornfels and highly micaceous schist.

Micaceous schist is the second most abundant rock type in the metamorphic belt. This shiny, well-foliated rock is gray on fresh surfaces and weathers to dark shades of brown or red. Some schist is composed dominantly of reddish-brown biotite and untwinned K-feldspar, with minor muscovite, quartz, and plagioclase. Other schist is dominated by intermediate plagioclase and biotite. In some layers, muscovite is abundant, but rarely is it more abundant than biotite. Sillimanite is notably abundant in these rocks, in part in coarse skeletal crystals, but more commonly in bundles of acicular crystals. Red garnets are also present in some layers. Shimmer aggregates of muscovite are rather widespread. Cordierite, which has been identified in two of these rocks, probably was originally more widespread and has been largely converted to muscovite.

As previously noted, the metamorphic belt contains rock types intermediate between quartzite and micaceous schist. Most of these rocks are strongly foliated and probably are best classified as quartzofeldspathic schist to hornfels. Some of the schist is dominantly a granoblastic mat of varying proportions of quartz, intermediate plagioclase, and K-feldspar, with a directed fabric shown in thin section only by aligned reddish-brown biotite and lesser muscovite. Small amounts of pink garnet and green hornblende are also present.

Two unusual rock types with very limited outcrop bear special note. Near the center of the largest pendant, dark layers in a dominantly quartzite sequence are composed of a decussate mat of olive-brown hornblende, minor epidote, and some sphene. These dark layers are partly replaced by light-colored thin layers and narrow tongues consisting of plagioclase (approx An_{50}), pale-green clinopyroxene, and abundant tiny round grains of highly pleochroic sphene. Some calcite is present locally in the lighter layers. The general appearance, composition, and setting of this rock suggest that it was derived from a calcareous sedimentary rock.

A dark chloritic schist was noted in a poorly exposed sequence of quartzite and quartzofeldspathic schist about 700 m S. 60° E. from San Emigdio Mountain along the road that skirts a tremendous cliff (landslide scarp?). The rock is composed almost entirely of pale-green chlorite and colorless to pale-green amphibole, with abundant, scattered opaque metallic grains and calcite. Although it is tempting to suggest that this rock is a volcanic greenschist, the absence of feldspar, the presence of abundant calcite, and the geologic setting suggest otherwise. I could speculate that this layer is a retrograde hornblendite dike or sill, but it seems more reasonable to suggest that it was a sedimentary layer. By any account, this is an unusual rock type for this area, although some chlorite is present in other schistose layers in the metamorphic belt. The occurrence of these two rock types points out the perils of assuming that hornblende-rich metamorphic rocks have an igneous origin.

GNEISS, AMPHIBOLITE, AND GRANULITE OF SAN EMIGDIO-TEHACHAPI MOUNTAINS

SETTING

High-grade, generally dark-colored metamorphic rocks form a distinctive belt along the north slope of the San Emigdio and Tehachapi Mountains from San Emigdio Creek eastward to Tejon Creek; they underlie an area of about 300 km^2. The characteristic rock types are gneiss, amphibolite, granofels, granulite, and unquestionably metasedimentary rocks (impure quartzite and calc-hornfels). Throughout the largely mafic metamorphic terrane, coarse red garnets are a widespread and striking feature. These garnets, as large as 10 cm in diameter and commonly accentuated in outcrop by bleached haloes, are discussed below in more detail in the subsection entitled "Coarse-Haloed Red Garnet." Gradation between rock types and intermixing are characteristic, and the rock types cannot be separated in most places by reconnaissance mapping. Felsic quartzofeldspathic gneiss that has both metasedimentary and metaigneous components is separable locally at the map scale. Slivers of similar and presumably correlative, red-garnet-bearing gneiss and amphibolite are strung out along the Garlock fault from the area of Tehachapi Pass eastward to Cinco. Within the largely dark gneissic terrane are areas of relatively homogeneous diorite to tonalite that are partly intrusive and magmatic. Indistinct, subtle contacts and migmatitic mixing with gneiss and amphibolite suggest, however, that at least some of the diorite to tonalite is a locally sweated out, ultrametamorphosed derivative of the gneiss and amphibolitic terrane. These relatively homogeneous and, at least in part, intrusive rocks are discussed separately below in the subsection entitled "Granitic Rocks Related(?) to the San Emigdio-Tehachapi Terrane."

DESCRIPTION OF ROCK TYPES

Only a summary description of the major rock types in the study area is presented here. Additional information on the textures and mineral contents of selected thin sections and localities where the major rock types can be observed and sampled was presented by Ross (1983c).

GNEISS

Strongly foliated gneiss is probably the dominant rock type, and dark, hornblende-rich gneiss is most common. These rocks are probably best exposed along the road by the Tehachapi Crossing of the California Aqueduct. Many of the gneissic rocks have a mineralogy of quartz diorite or tonalite and contain sharply twinned andesine, abundant quartz, olive-brown hornblende, and brown biotite. Also abundant is amphibolitic gneiss, composed almost entirely of intermediate plagioclase and pale- to dark-green and brown hornblende. Much of the gneiss is to some extent cataclastic.

A belt of light-colored gneiss lies along the north slope of the San Emigdio Mountains and in the area of Grapevine Canyon is bounded rather sharply against dark gneiss on the south. The felsic gneiss is rich in quartz and sharply twinned andesine, and contains widely scattered biotite and hornblende crystals. Pink garnet is common and occurs in trains that accent the foliation. Coarse-grained graphite is abundant in some layers. Much of the foliation is imparted by well-aligned trains and layers of biotite. Surprisingly, much of the felsic gneiss contains little or no K-feldspar. Augen and flaser structures are well developed in some of the felsic gneiss layers. Much felsic granitic rock intrudes the felsic gneiss in and west of Grapevine Canyon. In places, the granitic material dominates, but there is generally ample evidence of gneissic host rock. Subordinate layers of dark gneiss and amphibolite are present in the felsic gneiss areas.

Although much of the hornblende-rich mafic gneiss is probably metaigneous (orthogneiss), the protolith is less evident for much of the felsic quartzofeldspathic gneiss. Some quartz- and mica-rich layers containing abundant garnets and graphite almost surely had siliceous and pelitic sedimentary protoliths. As an aid for distinguishing orthogneiss from paragneiss, some preliminary determinations of the $\delta^{18}O$ of selected samples have been made (fig. 3; table 1). A biotite quartzofeldspathic gneiss (sample 3933) from near the mouth of Tejon Creek has a $\delta^{18}O$ value that indicates very little enrichment (8.72 permil

Standard Mean Ocean Water [SMOW]; Ivan Barnes, written commun., 1983); this rock is a foliated igneous rock. A quartzofeldspathic gneiss (sample 3058B) from the north side of the San Emigdio Range west of Grapevine Canyon and two more samples (3369, 3420) from east of Grapevine Canyon have $\delta^{18}O$ values that indicate considerable enrichment (13.82, 14.01, and 14.97 permil SMOW, respectively; Ivan Barnes, written commun., 1983); these gneiss samples most probably have a sedimentary protolith.

Amphibolite

The dark, hornblende-rich gneiss commonly grades into more massive rocks composed almost exclusively of intermediate plagioclase and hornblende. Coarse red garnets with bleached haloes in poikilitic anhedral masses to euhedral crystals, as large as 10 cm in diameter, are particularly abundant in the amphibolitic rocks. Clinopyroxene, biotite, and quartz are subordinate in some samples.

It is commonly difficult to decide whether a given specimen has a xenoblastic (metamorphic) or a xenomorphic granular (magmatic) texture. Wiese (1950) called many of these amphibolitic rocks diorite and gabbro. After much confusion and indecision, I began to notice that the homogeneous, magmatic-looking rocks invariably were closely associated with unquestionably metamorphic rocks and were not in discrete large masses. I concluded that the dominant fabric of these dark rocks was metamorphic, although I fully realized that there probably are local melted and mobilized spots in the amphibolitic rocks. The hornblende-rich diorite to tonalite of the Tehachapi Mountains (see section below entitled "Plutonic Rocks") constitutes the sum total of these local homogeneous areas that appear to grade from dark amphibolite and gneiss.

Two samples of amphibolite (3886B, 3091B, table 1) have $\delta^{18}O$ values of 6.45 and 9.12 permil SMOW, respectively (Ivan Barnes, written commun., 1983). The first sample is definitely metaigneous, but the second sam-

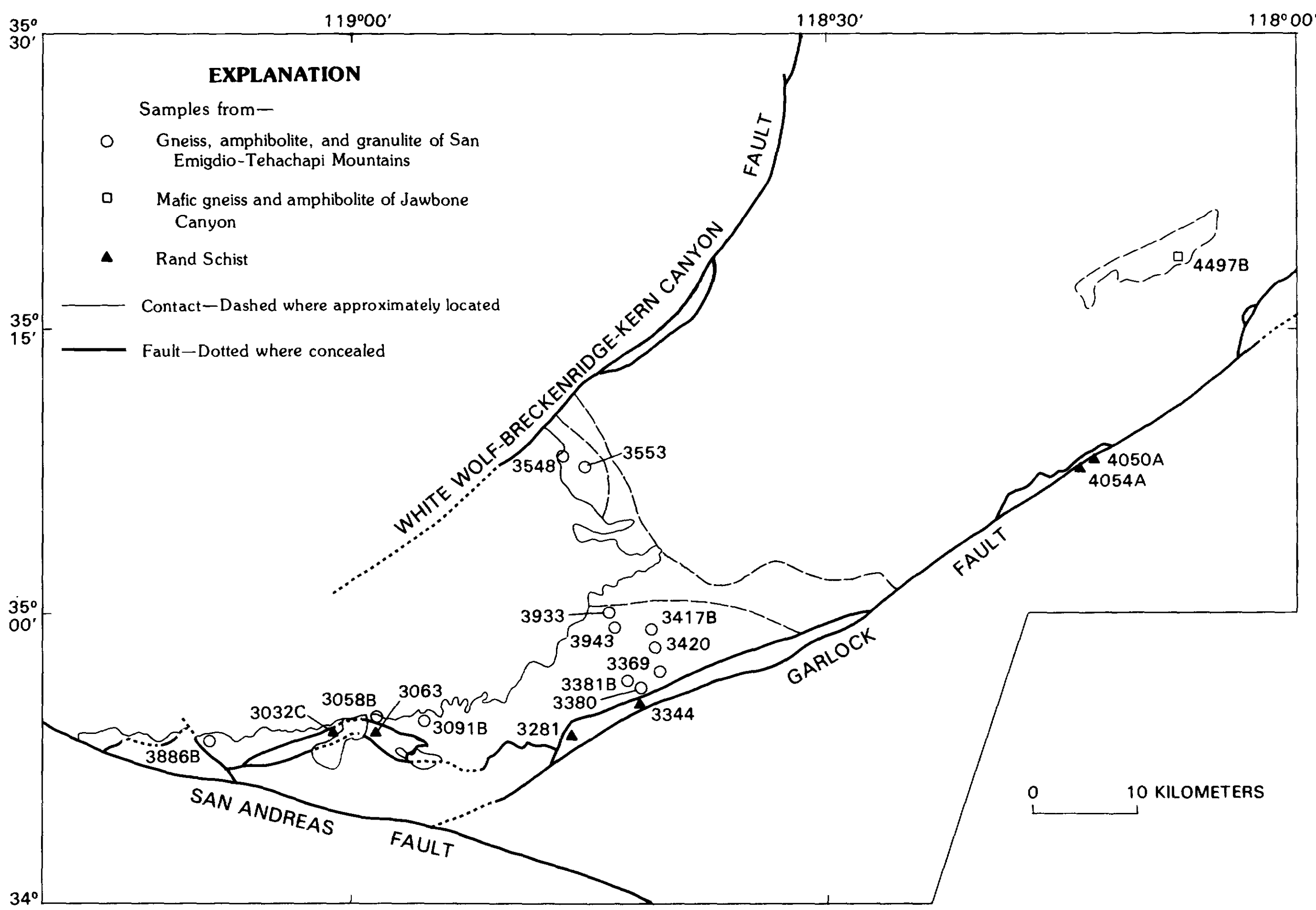


FIGURE 3.—Skeletonized geologic map of the southernmost Sierra Nevada showing locations of samples for which oxygen-isotopic data ($\delta^{18}O$ permil SMOW) have been determined. Data and sample descriptions are listed in tables 1 and 2. See plate 1 for detailed geology.

TABLE 1.—*Oxygen-isotopic data for selected samples from the mafic terranes of the San-Emigdio-Tehachapi Mountains and Jawbone Canyon*

[See figure 3 for locations of samples. Minerals listed in order of decreasing abundance. All values in permil standard mean ocean water (SMOW). Data from Ivan Barnes (written commun., 1983)]

Sample	Rock type	Mineral content	$\delta^{18}O$
3058B	Quartzofeldspathic gneiss---	Quartz, andesine, reddish-brown biotite, muscovite, graphite, garnet, apatite.	14.01
3091B	Amphibolite-----------------	Andesine, olive-brown hornblende, quartz, garnet, reddish-brown biotite, opaque minerals, sphene, zircon.	9.12
3369	Quartzofeldspathic gneiss to granofels.	Quartz, plagioclase (approx An_{50}), clinopyroxene, reddish-brown biotite, garnet, graphite, hornblende.	13.82
3380	do------------------------	Quartz, calcic andesine, reddish-brown biotite, garnet, graphite.	14.97
3381B	Quartzite-------------------	Quartz (90-95 percent), reddish-brown biotite, pink garnet, chlorite, graphite, muscovite.	19.18
3417B	Graphitic quartzite---------	Quartz (75 percent), graphite, reddish-brown biotite, pale-brown hornblende, plagioclase, apatite.	16.82
3420	Hypersthene quartzofeld-spathic granofels (granulite).	Plagioclase (approx An_{50}), quartz, reddish-brown biotite, hypersthene, clinopyroxene(?), zircon, apatite, graphite.	14.97
3548	Calc-hornfels---------------	Clinopyroxene, plagioclase, quartz, epidote-clinozoisite, sphene, zircon.	12.59
3553	Impure quartzite------------	Quartz (75-80 percent), plagioclase, reddish-brown biotite, red opaque-mineral grains ("hematite").	13.12
3886B	Amphibolite, fine-grained, schistose.	Andesine, gray-green hornblende, quartz, opaque minerals, epidote.	6.45
3933	Quartzofeldspathic gneiss---	Quartz, andesine, brown to reddish-brown biotite, pale-olive hornblende, apatite, allanite, epidote.	8.72
3943	Hypersthene amphibolite (granulite).	Labradorite, gray-green to olive hornblende, hypersthene, quartz, opaque minerals.	7.68
4497B	Amphibolite (Jawbone Canyon body).	Plagioclase (approx An_{50}), green hornblende, brown biotite, opaque minerals, epidote, prehnite.	8.94

ple is enriched enough to suggest some sedimentary contamination.

Granofels

Widespread and abundant granofels is essentially a more massive equivalent of the mafic and quartzofeldspathic gneiss. The distinction between granofels and tonalite is problematic in places. All gradations occur from highly foliated gneiss to massive granofels, with an increasing content of quartz to impure quartzite. Mostly, these rocks are medium grained (locally fine grained), have a dominantly granoblastic texture, and are composed of varying amounts of plagioclase, hornblende, quartz, biotite, and clinopyroxene. Pink to red garnets and coarse gray, shiny graphite flakes are characteristic of many layers that bear a striking similarity to rocks in the western Santa Lucia Range south of Monterey (Ross, 1976c).

Granulite

The granulite includes only hypersthene-bearing rocks of amphibolite and granofels composition. It is particularly significant because it signals the presence of granulite-grade metamorphism. I have observed and collected hypersthene granulite samples from only a few localities that mark the minimum extent of confirmed granulite-grade rocks.

Impure Quartzite and Calc-hornfels

Quartzite is scattered throughout the mafic terrane. It is composed dominantly of quartz and varying admixtures of plagioclase, hornblende, biotite, K-feldspar, and muscovite. These quartz-rich rocks are unquestionably metasedimentary. Three selected samples (3553, 3417B, 3381B, table 1) have highly enriched $\delta^{18}O$ values of 13.12, 16.82, and 19.18 permil SMOW, respectively (Ivan Barnes, written commun., 1983); the second and third samples may be metachert.

Calc-hornfels is rare in the San Emigdio-Tehachapi terrane. Fine-grained, gray-green laminated rocks are rich in plagioclase, clinopyroxene, and epidote. Hornblende and quartz are also present. One sample has an enriched $\delta^{18}O$ value of 12.59 permil SMOW, indicating a sedimentary protolith.

MAFIC METAMORPHIC AND PLUTONIC BODIES RELATED(?) TO THE SAN EMIGDIO-TEHACHAPI TERRANE

Various dark rock masses in the study area have in common a possible relation to the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains; the main bodies are described below and are located in figure 4. Several smaller bodies that are not discussed and that were not all visited are shown in plate 1; they are particularly abundant in the Lone Tree Canyon drainage and to the southwest within a few kilometers of the Garlock fault. These small mafic masses, which are partly similar to the ovoid mafic inclusions that are abundant in the tonalite of Bear Valley Springs, are with much less confidence related to the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains. Nevertheless, the relative concentration of these small mafic bodies near the Cameron slivers suggests, at least, a similar origin.

LIVE OAK

The Live Oak mafic body, covering an area of about 6 km^2, was mapped as "gabbro and gabbro-diorite, including olivine norite" by Dibblee and Chesterman (1953). They noted gradational contacts with the surrounding tonalite but found the contact difficult to map. Near the east side of the Live Oak body, I mapped hypersthene-bearing tonalite that contains gneissic streaks and patches.

This mafic body is evidently igneous, at least in part, because it contains patches of strongly retrograde olivine norite. Nevertheless, the presence of fine- to medium-grained plagioclase amphibolite, hornblende-rich tonalite, and gneissic remnants in the surrounding tonalite suggests metamorphic lithologies similar to the major rock types of the San Emigdio-Tehachapi terrane. Particularly noteworthy is the tonalitic outcrop, about 3 km east of the Live Oak body, in roadcuts where the outcrop is

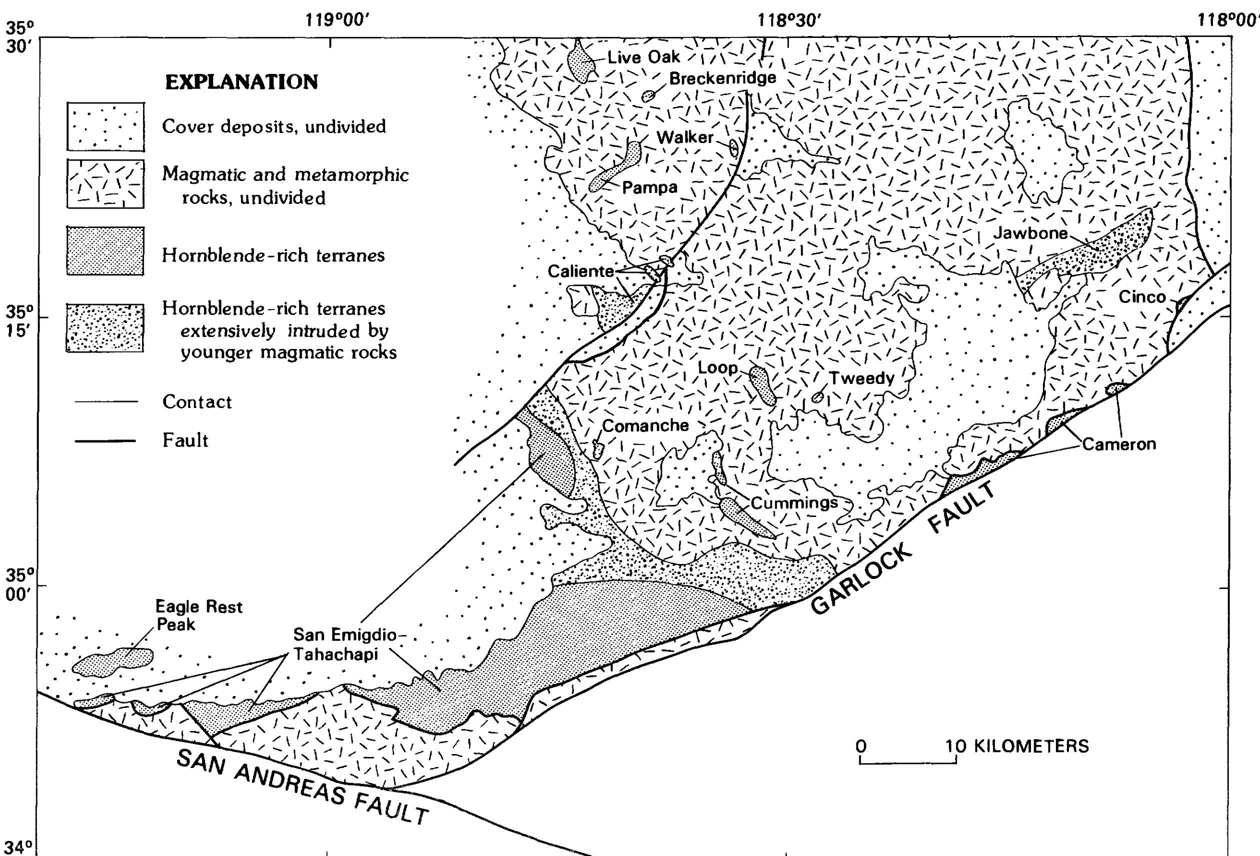


FIGURE 4.—Locations and names of hornblende-rich metamorphic and magmatic terranes in the southernmost Sierra Nevada.

at least 50 percent dark, elongate gneissic streaks and lenses. Just north of the Live Oak mafic body, a large, dark inclusion mass in the tonalite was found to be composed chiefly of pale-brown hornblende and labradorite. Most likely, this mass is a fragment of amphibolite from the Live Oak body.

BRECKENRIDGE

Small remnants of strongly retrograde olivine gabbronorite are exposed along Breckenridge Mountain Road, about 6 km east-southeast of the Live Oak body (Mount Adelaide 7½-minute quadrangle). The mafic rocks are remnants in an intrusion breccia of tonalite and fine-grained felsic granitic rock. The remnant mafic patches contain no dark gneiss, amphibolite, or any other rock type that would tie them directly to the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains. They do have a mineralogy and texture to suggest that they are correlative with part of the Live Oak body.

The mafic rocks have gabbroic to granoblastic (polygonal) textures and are dominated by well-twinned, fresh labradorite and colorless to pale-green amphibole. Studded throughout the samples are pleochroic (pink to green), lamellar, twinned orthopyroxene, as well as olivine that is altered along curving fractures. Some of the orthopyroxene appears to have schiller inclusions. Most olivine and orthopyroxene grains are surrounded by reaction rims of amphibole, with vermicular intergrowths of green spinel. In addition, scattered grains of green spinel and opaque minerals are present. The mineralogy and texture in general, and the conspicuous vermicular green spinel in particular, suggest a correlation with the Live Oak body.

Sams and others (1983) reported "layers of deformed mafic-ultramafic cumulates" in the San Emigdio-Tehachapi terrane. This observation strengthens the suggestion that both the Breckenridge and Live Oak bodies may be related to the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains.

PAMPA

A few kilometers south of the Live Oak body is a thin mafic body, about 6 km long, that was described as gabbro by Dibblee and Chesterman (1953). I examined this body only cursorily and only at its two ends. Particularly noteworthy near the northeast end of the body is a hypersthene-bearing tonalite composed of fresh, well-twinned plagioclase (approx An_{50}), abundant coarse-grained anhedral quartz, reddish-brown biotite, pleochroic hypersthene, and less common olive-brown hornblende. Most hypersthene has a sharp contact with plagioclase, with only rare, thin amphibole reaction rims. Associated with the tonalite is retrograde gabbro containing blocky, fresh subhedral plagioclase (approx An_{50}) and abundant pale-green to pale-olive-brown, generally acicular amphibole aggregates. Some amphibole cores contain skeletal clinopyroxene, and some acicular amphibole masses are pseudomorphs, probably of pyroxene. Large crystals of reddish-brown biotite, interstitial quartz, and aggregates of pale-green chlorite make up the rest of the rock. Also present is a fine-grained granoblastic rock composed of clinopyroxene, olive-brown hornblende, and plagioclase. Some hornblende aggregates, as much as 3 mm long, suggest pseudomorphs of phenocrysts and a possible volcanic protolith.

One of the samples from the southwestern part of the Pampa body is composed almost entirely of fibrous amphibole and chlorite, and could be a strongly retrograde ultramafic rock. More common are probable gabbroic rocks, in part retrograde, composed chiefly of labradorite and pale-green to pale-brown hornblende with included skeletal clinopyroxene. The physical resemblance of these rocks to some of those of the San Emigdio-Tehachapi mafic terrane is enough to raise suspicions that, at least in part, the Pampa rocks are metamorphic amphibolite rather than magmatic gabbro.

Most distinctive and, probably, most abundant in the southwestern part of the Pampa mafic body is hypersthene-hornblende tonalite. The rock has a decidedly granitic texture and contains 50 to 60 percent subhedral, well-twinned plagioclase (approx An_{50}), 5 to 10 percent quartz, 15 to 30 percent pale-brown hornblende, 10 to 20 percent hypersthene, less than 5 percent reddish-brown biotite, and about 1 percent opaque minerals. Some of the hypersthene sharply contacts plagioclase, but most has an amphibole reaction rim. One specimen contained both clinopyroxene and hypersthene, but bright-olive-green hornblende and dark-brown biotite—interference colors typical of normal granitic rocks (particularly tonalite) in the region.

WALKER

Dibblee and Chesterman (1953) mapped a small body of hornblendite and other ultrabasic rocks just west of the Walker Basin and noted the presence of hornblende-rich gneiss. I made a cursory investigation of the south end of this inclusion and collected and studied some highly retrograde ultramafic(?) rock that is composed of a decussate mat of pale-green to colorless, acicular to blocky amphibole and lesser pale-green chlorite with dull-gray interference colors. Local coarse-grained hornblende remnants that have hints of a pale-brown color contain well-aligned, bladed purple inclusions (schillerlike?). I also noted quartzite and biotite quartzofeldspathic gneiss. Float boulders at the foot of the slope just east of the Walker body and, presumably, derived from it are described in my field notes as "hornblende gabbro-diorite—hornblende-rich inclusion of dark gneissic complex." The Walker body thus is somewhat similar to the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains.

CALIENTE

South and west of Caliente, several square kilometers of mafic plutonic rocks are notably rich in hornblende. These Caliente rocks appear to be a large mafic enclave, much mixed with and intruded by younger granitic rocks.

The most common rock type is hornblende-rich gabbro to tonalite that consists chiefly of well-twinned andesine to labradorite and light- to moderate-olive-green hornblende. Much of the hornblende is in aggregates that are partly composed of pale-green acicular crystals. Skeletal clinopyroxene crystals form the cores of some hornblende. Strongly pleochroic, deep-brown to reddish-brown biotite is common in the tonalite but absent in the more mafic rocks. Some associated fine-grained amphibolite may be metavolcanic.

Several small bodies of ultramafic rock, enclosed in the Caliente rocks, are dominantly hornblende but locally contain altered remnants of olivine and orthopyroxene. In places, hornblende-rich gabbro intrudes the ultramafic rock and forms an intrusion breccia. This feature is well developed in a large roadcut on the north side of California Highway 58, about 4.5 km west of the junction with California Highway 223 to Arvin.

The most distinctive and widespread ultramafic rock is coarse, knobby weathering, and contains subhedral hornblende crystals, as long as 2 cm, set in a fine-grained groundmass of hornblende, clinopyroxene, and lesser labradorite. The hornblende is pale brown in thin section. Less common are patches of coarse-grained rock dominated by pale-green hornblende that contains abundant bladed opaque-mineral inclusions (schillerlike). These hornblende crystals enclose many small crystals of highly altered and fractured olivine. Less common are lamellar, twinned orthopyroxene and green spinel. These rocks may be retrograde remnants of peridotite(?) bodies now immersed in a gabbroic and tonalitic matrix.

Gneissic streaks, patches, and lenses are common in the Caliente terrane. Felsic gneiss, reminiscent of some of the gneiss in the San Emigdio-Tehachapi Mountains, is exposed in a roadcut on the Caliente-Bodfish road about 2 km west of Caliente. Dark, hornblende-rich gneiss that is even more reminiscent of the San Emigdio-Tehachapi terrane is exposed along California Highway 58 near the previously described intrusion breccia and on the grassy slopes to the southeast. The first report of these rocks was by Dibblee and Chesterman (1953, p. 32), who noted "* * * layers of banded hornblende-biotite gneiss made up of layers rich in hornblende alternating with layers containing less."

COMANCHE

A small mafic body, a few hundred meters across, is included in tonalite on the north side of Horsethief Flat on Comanche Point Road (fig. 2). I mapped the body as "hornblende gabbro" and noted considerable grain-size variation in the outcrop. Thin-section study revealed fresh, well-twinned labradorite and pale-green hornblende, in large crystals and in aggregates. One fine-grained sample has a granoblastic texture and is composed of a mat of cleanly twinned zoned labradorite, olive-brown hornblende, hypersthene, and abundant opaque-mineral grains. The hypersthene has clean, sharp boundaries with adjacent hornblende and plagioclase grains. Here is yet another example of a mafic rock that suggests a magmatic rock (cleanly twinned and zoned plagioclase), although the overall texture and mineralogy points to a high grade (granulite) metamorphic rock.

Along Comanche Point Road just west of Horsethief Flat (near the transmission line), the tonalite is darker and contains numerous dark inclusions and amphibole-rich clots. The one inclusion sample studied has a granoblastic to polygonal texture and is composed of labradorite, brown to pale-green hornblende, common opaque-mineral grains, and minor reddish-brown biotite. Strongly pleochroic hypersthene is common as cores of hornblende crystals. Generally colorless to pale-green amphibole forms a bleached zone between brown hornblende and hypersthene. The two different colored amphiboles are in optical continuity, however, and so this is not a standard reaction rim. Also, several hypersthene grains are in sharp, clean contact with brown hornblende. I suggest that these inclusions west of Horsethief Flat are analogous to the much larger inclusion swarm of the Loop area.

CUMMINGS

A thin, disrupted, arcuate belt of mafic rocks extends some 10 km eastward and southeastward of Cummings Valley. Petrographic notes for selected samples were presented by Ross (1983c) for this belt, which is not easily visited because of access restrictions. In the field, these rocks were variously referred to as diorite, hybrid rock, gabbro, and dirty, dark tonalite. It soon became apparent that hypersthene was a common characterizing accessory. A recent (1982) careful reexamination of petrographic notes and thin sections made it clear that there are two rather distinct, hypersthene-bearing rock types in this belt.

The smaller, northern body and the southern part of the larger, southern body are characterized by mafic, medium-grained granoblastic to gneissic rocks that are composed dominantly of sharply twinned and somewhat zoned labradorite and pale-brown to olive-brown hornblende. Reddish-brown biotite and quartz are present in small amounts in some samples. Strongly pleochroic hypersthene in fresh, clean anhedral crystals is common, and some samples also contain clinopyroxene. In part, the hypersthene has sharp, unaltered boundaries with hornblende and labradorite; but in other samples, clear to pale-green, somewhat acicular amphibole forms a cushioning reaction rim around the hypersthene. Almost certainly, these rocks are hypersthene granulites similar to other granulites of the San Emigdio-Tehachapi terrane, particularly in the Tunis Creek area.

In rather strong contrast, samples from the northern part of the larger, southern body consist of a hypautomorphic granular combination of andesine, abundant quartz (10–20 percent), and olive-green (granitic) hornblende in excess of brown to reddish-brown biotite. Accessory hypersthene (max 7 percent) generally has sharp, unaltered grain boundaries with adjacent minerals. The mineralogy and texture of these rocks indicate that they are magmatic tonalite.

LOOP INCLUSION SWARM

Ovoid mafic inclusions, as large as a few tens of centimeters in maximum dimension, are packed together in a tonalite matrix over an elliptical area (approx 1 by 4 km) in and southeast of the Tehachapi Loop on the Southern Pacific Railroad line (northwest of Tehachapi). As much as 75 to 90 percent of individual outcrops are composed of inclusion material. Although these inclusions have the appearance of typical hornblendic Sierran inclusions and are rarely foliated, the two inclusions that I selected as typical both contain hypersthene.

One sample consists of a granoblastic or polygonal mat of fresh twinned labradorite (55 percent), light- to moderate-olive-brown hornblende (35 percent), pink to green pleochroic hypersthene (9 percent), and metallic opaque-mineral grains (1 percent). The hypersthene seems to be in equilibrium because it forms clean, sharp contacts against hornblende and plagioclase, with no reaction zones. The other sample is somewhat coarser grained and xenomorphic granular or granoblastic. It likewise is dominated by fresh, well-twinned plagioclase of about An_{50} (50 percent) and olive-green hornblende (30 percent). Brown biotite (10 percent), quartz (5 percent), pleochroic hypersthene (5 percent), and scattered opaque-mineral grains are also present. Both of these inclusion samples are similar to the hypersthene granulite of the Tunis Creek area.

Two specimens in 4 km^2 are hardly an exhaustive sample, but they do suggest an affinity with the other mafic, metamorphic terranes of the region, particularly the San Emigdio-Tehachapi terrane. The Loop body may be a somewhat dismembered equivalent of the Cummings belt.

A hypersthene granulite sample from streaky inclusion and schlieren material in the uppermost part of Tejon Creek is nearly identical to the granulite samples from the Loop swarm. The matrix rock in Tejon Creek is a foliated hornblende-rich tonalite that is similar to the hornblende-rich rock which dominates the Caliente terrane.

TWEEDY

The Tweedy mafic body is engulfed by granodiorite about 5 km north of Tehachapi. It covers an area of only about ⅛ km^2, and in my short examination I could not determine its relation to the surrounding granitic rocks. In my field notes, I described the rock as "coarse knobby hornblende gabbro." Samples range in texture from fine to coarse grained and consist of 40 percent twinned, in part subhedral plagioclase (approx An_{50}), 40 percent pale-olive hornblende, 20 percent clinopyroxene, abundant coarse-grained sphene, and traces of quartz and K-feldspar. From the texture and mineralogy of the samples alone, this body could be either a plagioclase amphibolite or a gabbro. Just southwest of the mafic body are streaky gneissic layers in the granodiorite and coarse-grained red garnets, in crystals as large as 3 cm across. These gneissic rocks and the coarse-grained garnets are reminiscent of some amphibolitic gneiss of the San Emigdio-Tehachapi Mountains.

CAMERON

Various slices and slivers of mafic rocks are strung out along the Garlock fault from Cameron Canyon (approx 12 km southeast of Tehachapi) eastward for more than 20 km. These mafic slivers abruptly terminate on the south against the southern strand of the Garlock fault. Retreat of the north contact of the largest (westernmost) mafic sliver up canyons in the mountain front suggests a low-angle (thrust?) contact with granitic rocks to the north. The more easterly mafic slivers are intertwined with granitic rocks along anastomosing faults. Weathering, alteration, and shearing along the mountain front here greatly hinder determination of rock relations.

The westernmost fault sliver is relatively coherent and well exposed in several canyons, and mafic rocks are readily visible in parts of the eastern slivers. Dark amphibolite, in which pale-green amphibole aggregates with brown cores generally exceed plagioclase, is the most conspicuous rock type. These rocks contain abundant opaque grains. Interstitial quartz commonly is cataclastically deformed. Cataclasis and retrograde metamorphism to varying degrees is widespread in these rocks. Much of the amphibolite has a strong gneissic fabric.

Coarse-haloed red garnet is abundant and widespread. A spectacular exposure of garnets, as large as 10 cm, is in Waterfall Canyon, on the north side of California Highway 58, about 3 km east of Cameron Road. These garnets appear to be identical to those of the San Emigdio-Tehachapi terrane, but they are lower in almandine and higher in pyrope and grossularite (see subsection below entitled "Coarse-Haloed Red Garnet"). The presence of carbonate rocks in the Cameron slivers and their virtual absence in the San Emigdio-Tehachapi terrane may account for the compositional differences of the garnet.

Felsic gneiss, containing abundant quartz, plagioclase, and lesser K-feldspar, olive to reddish-brown biotite, and minor olive hornblende, is relatively abundant and commonly is strongly cataclastically deformed. Marble also is relatively common locally. In the easternmost slivers, the marble appears to be interlayered with garnetiferous amphibolite. Ted Antonioli (written commun., 1980) made a detailed study of the Cameron slivers and noted a dark, garnetiferous gneiss facies as well as a marble-bearing, felsic gneiss facies. The felsic gneiss that Antonioli noted is similar to the felsic (quartzofeldspathic) gneiss of the San Emigdio-Tehachapi mafic terrane. The dark, garnetiferous plagioclase amphibolite and gneiss of the Cameron slivers are virtually identical to much rock of the San Emigdio-Tehachapi mafic terrane.

In the course of his work, Antonioli mapped a small body of peridotite in the mafic terrane. Sams and others (1983) reported the presence of ultramafic rocks within the main San Emigdio-Tehachapi mafic terrane, and scraps of ultramafic rock are known from the possibly related Live Oak, Pampa, Caliente, and Eagle Rest Peak bodies. The Cameron peridotite thus helps tie this area to several of the other mafic-rock areas.

CINCO

A small body of mafic rocks is exposed at the base of the mountain front directly west of Cinco on California Highway 14. The mafic rocks are most likely a sliver in the Garlock fault zone, but the relations were unclear to me during a short visit to the outcrops. Samsell (1962) first noted these rocks and described several rock types, including hornblende-andesine gneiss.

In my short field visit, I noted diorite, dark- and light-colored gneiss, augen gneiss, fine-grained diorite (amphibolite) containing coarse-haloed red garnets, as large as 5 cm across, and some marble. The garnet-bearing amphibolite and the gneissic rocks at Cinco are strikingly reminiscent of some of the major rock types of the San Emigdio-Tehachapi terrane.

Petrographic examination revealed that what was called dark diorite in the field is amphibolite which is dominated by well-twinned plagioclase (approx An_{50}) and dark-green hornblende with distinctly brown cores that occurs in aggregates or in discrete crystals. Opaque grains are abundant, as are coarse red garnets. Minor quartz is present, and cataclasis is common. The gneissic rocks, composed of varying amounts of plagioclase, quartz, epidote, and abundant pale-green hornblende, are commonly cataclastic augen to flaser gneiss. Some of the green hornblende in the gneiss has brown cores. Strongly retrograde amphibolite and gneiss are present in this body, along with strongly seriticized plagioclase, and hornblende completely replaced by epidote and chlorite.

The Cinco and the previously described Cameron mafic rocks are both structurally isolated in the complex Garlock fault zone. I previously speculated (Ross, 1980) that these rocks were originally parts of the San Emigdio-Tehachapi mafic terrane and were separated from it by lateral movement on the Garlock fault. Although lithologic correlation with the San Emigdio-Tehachapi terrane is apparent, these mafic fragments may reflect uplift along the Garlock fault of parts of widely distributed mafic basement, rather than lateral fault transport of pieces of the more limited gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains.

JAWBONE

Along Jawbone Canyon and to the north in Hoffman Canyon (approx 8 km northwest of Cinco), mafic gneiss and amphibolite, covering an area of about 50 km^2, are much intruded by younger granitic rocks. Samsell (1962) first noted the presence of abundant schistose and gneissose xenoliths in this area. Early in my reconnaissance studies of the southernmost Sierra Nevada in 1978, I was surprised by the presence of these anomalously dark rocks well east of the quartz diorite line of Moore (1959), in an area where the metamorphic rocks are generally marble-rich metasedimentary pendants and inclusions. At that time, I noted amphibolite, hornblende-rich diorite, and gneiss, and suggested that these dark rocks were correlative with the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains.

Recent (1982) reexamination of thin sections from the Jawbone body reinforced my earlier suspicions of similarity to the San Emigdio-Tehachapi terrane, but my overall impression is that the Hoffman rocks may be of somewhat lower metamorphic grade. Most common in the Hoffman mafic rocks is amphibolite (in part quartz bearing) composed of well-twinned andesine-labradorite, moderate-green to olive hornblende, and abundant opaque-mineral grains. These rocks grade to gneissic rocks of similar mineralogy. Also present is quartzofeldspathic gneiss composed of andesine, brown biotite, and quartz.

Biotite is present, but much less common than hornblende, in the amphibolite and dark gneiss and is brown to reddish brown. Clinopyroxene is present, but scattered, in the dark rocks. Only one occurrence of coarse-haloed red garnets was observed near the west end of the Jawbone mafic mass.

The overall appearance and general mineral content of these dark rocks are strongly compatible with those of the San Emigdio-Tehachapi terrane. In the Jawbone mass, however, coarse-haloed garnet is rare, orthopyroxene is apparently absent, and clinopyroxene is scattered. Hornblende is green (rarely brown), and biotite is mostly brown to reddish brown. Plagioclase may be less calcic than in the San Emigdio-Tehachapi rocks.

The Jawbone mafic mass (fig. 4) is isolated from the other mafic bodies and is immersed in a granite-granodiorite terrane whose initial $^{87}Sr/^{86}Sr$ ratios of 0.707 to 0.708 (Kistler and Peterman, 1978) indicate intrusion through Precambrian continental basement. Thus, the Jawbone mafic rocks, with oceanic(?) affinities, are out of place and seem to call for lateral and (or) vertical displacement from kindred mafic rocks of the region. Large masses of mafic rock in a similar(?) setting were mapped as "Mesozoic basic intrusive rocks" by Smith (1965). The same rocks, some 40 to 60 km north of the Hoffman body, are the Summit gabbro of Miller and Webb (1940). These rocks need to be investigated further to determine whether they have mafic metamorphic affinities and bear any similarity to the Jawbone mafic rocks.

EAGLE REST PEAK

About 26 km^2 of gabbro, pyroxenite, tonalite, and metavolcanic rocks crop out north of the west tip of the main basement exposures in the San Emigdio Mountains. The rocks were first described by Hammond (1958); more recently, they were reexamined and sampled by Ross (1970), who referred to this locality as the Eagle Rest Peak area.

About half the outcrop area is made up of gabbro and pyroxenite. The gabbro, by far the most abundant rock type, is in part anorthositic gabbro, consisting of calcic plagioclase (labradorite-bytownite), pale-green amphibole, clinopyroxene, minor orthopyroxene, and metallic opaque minerals. Layering of some of the gabbro suggests a cumulate texture. The pyroxenite, which is composed primarily of clinopyroxene and orthopyroxene, is partially serpentinized.

Presumably intrusive into the gabbro and pyroxenite is a small body of hornblende-quartz gabbro to tonalite. These rocks are about one-half plagioclase (andesine to labradorite), about one-third green hornblende, and about one-fourth quartz; they also contain minor biotite and metallic opaque minerals.

The wallrock of these gabbroic rocks is a fine- to medium-grained, locally coarse-grained, mafic metavolcanic-rock type with local diabasic texture. This wallrock is dominantly plagioclase, as calcic as labradorite and pale-green hornblende; it is now essentially an amphibolite.

It is problematic whether the Eagle Rest Peak mafic rocks are related to the rest of the exposed mafic rocks in the southernmost Sierra Nevada. Some dark amphibolites in the San Emigdio-Tehachapi terrane resemble some of the rocks at Eagle Rest Peak, but common dark amphibolite is not a distinctive rock for correlation. I have seen no counterparts of either the coarse-grained anorthositic gabbro or the hornblende-quartz gabbro in other basement-rock outcrops of the southernmost Sierra Nevada. The Eagle Rest Peak body may be a fragment of either the Kings River ophiolite (Saleeby, 1978) or the Coast Range ophiolite (Bailey and Blake, 1974), and structurally separate from the basement rocks of the Sierra Nevada tail.

Reitz (1983) suggested that the absence of sheeted dikes and peridotite tectonite in the Eagle Rest Peak area makes it more likely that these rocks are part of an arc-type plutonic complex rather than part of an ophiolite, as I originally suggested. Reitz furthermore noted that the plutonic association and metamorphic overprint of these rocks more closely ally them to the western Sierra Nevada foothills belt than to the Coast Range ophiolite.

FELSIC GNEISS BODIES RELATED(?) TO THE SAN EMIGDIO-TEHACHAPI TERRANE

The north-south-trending belt of metasedimentary rocks at the longitude of Tehachapi contains three elongate bodies that closely resemble some of the quartzofeldspathic gneiss of the San Emigdio-Tehachapi terrane. All three bodies are strongly cataclastically deformed. The northern two bodies (Brown Meadow and Tweedy Creek) appear to be derived, at least in part, from relatively coarse grained, felsic granitic rock. The southern body (Mountain Park) contains some metasedimentary rocks and is more or less transitional between the metasedimentary rocks of the Keene area and the quartzofeldspathic gneiss of the San Emigdio-Tehachapi terrane. These three cataclastically deformed bodies appear to mark a strongly compressional zone of deformation at or near the boundary between largely tonalite to the west and granite to the east the (quartz diorite line of Moore, 1959). The possible significance of this zone is considered below in the subsection entitled "Structural Elements * * *."

MOUNTAIN PARK

The lower mountain slopes 6 km south of Tehachapi expose about 6 km^2 of cataclastic felsic gneiss in a northwest-trending belt. This gneiss is composed mostly of plagioclase, quartz, reddish-brown biotite, and lesser garnet. Layers of impure to nearly pure quartzite are present; one layer contains abundant sillimanite. Other layers are of mica schist containing abundant red-brown biotite, and muscovite. Marble lenses are also present.

The larger (gneissic), Mountain Park mass generally resembles the felsic gneiss of the San Emigdio-Tehachapi terrane but has some affinities (sillimanite, mica schist, marble) with the metasedimentary rocks to the north and east. The smaller eastern body is even more enigmatic. It contains significant amounts of marble and calc-hornfels, as well as quartzite, but also amphibolite composed of plagioclase (approx An_{50}) and olive-brown hornblende, liberally sprinkled with opaque-mineral grains. In this rock are also coarse, poikilitic red garnets with bleached haloes—in short, a rock typical of the San Emigdio-Tehachapi terrane to the west.

It is unclear to me why there is an apparent association of gneiss and garnetiferous amphibolite with marble-bearing metasedimentary layers here in the Mountain Park bodies. The gneiss and amphibolite may be inclusions in the tonalite that surrounds the Mountain Park bodies, and the whole tonalite package may intrude the marble-bearing metasedimentary rocks. This interpretation is possible, but considering the abundance of marble-bearing metasedimentary layers on all sides of the two Mountain Park pendants, it seems like special pleading.

The Mountain Park rocks lie along a transition zone between largely marble-free metamorphic rocks to the west and a relatively marble-rich metamorphic terrane to the northeast. Marble associated with felsic gneiss and garnetiferous amphibolite to the east in slivers in the Garlock fault (Tehachapi Pass area) might be analogous to the Mountain Park rocks. In addition to the lithologic change, this may be a transitional zone between the higher grade (granulite) rocks to the west and the somewhat lower grade sillimanite-bearing, metamorphic rocks to the northeast.

TWEEDY CREEK

Along Tweedy Creek, about 12 km north-northwest of Tehachapi (lat 35°13′ N., long 118°29′ W.), a body of augen gneiss was traversed that on preliminary field examination was considered to be part of the Keene metasedimentary terrane. Subsequent thin-section study suggested that this gneiss is a strongly deformed felsic granitic rock (see fig. 49*F*) which is correlative with the felsic rocks associated with the tonalite of Bear Valley Springs. However, a preliminary Pb-U age on zircon from this gneiss is 113 m.y. (Sams and others, 1983), coeval with quartzofeldspathic gneiss in the Sierran tail and somewhat older than the nearby tonalite of Bear Valley Springs, from which zircons have been dated at about 100 m.y. (Sams and others, 1983). The strongly mylonitized rocks of Tweedy Creek (see fig. 49*F*) may be derived from felsic granitic rock that is a common associate of the quartzofeldspathic gneiss of the San Emigdio-Tehachapi terrane. The zircon age suggests that these Tweedy Creek rocks are a scrap of the gneissic and amphibolitic terrane of the San Emigdio-Tehachapi Mountains. Note that the Tweedy Creek rocks are essentially on strike with the cataclastically deformed, felsic gneissic rocks of Mountain Park.

BROWN MEADOW

A belt of strongly cataclastically deformed rocks, about 12 km east of the center of the Walker Basin, extends southward some 15 km from Brown Meadow (approx 1.5 km southeast of Brown Peak) along the east margin of the Mount Adelaide mass. In the field, these rocks were first interpreted as a protoclastic margin of the Mount Adelaide body. Subsequent petrographic study showed the augen gneiss and mylonite of the deformed belt to be rich in K-feldspar, whereas the Mount Adelaide body near the deformed belt contains only minor amounts of K-feldspar. Therefore, it seems unlikely that the deformed rocks are related to the Mount Adelaide body.

The augen gneiss with conspicuous K-feldspar augen and the mylonitic rocks of the Brown Meadow belt do, however, strikingly resemble the rocks of Tweedy Creek to the south. There thus appears to be a discontinuous belt of strongly deformed rocks, extending from Brown Meadow south to the Mountain Park area. Scattered amphibolite (metagabbro?) in the Brown Meadow belt that contains pale-green to brown hornblende is similar to some of the hornblende-rich rocks of the San Emigdio-Tehachapi terrane. This feature helps link the Brown Meadow belt to the Tweedy Creek and Mountain Park bodies and suggests that they all may be related to the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains.

METAVOLCANIC ROCKS OF FRENCH GULCH

SETTING

At the north edge of the study area, a small belt of dark-gray metavolcanic rocks is exposed over an area of about 10 km^2. These rocks appear to contact the metasedimentary rocks of the Keene area on the west, but the relations are unknown. These metavolcanic rocks were shown by Saleeby and others (1978) to extend some 16 km farther northward to the vicinity of Isabella Lake, and reconnaissance studies to the north suggest to me that this metavolcanic belt may extend even farther.

ROCK TYPES AND PETROGRAPHY

The metavolcanic rocks of French Gulch make up a somewhat monotonous section that is generally dark gray, fine grained, and weakly to strongly foliated. Some marble interbeds occur in the section, along with some dark, dense layers that probably were originally chert. Relict megacrysts of plagioclase, as much as 2 mm long, are scattered through some samples. Some megacrysts are euhedral and could be phenocrysts, whereas others are angular and could be broken phenocrysts or clasts from a tuff. The occurrence of some composite megacrysts suggests that at least part of the metavolcanic section is made up of crystal tuff. The interlayers of chert and marble would accord well with a water-laid tuff origin for at least part of the volcanic section. Some of the samples are strongly sheared, and their megacrysts are deformed into augen. A volcanic parentage has been assumed for these rocks, although structural contortions have obliterated almost all traces of original fabric.

Although many of the metavolcanic layers are quite dark, the original compositions were intermediate to felsic, judging from the present mineral assemblage. Quartz, andesine, and K-feldspar in varying proportions dominate, along with abundant red to brown biotite. In addition, some layers contain abundant hornblende and epidote. Small amounts of tourmaline are present in most samples. The metavolcanic section was probably mostly dacitic with variations to andesitic, and it may have contained some rhyolitic layers.

METAMORPHIC ROCKS IN THE SUBSURFACE OF THE SOUTHEASTERN SAN JOAQUIN VALLEY

Core material from nearly 100 oil wells that were drilled to basement in the southern San Joaquin Valley south of lat 35°30′ N. was collected, thin-sectioned, and studied by May and Hewitt (1948). Although the whereabouts of the core material is presently unknown, the thin sections, now in the custody of the California Academy of Sciences, were kindly loaned to me for restudy. The following discussion is based on the data gleaned from these tiny thin sections, only 1 to 2 cm^2 in area. The locations of the wells from which the thin-section material was obtained and other data about the wells and the rocks were summarized by Ross (1979).

METAIGNEOUS ROCKS

In three areas (pl. 1) that encompass an area of about 230 km^2, the basement-rock subcrop is largely greenschist and amphibolite. The greenschist samples are dark, dense, schistose rocks dominated by chlorite, epidote-clinozoisite, and actinolitic amphibole. Plagioclase, generally sodic (in part albite), is much less abundant in most sections. Minor amounts of quartz, biotite, and muscovite are found in some sections. Metallic opaque minerals are common locally. Particularly abundant in some samples are sugary grains and aggregates of sphene; the metallic opaque-mineral and sphene grains probably were derived from ferromagnesian minerals that have been subjected to retrograde metamorphism. Some of the greenschist samples preserve a felty to trachytic texture. From the composition and appearance of these rocks, they are almost certainly metavolcanic.

Somewhat higher grade rocks are also preserved in this suite. Some hornblende-andesine rocks in part preserve a porphyritic texture, but even these rocks generally contain some chlorite and epidote-clinozoisite, as well as actinolitic aggregates that mimic the original hornblende or pyroxene crystals.

Most of these subsurface rocks were originally quite mafic, but one section shows conspicuous square β-quartz phenocrysts, as much as 1.5 mm long, smaller plagioclase phenocrysts (now albite), and clots of fine-grained muscovite set in a hornfelsed groundmass of plagioclase, quartz, and muscovite—probably a quartz porphyry. This sample, however, is north of and isolated from the three large greenschist areas. Some thin sections have a sandy texture suggesting tuff or tuffaceous sedimentary rocks.

In summary, most of the metaigneous rocks that subcrop in the San Joaquin Valley are classic greenschist that partly preserves a volcanic texture (felty or trachytic). Most probably their protolith was basalt or, possibly, andesite. Some of the amphibolites were probably intrusive rocks, but they are similar in composition to the greenschist and are certainly related in origin.

METASEDIMENTARY ROCKS

East of the southernmost mass of greenschist and amphibolite is an elongate belt of metasedimentary rocks that is interrupted by intrusive granitic rocks (pl. 1). These rocks are dominantly schistose, and texturally they resemble the Pampa Schist of Dibblee and Chesterman (1953).

The dominant rock type is a dark, strongly schistose rock containing varying amounts of quartz, biotite, muscovite, and chlorite; plagioclase is minor. Opaque carbonaceous dust is common, and garnet and tourmaline occur locally. This carbonaceous schist is similar to some rocks of the Pampa Schist. No andalusite was seen in the subsurface samples, but I note that large areas of the Pampa Schist also lack andalusite.

Impure micaceous quartzite, marble containing grains of quartz and plagioclase, muscovite-quartz schist (clay shale), and calcareous metasiltstone containing epidote are represented by one thin section each. The relation of these rocks to the carbonaceous schist and the outcrops of the Pampa Schist is uncertain; calcareous rocks, for example, are unknown in outcrops of the Pampa Schist.

RELATION TO THE PAMPA SCHIST

Both the metaigneous and metasedimentary rocks of the San Joaquin Valley are dominantly dark and strongly schistose, and although these rocks are generally of a lower metamorphic grade than the exposures of the Pampa Schist, there is certainly overlap in grade between the two areas. The Pampa Schist appears to be dominantly metasedimentary, whereas the San Joaquin Valley subcrop is dominantly metaigneous, but some rock types are common to the two areas. The overall appearance and composition suggest that all these rocks are closely related.

Across the White Wolf fault to the southeast, the metamorphic-rock outcrops are gneissic and amphibolitic, and the metamorphic grade is generally higher. The small area about 15 km south-southeast of Arvin (pl. 1) that is lumped with the greenschist and amphibolite unit for convenience is based on one thin section that features subangular clasts of quartz set in a groundmass of chlorite and lesser olive to green biotite. This low-grade metasedimentary or metatuffaceous rock is near outcrops of gneissic rock.

BEAN CANYON FORMATION

SETTING

The metamorphic rocks on the south fringe of the Tehachapi Mountains, south of the Garlock fault, were originally assigned to the Bean Canyon Series by Simpson (1934). Dibblee (1963, 1967) renamed these rocks the "Bean Canyon Formation," but for some unexplained reason he excluded the large body of dominantly marble, at the west end of the belt of metamorphic rocks south of the Garlock fault (pl. 1), from the unit. My work suggests that the whole metamorphic belt south of the Garlock fault is part of the Bean Canyon Formation. Therefore, in this report, all these rocks are included within the Bean Canyon Formation, but I note that various rock types are present and that the Bean Canyon Formation can be subdivided, at least locally, even though the dismemberment of this section by granitic intrusive rocks makes the task difficult.

The Bean Canyon extends discontinuously for about 50 km from a point near Castaic Lake to Oak Creek Road. It is reasonable to assume that the belt is cut off to the west by the San Andreas fault, even though the westernmost exposures are truncated by granitic rocks about 5 km from the fault zone. The Bean Canyon Formation is terminated on the east by overlapping Quaternary alluvial deposits. The trend of layering (at least in part, bedding) generally parallels the northeastward trend of the entire metamorphic belt; dips are generally steep, but in the westernmost body is much folding and some gentle dips. The two largest bodies on the west may be relatively flat bottomed, judging by the relation of their contacts to the granitic rocks and the topography. Wiese (1950) suggested that these granitic rocks may have intruded along a pregranite thrust surface. The contacts are definitely intrusive, however, and contact-metamorphic-mineral deposits are developed locally along them. The easternmost exposures of the Bean Canyon Formation appear to be normal roof pendants, keeled down into the granitic rocks. Although structural details of the metamorphic rocks remain to be worked out, G.A. Davis (written commun., 1977) suggested that the metamorphic rocks in Bean Canyon are isoclinally folded into a synformal(?) structure.

ROCK TYPES AND PETROGRAPHY

White to gray marble that commonly is coarsely crystalline is probably the most abundant rock type in the Bean Canyon Formation. Possibly the most distinctive rock type is gray to green, dense, in part thinly layered calc-hornfels. The porcelaneous appearance of fresh surfaces typifies these rocks. Metavolcanic rocks and dark andalusite hornfels occur locally. Individual bodies of the metamorphic rocks from west to east are described below.

Dark, dense hornfels occurs on the north side of the westernmost, largest, relatively flat bottomed metamorphic mass. The hornfels is quartz rich and contains varying amounts of clinopyroxene, epidote-clinozoisite, and K-feldspar; locally, hornblende and scapolite are present. One specimen has a notable volcanic texture; equant plagioclase crystals, as much as 2 mm long, are set in an acicular mat of olive hornblende and minor plagioclase that is liberally sprinkled with sphene and some pyrite. The relation of this volcanic rock to the rest of the hornfels is uncertain. The resistant hornfels probably is preferentially exposed at the expense of such rocks as mica schist, which I found to be very rare here, but which were noted in the area by Crowell (1952). The westernmost metamorphic body probably was originally composed of limestone, sandstone, and highly siliceous siltstone. Into these rocks, andesitic(?) flows or intrusive rocks were introduced. The marble here and elsewhere in the Bean Canyon Formation is largely white to gray calcite marble. Graphite is locally abundant, and the marble is in part dolomitized. Bedding is partly preserved, but in much of the coarsely crystalline marble all traces of original structures have been obliterated. Some areas are shown on the map (pl. 1) as marble where, in fact, they contain significant amounts of other rock types. For example, within the area mapped as marble in the large western metamorphic body, layers of green to dark-gray calc-hornfels are present that are composed of plagioclase, hornblende, clinopyroxene, epidote in varying proportions, and lesser amounts of K-feldspar, scapolite, and garnet.

The next mass of the Bean Canyon Formation, to the east near the Quinn Ranch, also is relatively flat bottomed; it is dominantly composed of marble but contains two belts of other lithologies. These two belts consist of gray to green, dense calc-hornfels and fine-grained quartzite, and minor amounts of pale-green amphibole, biotite, plagioclase, and clinopyroxene. These rocks probably were originally quartz sandstone with minor amounts of calcareous and argillaceous cement. They grade into schistose rocks that are rich in muscovite and biotite and locally contain andalusite; garnet is a sparse accessory. The schistose rocks may have originally been siltstone and argillite. No recognizable metavolcanic rocks were found in this pendant, but one layer of plagioclase-hornblende schist contains pale-green hornblende crystals, as much as 3 mm long, that may be remnant crystals or fragments from a tuff or tuffaceous sedimentary rocks.

Between the Quinn Ranch and Cottonwood Creek are numerous small pendants composed chiefly of marble. A thin section from a dark layer of thinly laminated rocks in marble in an abandoned quarry just east of Canyon del Gato-Montes may contain relict silty grains. The laminated rocks are composed of alternating quartz-rich and quartz-poor layers. The quartz-poor layers contain varying proportions of brown biotite, K-feldspar, and plagioclase. Minor andalusite is present in some of the biotite-rich layers. Rounded detrital grains of apatite, zircon, garnet(?), and opaque material testify to the clastic origin of these rocks.

From Cottonwood Creek to the east limit of exposures of the Bean Canyon Formation, marble, though still abundant, no longer dominates the section. Well exposed west of Cottonwood Creek is a belt, nearly a kilometer wide, of dark schist, impure quartzite, calc-hornfels, and minor pure quartzite. Commonly the dark schist contains abundant andalusite (some with well-formed chiastolite crosses), abundant brown to red-brown biotite, muscovite, and some K-feldspar. Quartz is abundant in some layers, and hints of original silty texture are present. These rocks grade by degrees into impure quartzite with micaceous layers and into rather pure quartzite with scattered mica and amphibole that reflect original argillaceous and calcareous cement. Calc-hornfels rich in clinopyroxene, quartz, and plagioclase is also present. Immediately north of the schist belt are scattered exposures of dark amphibolite that suggest a metavolcanic parentage. These rocks are massive to weakly layered and are rich in fibrous green amphibole and epidote. Largely untwinned plagioclase crystals, as much as 2 mm long, are set in this dark groundmass; locally they are in polygonal aggregates, with fibrous amphibole molded around them. The rocks do not resemble the gneissic and amphibolitic mafic rocks north of the Garlock fault but do resemble metabasalt or metaandesite related to the Bean Canyon Formation.

Between Cottonwood Creek and Gamble Spring Canyon, Dibblee (1967) delineated several small bodies that he called hornblende diorite. Most of the samples that I have studied from those bodies are dark granodiorite to tonalite and probably represent the granodiorite of Gato-Montes contaminated by interaction with metamorphic rocks of the Bean Canyon Formation. Locally unusual gabbroic rocks were found that are probably related to the metavolcanic rocks of the Bean Canyon Formation. One sample from just east of Cottonwood Creek features stubby euhedral crystals of weakly zoned pale-brown to colorless hornblende(?) set in a matrix of clinopyroxene, labradorite, K-feldspar, and minor quartz.

The east-west-trending elongate pendant of the Tylerhorse-Gamble Spring Canyon area has a conspicuous belt of marble along its south side that contains some dark hornfels layers. One dense, dark layer is dominantly quartz but also is rich in K-feldspar. It is liberally sprinkled with opaque carbonaceous(?) matter, as well as tremolite and pale-reddish-brown tourmaline. North of the marble is a section of dark rocks much like those exposed west of Cottonwood Creek: dark micaceous schist, in part containing andulusite; dense, dark siliceous calc-hornfels; and lesser marble and light quartzite. Particularly noteworthy in this layer are porphyritic volcanic rocks or crystal tuff. These rocks in various shades of gray have phenocrysts (or crystal fragments) of twinned and zoned plagioclase, embayed quartz, and biotite aggregates, as much as 4 mm long, in a dense granoblastic matrix of quartz, plagioclase, biotite, and pale-green hornblende. Also present are layers with a well-preserved felty texture of small tabular plagioclase crystals. These rocks contain small phenocrysts of plagioclase and altered hornblende, as much as 2 mm long, set in a groundmass rich in acicular, pale-green hornblende crystals. The mineral content suggests that these volcanic rocks were originally dacite to basalt.

Bean Canyon was designated the type section of the Bean Canyon Formation (Dibblee, 1963, 1967). Here is well exposed, in a rather limited area, the lithologic range of the formation. Dense, dark siliceous calc-hornfels, micaceous schist (in part containing andalusite), impure quartzite, a thick belt of pure quartzite, and marble make up most of the section. A dark belt of altered peridotite(?) that extends across the canyon is highly distinctive. It is made up of fractured, altered olivine laced by colorless to pale-yellowish-green serpentine minerals; abundant acicular, colorless amphibole accompanies the olivine. North of the ultramafic belt is a gray porphyritic metavolcanic layer containing ovoid to equant phenocrysts of plagioclase and quartz set in a granoblastic mat of untwinned plagioclase, quartz, brown biotite, and minor muscovite. Accessory tourmaline is found in this layer and in several other metavolcanic-rock types in the Bean Canyon Formation. This layer somewhat resembles the metavolcanic rocks in Gamble Spring Canyon, but the phenocrysts here are less distinct.

Dunne and others (1975), in a study of the strata in and near Bean Canyon, suggested that the protoliths of the Bean Canyon Formation were about equal amounts of feldspathic and quartz wacke, quartz arenite and siltstone, argillite and shale, and marble; volcanic rocks were subordinate in the original section.

Various scraps, slivers, and pendants of the Bean Canyon Formation are also exposed in the area between the Tehachapi-Willow Springs and Oak Creek Roads. Although marble probably dominates in this area, the bodies mapped as marble (pl. 1) generally include other rock types. For example, the marble belt that is cut by Willow Springs-Tehachapi Road has a dark hornfels layer that upon thin-section study proves to be a hornfelsed metavolcanic rock in which somewhat embayed plagioclase crystals, as long as 3 mm, are set in a dense matrix of quartz, feldspar, biotite, and minor hornblende. In addition to the plagioclase megacrysts, clusters of fine-grained brown biotite, gray-green hornblende, opaque material, and sphene are present that probably mimic original ferromagnesian megacrysts.

On Oak Creek Road, about 1,500 m east of its junction with Willow Springs-Tehachapi Road, gray metavolcanic rocks are exposed with well-preserved volcanic textures. Hand specimens resemble porphyry, but thin-section study suggests that these rocks are pyroclastic. Plagioclase megacrysts, as much as 6 mm but generally from 2 to 3 mm long, are conspicuous and generally are partially hornfelsed. Smaller, rounded to equant quartz masses are suggestive of clasts, as are some broken-looking plagioclase crystals. These rocks have a dense hornfelsed matrix consisting of quartz, feldspar, brown biotite, and minor green hornblende. The other common rock types composing the Bean Canyon Formation are also seen on Oak Creek Road: micaceous hornfels, impure quartzite, quartzite, and quartz-plagioclase-pyroxene calc-hornfels.

About 800 m east of the Willow Springs-Tehachapi Junction with Oak Creek Road, a layer of gray dense hornfels was found that is studded with euhedral, tabular, grass-green crystals of clinopyroxene, as much as 1 cm long. This highly distinctive rock, with its groundmass of quartz, K-feldspar, minor plagioclase, accessory round sphene grains, and abundant graphite, has not been seen elsewhere in the outcrop area of the Bean Canyon Formation.

AGE

Dunne and others (1975) noted "* * * the presence of poorly preserved macrofossils" in the Bean Canyon Formation, but the fossils were not diagnostic for age determination. They suggested, however, "* * * that two lines of evidence support an early Mesozoic age for the Bean Canyon Formation. First, the lithostratigraphy of the formation is similar to that of Triassic rocks in the Mineral King pendant of the southern Sierra Nevada but not similar to that of Paleozoic formations in the region. Second, the inferred environment of deposition of the Bean Canyon Formation is compatible with regional early Mesozoic paleogeographic data." Dunne and others postulated that during the early Mesozoic, a nascent volcanic arc lay to the east and a subduction zone to the west, and that the Bean Canyon Formation represents an arc-trench gap deposit between those two features.

R.A. Fleck (written commun., 1976) analyzed dacitic metavolcanic rocks from Bean Canyon by the Rb/Sr method and determined ages averaging 150 m.y. The data are equivocal, however, and can be interpreted as providing either the age of the volcanism or the time of metamorphic homogenization of the Sr-isotopic systems of older volcanic rocks. Thus, the metavolcanic rocks are at least 150 m.y. old, possibly much older. The presence of numerous Upper Cretaceous to mid-Cretaceous plutons and the absence of Jurassic plutons (except in the Eagle Rest Peak area) in the southernmost Sierra Nevada makes it seem unlikely that the Rb-Sr age reflects a time of metamorphic homogenization. The Rb-Sr age of about 150 m.y. most probably reflects the age of the dacitic volcanism. Therefore, in the present report, the age of the Bean Canyon Formation is considered to be Jurassic.

PAMPA SCHIST OF DIBBLEE AND CHESTERMAN (1953)

SETTING

The predominantly dark schist west of the Walker Basin was named the "Pampa Schist" by Dibblee and Chesterman (1953). The schist forms an east-westerly-trending, sinuous belt that is intimately penetrated by granitic rocks. The schist layering ranges from vertical to steeply south dipping, and there are no obvious clues to original bedding or top directions in the section. The Pampa Schist differs from the other metamorphic rocks of the region. Rocks possibly related to the Pampa Schist are penetrated in oil wells drilled to basement over a rather extensive area in the adjacent San Joaquin Valley (Ross, 1979). Apparent limitation of the Pampa Schist and its possible San Joaquin Valley subcrop extension to the area northwest of the White Wolf-Breckenridge fault suggests that this structural zone marks a significant break in the metamorphic framework rocks.

ROCK TYPES AND PETROGRAPHY

The Pampa Schist is characterized by dark, notably micaceous rocks containing local conspicuous andalusite (chiastolite) crystals, some rocks with metavolcanic affinities, and an absence of marble and quartzite. Some quartzofeldspathic layers are indistinguishable from layers in the Keene and Salt Creek metamorphic units, but overall the Pampa Schist is distinctive and easily separable from other metamorphic units of the southernmost Sierra Nevada.

Probably the most distinctive feature of the Pampa Schist is the dominance of muscovite and biotite in many layers. Quartz and plagioclase are present in most layers, but they are subordinate. The rocks are dominantly a pelitic schist. Andalusite, partly in coarse crystals, is locally abundant in the dark schist. Sillimanite, in part after andalusite(?) is widespread in needlelike bundles and as clusters of acicular to prismatic crystals that are generally small but locally are as coarse as 1 cm across by 5 cm long. Powdery carbonaceous matter is common and contributes to the dark color of these rocks. In some layers, muscovite occurs in dense clots that are reminiscent of shimmer aggregates. Some thin sections contain small amounts of orange, nearly isotropic material that has a serpentinelike structure which may represent pinite alteration of cordierite, but no original cordierite was noted. Pale-green chlorite is present in some layers, but it is by no means common. Tourmaline and garnet are local accessories in the schist. I conclude that the protolith of much of the Pampa Schist was highly argillaceous.

Amphibolite is a less common but recurring rock type. Some specimens have original phenocrysts mimicked by clots of hornblende and opaque-mineral grains. Although pale-green acicular amphibole is present in some specimens, the association with andesine and (locally) with clinopyroxene suggests that these rocks, though now extensively retrograde, originally reached amphibolite grade. The easternmost exposures of the Pampa Schist include some unusual rocks that I would characterize as retrograde gabbro. Some layers are rich in antigorite(?), tremolite, and muscovite; others are rich in clinopyroxene and chlorite. Associated samples, composed chiefly of olive hornblende and labradorite with accessory metallic opaque-mineral grains and green spinel, show compositional layering and a granoblastic to polygonal texture. In this area, Dibblee and Chesterman (1953) noted a massive chlorite schist that they suggested was "* * * apparently of volcanic origin, probably a basalt." One of the samples from the southwesternmost exposures of the Pampa Schist has a fairly well preserved diabasic texture; some subhedral andesine crystals, as long as 1.5 mm, are set in a somewhat felted groundmass of plagioclase, biotite, and muscovite.

I have probably oversampled the micaceous and andalusite-bearing rocks relative to the sugary quartzofeldspathic rocks, which are abundant particularly in the northern exposures and in some of the smaller slivers on Breckenridge Mountain Road. Nevertheless, my overall impression of these rocks is that the original section was dominated by dark shale and silty rocks, with an admixture of volcanic rocks. Some of the dark argillaceous rocks might have been tuffaceous, although there is little direct evidence from the collected samples.

RAND SCHIST

SETTING AND AGE

Two horses of dark schist along the Garlock fault differ from other metamorphic rocks in the southernmost Sierra Nevada. The larger horse extends 34 km northeastward from Bear Trap Canyon and is as much as 2.5 km thick. The much smaller horse, exposed about 10 km northwest of Mojave, is only about 5 km long and no thicker than 0.5 km. It had largely escaped notice until this study, although Dibblee (1959) noted "gneiss and biotite schist" here on his map explanation. The larger horse has long been recognized (Wiese, 1950) and has been considered correlative with the Rand Schist (Hulin, 1925). The Rand Schist is now considered to be part of the widespread, informally named Pelona-Orocopia Schist (Ehlig, 1968, p. 294). Haxel and Dillon (1978, p. 453) indicated that the "* * * metamorphism of the Pelona-Orocopia Schist occurred in Paleocene (or Late Cretaceous) time." The age of the Pelona-Orocopia protolith is unknown, most probably Mesozoic. In the present report, the age of the Rand Schist is thus tentatively considered to be Mesozoic(?).

The rock types in the smaller horse are comparable to those in the larger horse, as well as to those in the Rand Schist in the Rand Mountains. These three masses are almost certainly correlative. The steep-sided horses may have been transported from the Rand Mountains area by movement on the Garlock fault (see subsection below entitled "Garlock Fault Displacement").

ROCK TYPES AND PETROGRAPHY

In my investigations of the southernmost Sierra Nevada, I have made only cursory examination of the correlatives of the Rand Schist in the Garlock fault zone. My data come from the west third of the larger horse and from the smaller horse near Mojave. In addition, I summarize the data of Wiese (1950), who examined and described rocks from part of the larger horse.

Probably the best and most continuous exposures of Rand Schist are at the blunted west end of the larger horse in roadcuts along the access road for the Tehachapi

TABLE 2.—*Oxygen-isotopic data for selected quartzite samples from the Rand Schist*

[See figure 3 for locations of samples. All values in permil standard mean ocean water (SMOW)]

Sample	Description	$\delta^{18}O$
	Largest body of Rand Schist	
3281	Dense quartz mosaic, with aligned trains of pale-pink garnet, minor muscovite, and pale-brown biotite.	19.53
3344	Thin-layered quartzite (no thin section)------------------------	16.97
	Body of Rand Schist northwest of Mojave	
4050A	Thin-layered quartzite (no thin section)------------------------	14.84
4054A	Mosaic of quartz, pale-pink garnet, well-aligned olive-brown biotite, and lesser muscovite.	11.87
	Body of Rand Schist in the San Emigdio Mountains	
3023C	Quartz, with thin layers relatively rich in red-brown biotite, muscovite, dusty sodic plagioclase, and pale-pink garnet. rounded zircon, apatite.	16.36
3063	Quartz, with thin layers relatively rich in lacy, untwinned sodic plagioclase, yellowish-brown biotite, and garnet.	18.06

crossing of the California Aqueduct. North of Bear Trap Canyon, thin-layered, light to dark schist containing varying proportions of quartz, muscovite, biotite, and minor sodic plagioclase is exposed. A common associate here is thin-layered quartzite (90–95 percent quartz), with trains of garnet, mica, and minor feldspar that probably reflect reconstituted argillaceous impurities from the original rocks. These rocks are partly dark colored from disseminated carbonaceous matter (graphite) and stained red by iron oxides. The overall appearance, composition, and setting of these quartzite layers suggest that they were originally chert. Oxygen-isotope determinations on quartz from two quartzite samples from the western part of the large horse give $\delta^{18}O$ values of 16.97 and 19.53 permil SMOW (table 2), well beyond the range of igneous quartz, and confirm that these rocks are, indeed, metachert (Ivan Barnes, written commun., 1979). The presence of some coarse-grained, green actinolite pods and local epidote-rich layers suggests metavolcanic rocks. Quartz sills are also found that range from thin wisps to lenses to thick bull-quartz veins. Field relations suggest that these veins have been sweated out of the schist during metamorphism. These white quartz segregations are an index to the Rand and related schists.

Two traverses across the larger horse about 8 km farther east revealed the same rock types as along the aqueduct road. Quartzite (metachert) is abundant, but because it is generally resistant and preferentially exposed, it may appear to be more abundant than it actually is. In addition to the common silvery schist, dark spotted schist occurs in which the spots are small albite crystals around which are molded pale-green actinolitic amphibole sprinkled with sphene and epidote. This rock is almost certainly metavolcanic. Associated with the dark spotted schist is a small knobby outcrop of serpentinite, a few tens of meters across at most, that is crisscrossed with carbonate veinlets. The contact with the surrounding rocks is not exposed. An X-ray-diffraction pattern for the serpentinite indicates that it is dominantly antigorite and talc. The nearby float of coarse massive amphibolite, though rare in this horse, is distinctive. A cursory look suggests that the amphibolite is dominated by a mat of blocky hornblende crystals, but closer observation reveals that the blocky crystals are acicular bundles of amphibole, probably actinolite. This amphibolite in no way resembles the dark, hornblende-rich rocks north of the horse composed of Rand Schist.

Wiese (1950), who examined the central 25 km of the largest horse, observed that the most common rock types are green amphibolite or chlorite schist, with subordinate quartzite and brown mica schist. He noted that albite porphyroblasts are present "in all but the purest quartzite." On the basis of my brief observations and Wiese's (1950) descriptions, I suggest that volcanic rocks and associated chert were the dominant protoliths for the rocks of the larger horse. Sandy and argillaceous rocks also were almost certainly present, but I suspect they were tuffaceous, at least in part. Structural contortions and disruptions preclude any statement about the original thickness of the section.

The smaller horse northwest of Mojave is composed dominantly of dark-gray to green, highly foliated schist that is maculose with poikilitic albite (in part black with included carbonaceous material). The schistose groundmass is dominated by acicular green actinolite and contains lesser biotite and epidote. Associated with this schist is thin-layered quartzite containing micaceous partings, minor albite, and small garnets. These rocks are almost certainly a metachert. $\delta^{18}O$ values of 11.87 and 14.84 permil SMOW (table 2) were determined on quartz from two samples (Ivan Barnes, written commun., 1979). The ever-present white quartz segregations are also found here. One specimen of dark albitic schist contains abundant clinopyroxene, which is anomalous in this suite. Another unusual specimen, which may be an albitized intrusive(?) rock, is spotted with scattered black poikilitic albite crystals, as much as 2 mm long, and, in addition, contains fresh-looking, discretely twinned albite crystals, large areas of quartz, and large discrete K-feldspar masses with planar faces against albite. Iron-stained relict prismatic crystals are present that may have been amphibole. Thus, the Mojave horse appears to be composed entirely of metavolcanic rocks and metachert.

RAND(?) SCHIST IN THE SAN EMIGDIO MOUNTAINS

Extending westward from Grapevine Canyon to the area of Antimony Peak (Ross, 1981) is a belt of distinctive dark quartzite and schist (pl. 1) that, early in my fieldwork in the San Emigdio Mountains, I considered to be part of the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. Petrographic study, however, revealed the disturbing fact that the quartzite and schist are generally of lower metamorphic grade than the associated gneissic and amphibolitic rocks. This difference in metamorphic grade suggests that the two terranes are separated by a structural break.

The belt of quartzite and schist is characterized by rocks in which coarse-grained, dark, biotite-rich layers alternate with impure quartzite layers, also dark colored. In thin section, the most striking feature of these rocks is the abundance of large poikilitic grains of virtually untwinned sodic plagioclase that are clouded with graphite or other carbonaceous matter. These plagioclase crystals impart a dark color to the quartzite layers. Another unusual feature of these rocks is the presence of discrete chlorite crystals whose habit and interference color suggest that they are not a penninitic alteration product but are primary and, as such, indicate relatively low metamorphic grade. The belt of quartzite and schist contains conspicuous thick white bull quartz veins.

The physical appearance and metamorphic grade of these rocks is strikingly reminiscent of rocks that I have examined in Rand Schist in the western part of the Rand Mountains. This similarity and the discordance in metamorphic grade between the rocks of the dark quartzite and schist belt and the enclosing, higher grade gneissic rocks suggest strongly that a sliver of the Rand Schist has been emplaced tectonically in the San Emigdio Mountains. Cursory examination of the sliver of probable Rand Schist in the San Emigdio Mountains suggests that it is dominantly metasedimentary, whereas the two bodies of Rand Schist along the Garlock fault have strong metavolcanic affinities. The structural significance of this newly discovered sliver of Rand Schist(?) is discussed below in the subsection entitled "Pastoria Thrust Controversy."

PLUTONIC ROCKS

UNITS NORTH OF THE GARLOCK FAULT

GRANITIC ROCKS RELATED(?) TO THE SAN EMIGDIO-TEHACHAPI TERRANE

Quartz Diorite to Tonalite of Antimony Peak

A relatively homogeneous granitic body underlies an area of a few square kilometers near Antimony Peak. The main body of quartz diorite to tonalite of Antimony Peak intrudes gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains and is faulted against granodiorite of Lebec to the west. In part, the main Antimony Peak body is slightly porphyritic and contains subhedral, well-twinned, weakly zoned andesine crystals. K-feldspar is present only locally. Quartz is abundant and commonly is strongly strained; in part, mortar texture has developed. Olive to green hornblende generally exceeds brown biotite in abundance. The dark minerals invariably occur in irregular anhedral crystals, and in part they are extensively altered. Prehnite is locally abundant as ovoid lozenges and rosettes in bleached(?) biotite. Some prehnite also occurs in veinlets and as primary(?) accessory crystals. West of the main Antimony Peak mass are several small bodies of grossly similar lithology.

The Antimony Peak unit is most probably magmatic but has been contaminated by the mafic metamorphic basement. The Antimony Peak rocks may also be local melts derived from gneissic and amphibolitic rocks. Certainly there is a marked contrast between the contaminated

Antimony Peak rocks and the younger, uncontaminated Lebec and Brush Mountain bodies that are clearly magmatic. An $^{87}Sr/^{86}Sr$ ratio of 0.70337 from one sample (DR-3023, table 28) of the Antimony Peak unit indicates a mafic oceanic source.

Note that the modal plot (figs. 5, 6; table 3)[1] includes only the more homogeneous-looking rocks of the Antimony Peak unit, but even these selected samples show a considerable range of mineral percentages, particularly of dark minerals.

[1]The igneous-rock classification adopted by the International Union of Geological Sciences (Streckeisen, 1976) is used in this report.

Diorite to Tonalite of Tehachapi Mountains

The Tehachapi Mountains unit is a rather ill-defined assemblage of the more homogeneous parts of the mafic terrane of the San Emigdio-Tehachapi Mountains. Though locally homogeneous and apparently intrusive, these rocks are more likely to be streaky, foliated, and somewhat gneissic. These are the rocks that Wiese (1950) presumably called diorite. The samples in the modal plot (fig. 7;

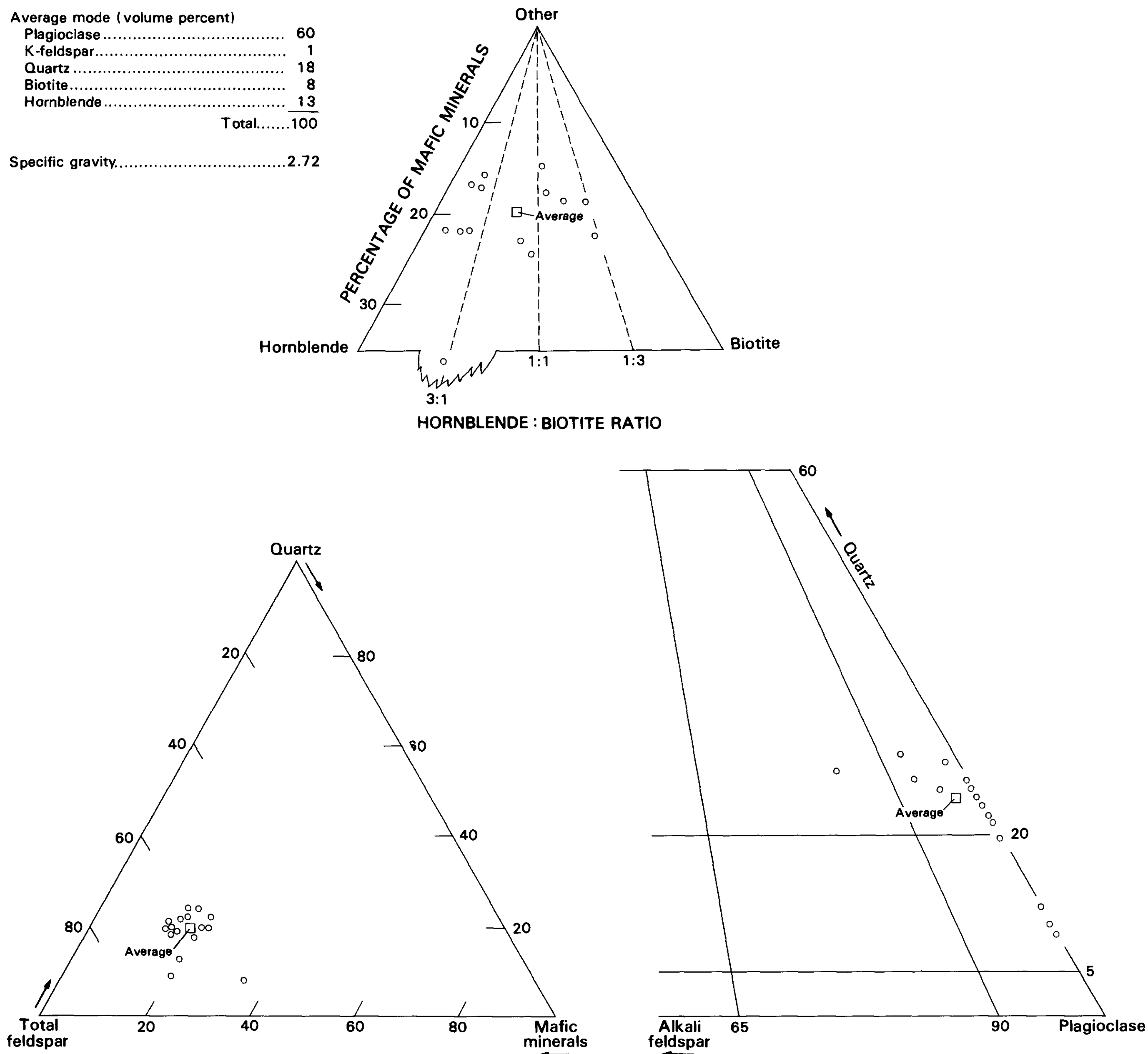


FIGURE 5.—Modal plots of the quartz diorite to tonalite of Antimony Peak.

table 4) were selected for their homogeneity—they represent the most intrusive-looking rocks of the diorite to tonalite. Although they are grossly similar to some of the rocks of the Antimony Peak unit, the Tehachapi Mountains rocks contain less quartz and more dark minerals, particularly hornblende. These hornblende-rich rocks appear to contain more metamorphic material than the Antimony Peak mass. I have speculated that at least some of the hornblende-rich rocks of the Tehachapi Mountains are locally derived from the metamorphic pile. These dark rocks are texturally and mineralogically transitional between dark amphibolitic rocks and the tonalite of Bear Valley Springs.

Some specimens of the Tehachapi Mountains unit contain clean, well-twinned, weakly zoned intermediate to calcic andesine (locally, labradorite). Quartz is also common and ranges from almost unstrained grains to highly undulatory and sutured grains. Locally, the quartz shows

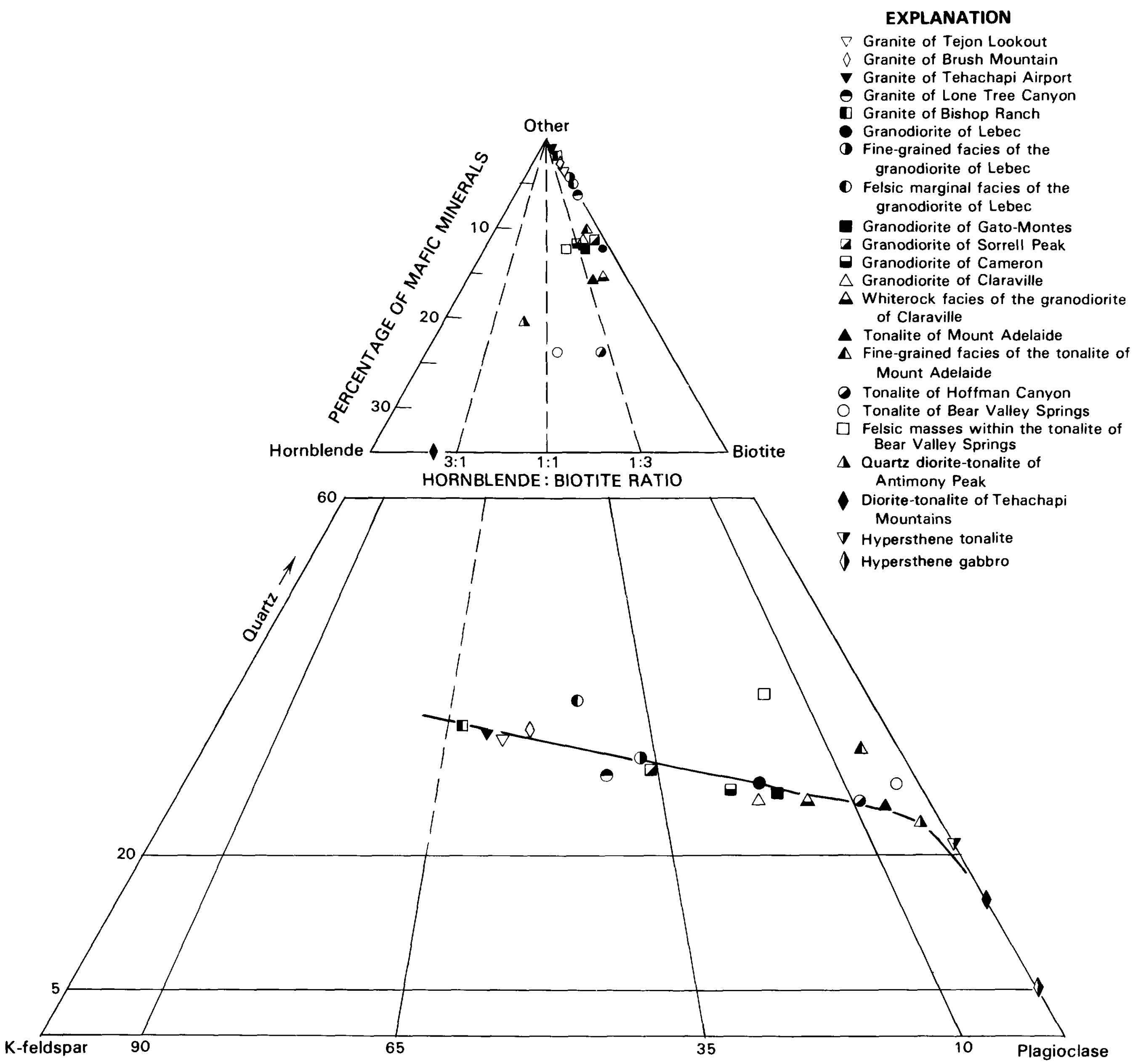


FIGURE 6.—Modal averages of plutonic-rock units. Line shows generalized trend, with noticeable deviations by some felsic masses, the tonalite of Mount Adelaide, the dark diorite to tonalite of Tehachapi Mountains, and the gabbroic hypersthene-bearing rocks.

TABLE 3.—*Modes of the quartz diorite to tonalite of Antimony Peak*

[All modes in volume percent. Others: E, epidote; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
784B	65	---	18	2	15	---	2.73
3000A	53	4	23	14	5	1(S)	2.70
3000B	58	1	23	10	8	---	2.68
3007	62	---	20.5	8	7	2.5(E)	2.71
3008A	56	---	7.5	9.5	27	---	2.80
3111	57	---	19	11.5	12.5	---	2.69
3022A	70	---	8	2	20	---	2.76
3023	66	---	12	3.5	18.5	---	2.73
3029B	64	---	19	17	---	---	2.72
3133	55.5	2.5	19	17	6	---	2.74
3150D	47	9	21	10	13	---	2.73
3152B	61	---	[1]17	4	18	---	2.75
3153	64	---	19	3	14	---	2.73
3158	56	4	21	12	7	---	2.69
3160	64	---	20	3	13	---	2.69
Average--------	60	1	18	8	13	---	2.72
Standard deviation.	6.0	2.6	4.9	4.9	6.3	---	.03

[1]Biotite+hornblende.

an impressive mortar texture. Hornblende, the chief dark mineral, occurs in various colors—pale-green and brown, light-olive, and deeper olive shades. The variety of hornblende color is another factor that points to the heterogeneity of this unit. Brown biotite is locally abundant, but generally it is uncommon. $^{87}Sr/^{86}Sr$ ratios of 0.70364 and 0.70446 for two samples (DR–3097, DR–3098A, table 28) attest to a mafic oceanic affinity for at least part of the Tehachapi Mountains unit.

TONALITE OF BEAR VALLEY SPRINGS

SETTING

The tonalite of Bear Valley Springs, rich in hornblende and biotite, is the most extensive granitic body in the study area, extending from the north edge of the area to the Garlock fault, a distance of about 60 km. It forms a generally north-south trending belt 20 to 25 km wide and crops out over an area of nearly 800 km^2 in the study area. The mass is cut off on the south by the Garlock fault; correlative rocks have not been found on the south side of the fault. These rocks extend northward from the study area for an unknown distance.

Larsen (1948, p. 69) noted that tonalites similar to the Bonsall Tonalite of the southern California batholith and carrying the abundant dark tabular inclusions characteristic of the Bonsall appear to be similar to tonalites widespread in the southern Sierra Nevada. He also noted that rocks similar to the Bonsall are found along the road from Bakersfield to Tehachapi (California Highway 58) and in Kern Canyon east of Bakersfield (along California Highway 178). At both these localities, the unit to which Larsen referred is the tonalite of Bear Valley Springs. From Larsen's statement, it is difficult to know whether or not he was suggesting a correlation between the two areas, but some geologists have interpreted his statement to mean that he was.

I suggest that the physical similarity between the Bonsall Tonalite (a rock type that may encompass more than one inclusion-bearing tonalite pluton) and the tonalite of Bear Valley Springs indicates a similar environment of emplacement for the two units, though probably not a correlation in the strict sense. Both of these tonalites are typical of batholithic terranes west of the quartz diorite line of Moore (1959), and other similar-appearing, dark, inclusion-rich tonalites occur farther north in the western Sierra Nevada.

PETROGRAPHY

The Bear Valley Springs mass is characterized by abundant anhedral dark minerals that occur in irregular aggregates and clots. Foliated, in part cataclastic (smeared) rocks are common. The abundance of hornblende is the next most characteristic feature of this mass; only rarely is the hornblende subhedral to euhedral, and commonly it is intergrown with biotite in aggregates and clots. Although the Bear Valley Springs mass contains irregular dark minerals, it has a definite granitic texture. Plagioclase is weakly zoned, well twinned, in part subhedral, and generally in the intermediate andesine range. Quartz is common in large irregular grains that generally show some strain or mosaicking; some is granulated. Generally, K-feldspar is absent (table 5) or is present only as small interstitial grains; locally, particularly near Bear Valley, in Caliente Canyon, and north of the Walker Basin, it makes up 5 to 14 percent of the rock. The K-feldspar is largely untwinned, but some shows faint grid twinning and weak perthitic structure. In Caliente Canyon, conspicuous salmon-colored K-feldspar phenocrysts are present.

Biotite is generally brown to opaque and locally is reddish brown in thin section. It occurs in anhedral grains that are partly chloritized, and contains round grains and

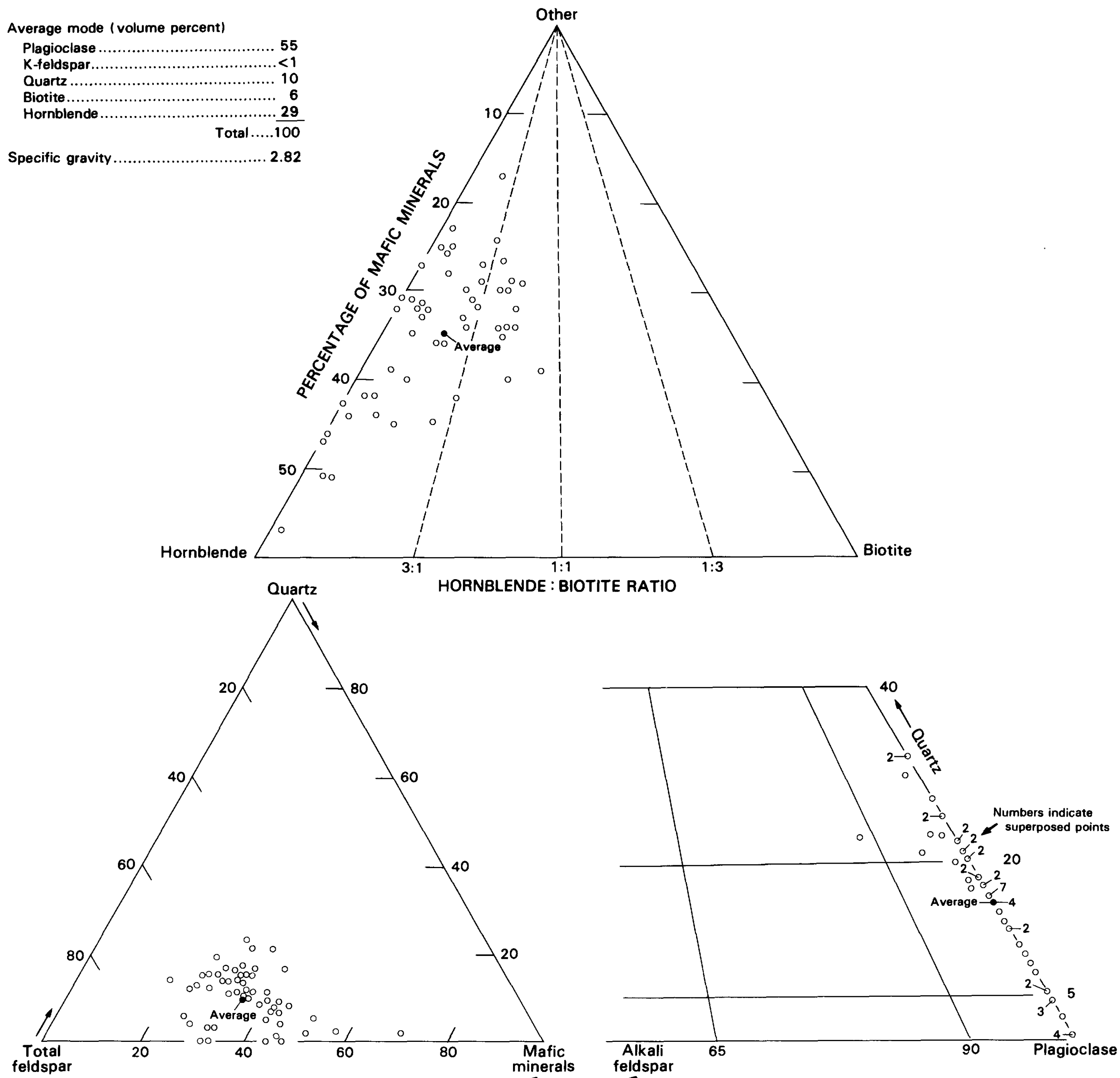


FIGURE 7.—Modal plots of the diorite to tonalite of the Tehachapi Mountains.

TABLE 4.—*Modes of the diorite to tonalite of Tehachapi Mountains*

[All modes in volume percent; n.d., not determined. Others: G, garnet; O, opaque minerals]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3097	53	---	8	3	36	---	2.85
3098A	47	---	8	6	39	---	2.86
3098B	44	---	5	3	48	---	2.87
3100	50	---	7	---	43	---	2.89
3246	53	---	11	6	30	---	2.84
3252	48	[1]1	21	10(?)	20(?)	---	2.76
3254A	50	---	8	2	40	---	2.85
3266B	49	---	9	2	40	---	2.90
3266-1	53	---	<1	---	47	---	2.84
3268	53	---	1	1	43	2(O)	2.89
3270	56	---	<1	[2]4	40	---	2.89
3276	69	---	4	---	[3]27	---	2.79
3283	62	[1]1	13	6	18	---	2.81
3285A	65	---	3	3	29	---	2.86
3285B	66	---	3	1	30	---	2.85
3304A	55	---	10	3	32	---	2.85
3333	55	[1]1	12	2	30	---	2.79
3345	41	---	2	1	56	---	3.02
3359A	56	---	19	2	[4]23	---	2.79
3360A	69	---	---	---	31	---	2.84
3400B	60	<1	15	[2]1	24	---	2.81
3407A	69	---	6	1	24	---	2.81
3419A	68	---	14	3	14	---	2.75
3430	50	<1	16	12	22	---	2.83
3431	52	---	15	7	26	---	2.79
3432A	52	---	17	7	24	---	2.81
3435A	56	[1]<1	11	3	30	---	2.82
3439A	47	---	2	2	[5]40	---	2.85
3442	53	---	15	3	[6]29	---	2.81
3448A	57	---	14	11	18	---	2.76
3552A	51	---	7	11	31	---	2.83
3572A	54	<1	14	12	20	---	2.78
3605	55	[1]3	15	6	21	---	2.76
3630A	51	---	4	10	35	---	2.82
3633B	68	---	---	---	32	---	2.84
3634	54	---	14	8	24	---	2.77
3635	54	[1]1.5	16.5	3	25	---	2.75
3651	60	---	10	10	20	---	n.d.
3709	44	---	16	15	25	---	2.80
3710	44	---	21	12	23	---	2.80
3729	54	---	10	6	30	---	2.83
3730A	50	1	15	13	21	---	2.82
3777	56	---	10	12	22	---	2.83
3793	48	---	23	10	19	---	2.76
3867	56	[1]1	14	7	21	---	2.77
3867-1	50	---	11	18	21	---	2.81
3895B	48	---	5	---	47	---	2.80
3899A	57.5	---	11	8	18.5	5(G)	2.81
3950	54	---	12	8	26	---	2.79
4009	51	---	9	5(?)	35	---	2.84
4045A	48	6	16	6	24	---	2.73
4192A	65	---	12	1	22	---	2.77
Average--------	55	<1	10	6	29	---	2.82
Standard deviation.	7.0	---	6.0	4.6	10.4	---	.05

[1]Late veinlets.
[2]Chlorite.
[3]Includes epidote.
[4]Some colorless to pale.
[5]Brown and pale actinolite.
[6]Minor clinopyroxene and actinolite.

trains of sphene. Hornblende, also generally in anhedral grains but locally subhedral, is olive green in various shades (mostly dark), and is strongly pleochroic. The hornblende is partially altered to epidote and calcite. Lacy to irregular cores of pale-green clinopyroxene are found in some hornblende crystals.

Honey-colored sphene is the most common accessory mineral. Metallic opaque-mineral grains have a sporadic occurrence; in some specimens they are virtually absent, and in others they make up at least 1 percent of the rock. Allanite, primary(?) epidote, zircon, and apatite are also present in trace amounts. Local pink garnet may reflect contamination by wallrocks.

Even though the Bear Valley Springs mass varies considerably in mineral content (fig. 8; table 5) and has a highly irregular, messy texture, all of which suggests contamination and probably some deformation, nevertheless there is a gross homogeneity to much of this mass that is probably best characterized by its outcrops near Bear Valley. It is evidently a largely intrusive mass that may have incorporated material from the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains.

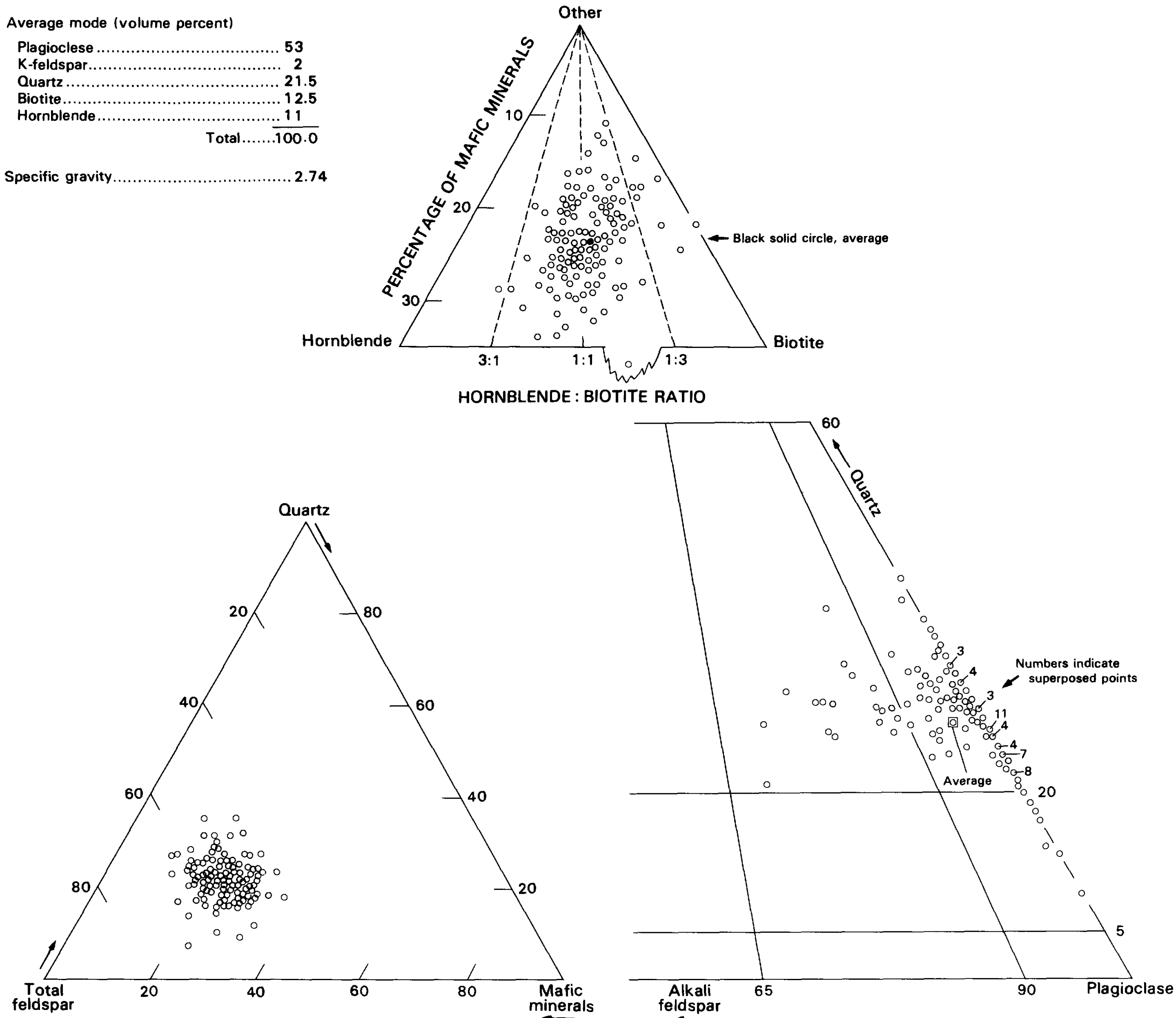


FIGURE 8.—Modal plots of the tonalite of Bear Valley Springs.

TABLE 5.—*Modes of the tonalite of Bear Valley Springs*

[All modes in volume percent; n.d., not determined. Others: A, allanite; C, clinopyroxene; G, garnet; O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3412	54	1	21	12	9	3(C)	2.77
3413	52	---	31	11	6	---	2.74
3427B	55	<1	26	11	8	---	2.76
3429A	62	<1	20	9	9	---	2.72
3444	61	1	20	10	8	---	2.75
3557	51	2	23	16	8	---	2.66
3559	52	---	25	13	10	---	2.70
3565	45	8	28	13	6	---	2.67
3571	51	1	23	13	12	---	2.735
3573	53	<1	20	12	15	---	2.76
3575	49	1	21	18	11	<1(C)	2.77
3576	51	<1	20	12	17	---	2.78
3578A	48	1	24	12	15	---	2.72
3582A	60	---	14	17	9	---	2.725
3586A	48	---	27	11	14	---	2.68
3600B	48	7	22	16	1	6(C)	2.75
3621	46	<1	35	15	4	---	2.69
3628	54	---	17	12	17	---	2.72
3650A	59.5	<1	24	7	9.5	---	2.72
3656	49.5	2.5	24	10	14	---	2.76
3664	44	3	23	14	16	---	2.76
3667	46	3	23	14	14	---	2.76
3668	47	3	20	13	17	<1(S)	2.77
3669	47	3	25	16	9	<1(S)	2.73
3670	57	<1	16	12	15	<1(S)	2.78
3672	47	<1	24	18	11	<1(S)	2.78
3674	44	<1	23	18	15	<1(S)	2.79
3678	49.5	---	19	13.5	18	<1(A)	2.78
3683	53	1	23	12	11	---	2.77
3690A	61	1	10	13	15	<1(S)	2.73
3691	48	3	24	15	10	---	2.71
3693	51	5	18	14	12	---	2.71
3694	47.5	6	25.5	14	7	<1(S)	2.72
3698	44	3	27	13	13	---	2.77
3699	45	4	21	15	15	<1(O), <1(S)	2.77
3715	45	1	32	16	6	<1(O)	2.70
3718	43.5	2	17.5	23	14	---	2.75
3728A	44	8	35	9	4	<1(O)	2.66
3791	52	---	24	12	12	---	2.78
3792	55	---	20	15	10	---	2.81
3795	55	---	22	10	13	---	2.78
3816	54	---	20	12	14	---	2.74
3831	55	---	20	13	12	---	2.76
3833A	49	---	31	12	8	---	2.73
3835	49	5	21	12	13	<1(S)	2.76
3837	47	7	20	11	15	---	2.81
3838	50	1	23	15	11	---	2.77
3839	53	<1	21	14	12	---	2.76
3840	51	---	18	16	15	---	2.80
3852	46	2	22	15	15	---	2.76
3853	51	4	19	12	14	---	2.77
3854	51	5	18	11	15	---	2.76
3855A	49	9	28	8	6	---	2.70
3856A	45	12	25	15	3	<1(S)	2.71
3858	51	3	20	14	12	<1(S)	2.78
3860	50	1	20	16	13	---	2.78
3863	58	---	9	15	18	---	2.82
3865	55	1	20	13	11	<1(A)	2.74
3869	51	2	18	15	14	---	2.78
3871A	49	5	23	11	12	---	2.75
3872	53	4	18	12.5	12.5	---	2.74
3873A	49	8	22	10	11	---	2.76
3874	34	14	18	15	19	---	2.79
3925	56	---	26	12	6	---	2.63

TABLE 5.—*Modes of the tonalite of Bear Valley Springs*—Continued

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3927	52	1	28	10	9	---	2.69
3937	51.5	---	20	16	12.5	---	2.68
3945	48	---	24	16	12	---	2.62
3962A	45	---	27	20	8	---	2.68
3963	48	2	22	22	3	3(C)	2.73
3973	59	---	21.5	13.5	6	---	2.67
3975C	47	11	25	12	5	---	2.68
3980A	69	---	7	22	---	2(G)	2.72
3991	60	---	19.5	13	7.5	---	2.71
4110A	54	<1	31	13	2	---	2.74
4111	48	4	27	13	8	---	2.73
4113A	54	---	17	8	21	---	2.79
4114	59.5	---	15	11.5	14	---	2.77
4115	51	---	18	10	21	---	2.78
4135	58	---	19	12	11	---	2.74
4138A	66	---	13.5	9.5	11	---	2.76
4141	54	---	28	8	10	---	2.74
4142	55.5	---	23.5	13	8	---	2.72
4154	56	1	23	10	10	---	2.71
4158A	54.5	---	19.5	17	9	---	2.77
4164	57	---	16	10	17	---	2.76
4175	53	---	21	14	12	---	2.72
4178	52	1	29	10	8	---	2.74
4179	62.5	---	17	10	10.5	---	2.77
4180	58	---	18.5	9	14.5	---	2.77
4181	52	---	25	11	12	---	2.75
4184	52.5	---	19.5	13	15	---	2.79
4185	53	3	23.5	7	13.5	---	2.71
4186	55.5	<1	18.5	15	11	---	2.76
4191B	51	---	19	13	17	---	2.78
4194A	59	---	21	6	14	---	2.74
4195	55	---	17	14	14	---	2.76
4196	59	---	19	9	12	1(O)	2.75
4202	58	---	16.5	8	17.5	---	2.78
4203	52.5	---	19	10.5	18	---	2.81
4204	60	1	18	9.5	11.5	---	2.76
4205	55	---	16	7	22	---	2.77
4210A	58.5	---	25.5	8	8	---	2.72
4213	59	---	22	9	10	---	2.75
4226	60.5	---	16	9.5	14	---	2.79
4228A	51	13	23	8	4	1(O)	2.68
4228B	56.5	1	21	12.5	9	---	2.73
4229	57	---	19	10	14	---	2.78
4230	54	---	12	13	21	---	2.82
4235	62	---	17.5	10	10.5	---	2.73
4239	66	---	17	5	11	1(O)	2.72
4240	60	---	22	14.5	3.5	---	2.73
4243	57.5	1	22.5	9	10	---	2.72
4250	55	---	22	14	9	---	2.77
4255	47	12	21	13	7	---	2.67
4258	49	6	21	15	9	---	2.75
4260	52	4	21	13	10	---	2.73
4261	49	6	22	15	8	---	2.75
4263	50	12	27	8	3	---	2.68
4277	52	3	24	14	7	---	2.72
4281	46	7	21	13	13	---	2.75
4305	54	1	23	19	3	---	2.75
4422	61	<1	17	12	9	1(S)	n.d.
4425	53	---	16	18	13	---	2.77
4443A	56	4	24	9	7	---	2.74
4458	50	<1	20	19	11	---	2.75
4565A	62	---	17	14	7	---	2.75
Average--------	53	2	21.5	12.5	11	---	2.74
Standard deviation.	5.5	3.2	4.5	3.4	4.3	---	.04

Relation to the Gneiss, Amphibiolite, and Granulite of San Emigdio-Tehachapi Mountains

Relations observed in the field and petrographic study indicate that the tonalite of Bear Valley Springs and the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains may be genetically related. Foliation, mafic schlieren, ghost gneiss patches, irregularly intergrown biotite and hornblende, and abundant dark ovoid inclusions contribute to the overall heterogeneous texture of the tonalite of Bear Valley Springs. In the study area, the Bear Valley Springs mass either has been strongly contaminated from the metamorphic basement it intruded or is the ultrametamorphosed end product of such a terrane.

It was extremely difficult—in fact, virtually impossible in some localities—to clearly distinguish between the tonalite of Bear Valley Springs and the diorite to tonalite of the Tehachapi Mountains. Table 4 lists samples assigned to the Tehachapi Mountains unit on account of a great preponderance of hornblende over biotite. The locations of these samples and their biotite/hornblende ratios are shown in figure 9.

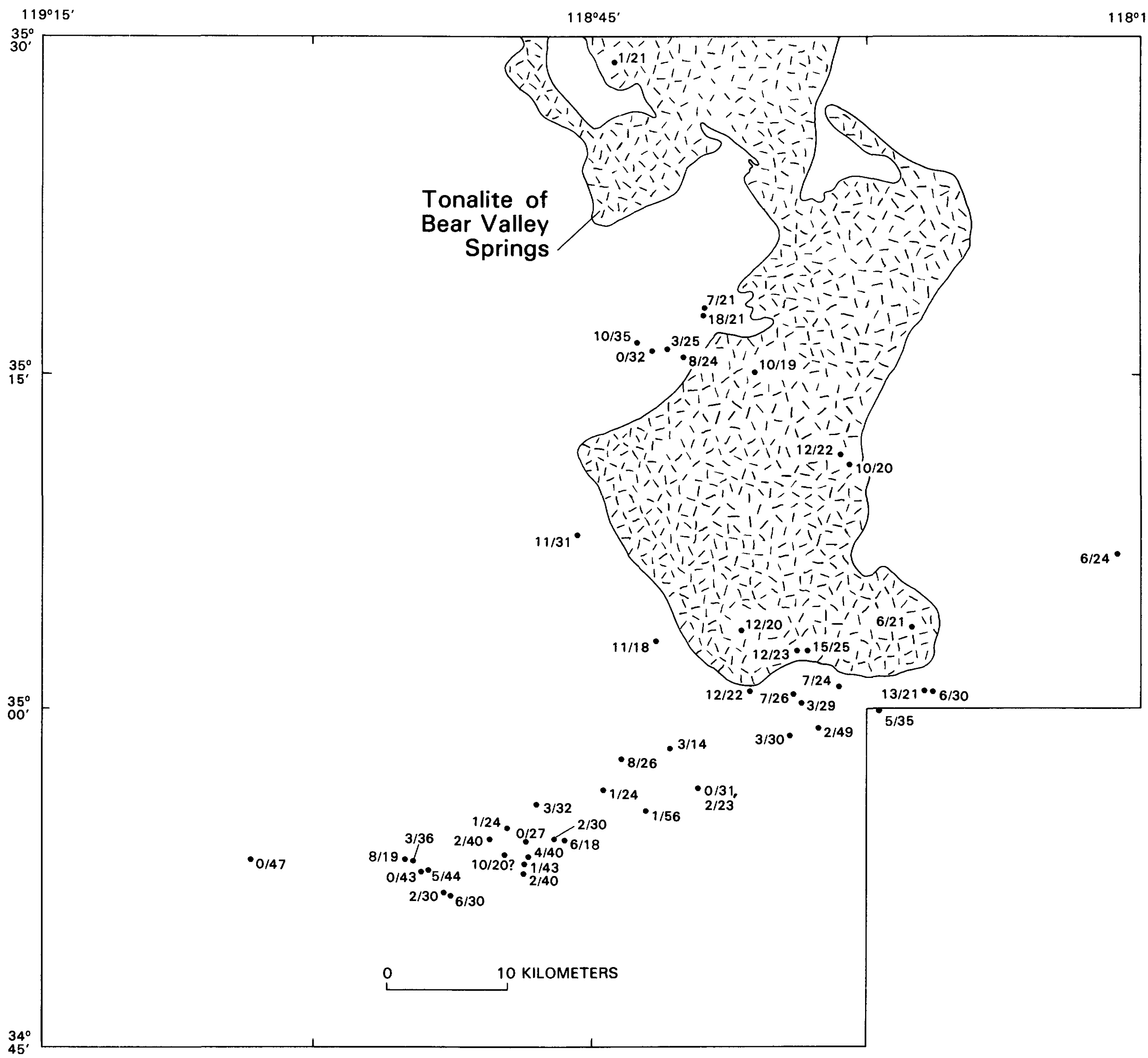


FIGURE 9.—Locations of samples of the diorite to tonalite of the Tehachapi Mountains, showing biotite/hornblende ratios. Generalized outline of the tonalite of Bear Valley Springs is shown for comparison with figure 11.

Table 5 lists samples assigned to the tonalite of Bear Valley Springs; in these samples, biotite and hornblende are present in grossly similar amounts. Figure 10 summarizes the dark mineral content of the Bear Valley Springs unit; it shows a clustering around 20 percent of total dark minerals, but with a considerable range in both total dark minerals and in biotite/hornblende ratio. Specimens of the more homogeneous rocks of the diorite to tonalite of the Tehachapi Mountains diverge from this cluster to more hornblende rich specimens. Interestingly enough, the dark-mineral content of the Antimony Peak mass is more like that of the Bear Valley Springs mass than like the diorite to tonalite of the Tehachapi Mountains. Exceptions—such as sample 4115 in the Bear Valley Springs mass, which contains 10 percent biotite and 21 percent hornblende, and sample 3867–1 in the Tehachapi Mountains unit, which contains 18 percent biotite and 21 percent hornblende—point out the dilemma of making a sharp distinction between these two units.

In an attempt to see whether some relation exists between the distance from the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains and the amounts of biotite and hornblende, particularly the biotite/hornblende ratio in the Bear Valley Springs mass, a plot of these features was made (fig. 11). These plots show no obvious zoning or trends based on the amount of dark minerals. A 10-percent contour vaguely suggests that both hornblende and biotite are decreasing near the north end of the study area, but this possible trend needs to be checked by means of further work to the north. The trend (if it is a trend) certainly is vague at best because the concentration of samples containing hornblende in excess of biotite is as great or even greater in the north part of the map area, and as previously noted, inclusion swarms and ghost gneiss patches occur close to the north margin of the study area. Hornblende in excess of biotite in the Bear Valley Springs mass is apparently a widespread characteristic and not solely a reflection of proximity to the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains. If contamination from metamorphic rocks is a factor in the composition of the Bear Valley Springs body, that contamination must be a regional effect from some depth below the present level of exposure.

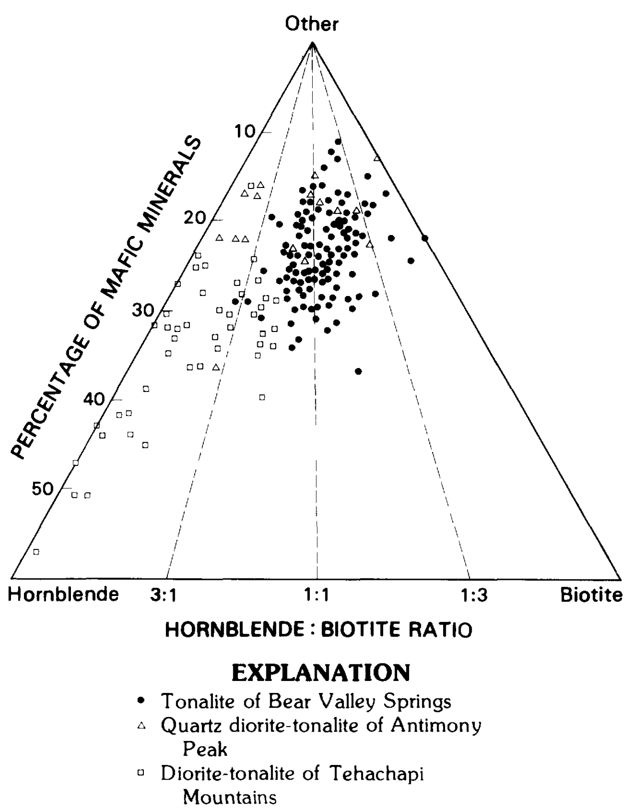


FIGURE 10.—Mafic-mineral compositions of the tonalite of Bear Valley Springs, the quartz diorite to tonalite of Antimony Peak, and the diorite to tonalite of Tehachapi Mountains.

Felsic Masses Within the Tonalite of Bear Valley Springs

Several bodies of tonalite have been mapped within the outcrop area of the tonalite of Bear Valley Springs. These bodies are coarser grained, lighter colored, and more quartz rich than the tonalite of Bear Valley Springs. Although their mapped form suggests that these bodies are plugs within the tonalite body, field study has not confirmed this interpretation. I found no dikes of these felsic rocks cutting the tonalite, and in several places the contact appears to be gradational, suggesting that the felsic masses may have been derived from later-crystallizing parts of the tonalite which were more or less in place—not later surges of magma. These felsic rocks are best exposed and most easily visited along California Highway 58 near Keene. These outcrops are part of a mass underlying an area of at least 20 km^2. The lighter colored, bouldery slopes of the felsic mass contrast strongly with the darker tonalite to the east and west. Other felsic masses are considerably smaller but are generally similar texturally and mineralogically. In addition to the mapped felsic bodies, other, smaller areas of similar rocks also appear to be local variations within the tonalite.

Included in this unit is a small mass on Caliente Creek Road east of Caliente that may correlate with the granodiorite of Sorrell Peak. The texture and modal composition of these Caliente Creek rocks are similar to the Sorrell Peak mass, and the Caliente Creek rocks contain more K-feldspar and less quartz than average felsic rocks within the tonalite of Bear Valley Springs. The Caliente Creek specimens, however, are within the modal range of felsic rocks of the Bear Valley Springs mass and for

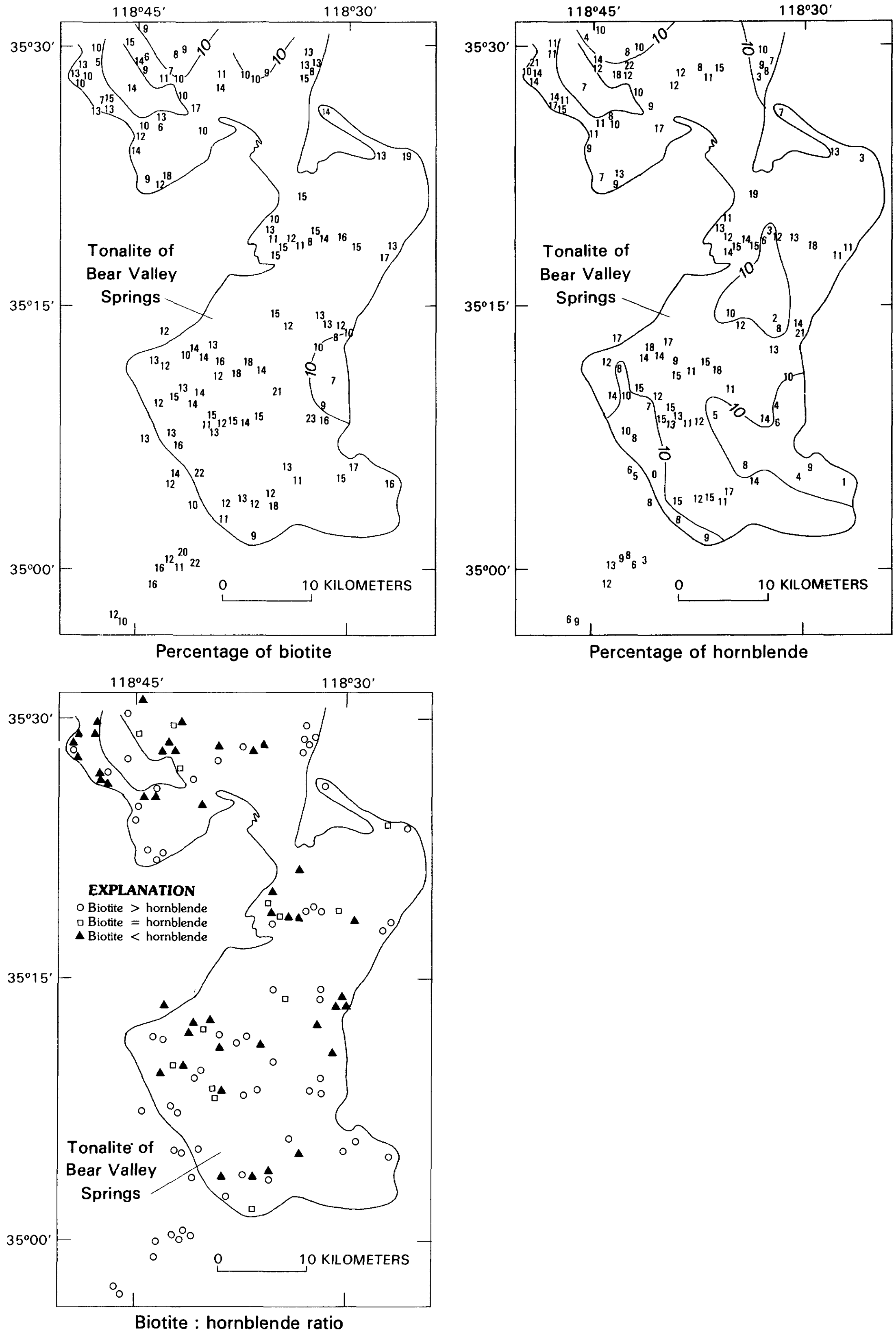


FIGURE 11.—Modal percentages of biotite and hornblende and biotite/hornblende ratios for samples of the tonalite of Bear Valley Springs. 10-percent contour is shown for biotite and hornblende. Compare with figure 9.

convenience are lumped with them on plate 1. These felsic rocks texturally are similar to the tonalite of Bear Valley Springs. Modal data for the felsic masses are presented in figure 12 and table 6.

Plagioclase is well twinned; however, the twin planes are thinner than in the tonalite of Bear Valley Springs, and determinations suggest sodic andesine or even, in part, oligoclase for these rocks. K-feldspar ranges in abundance from 0 to 20 percent; where present, it commonly forms large grains that range from untwinned to strongly grid twinned. Locally, the K-feldspar crystals are phenocrysts. Quartz is abundant in grains as large as 5 mm. Undulatory extinction, mosaicking, and granulation of the quartz reflect considerable strain in these rocks, particularly in foliated samples. Dark-brown biotite and dark-olive-green hornblende are nearly identical in appearance to those minerals in the darker tonalite of Bear Valley Springs. Discrete primary(?) muscovite flakes are found in some specimens. Sphene, zircon, and apatite are also found, but metallic opaque-mineral grains are rare.

Dibblee and Warne (1970) mapped several areas of altered rock along Tweedy Creek and suggested that the rock "* * * has been highly altered by hydrothermal solutions or steam, probably related to emplacement of dacite dikes of Tertiary age which crop out to the southeast." I mapped an additional altered body on the north side of

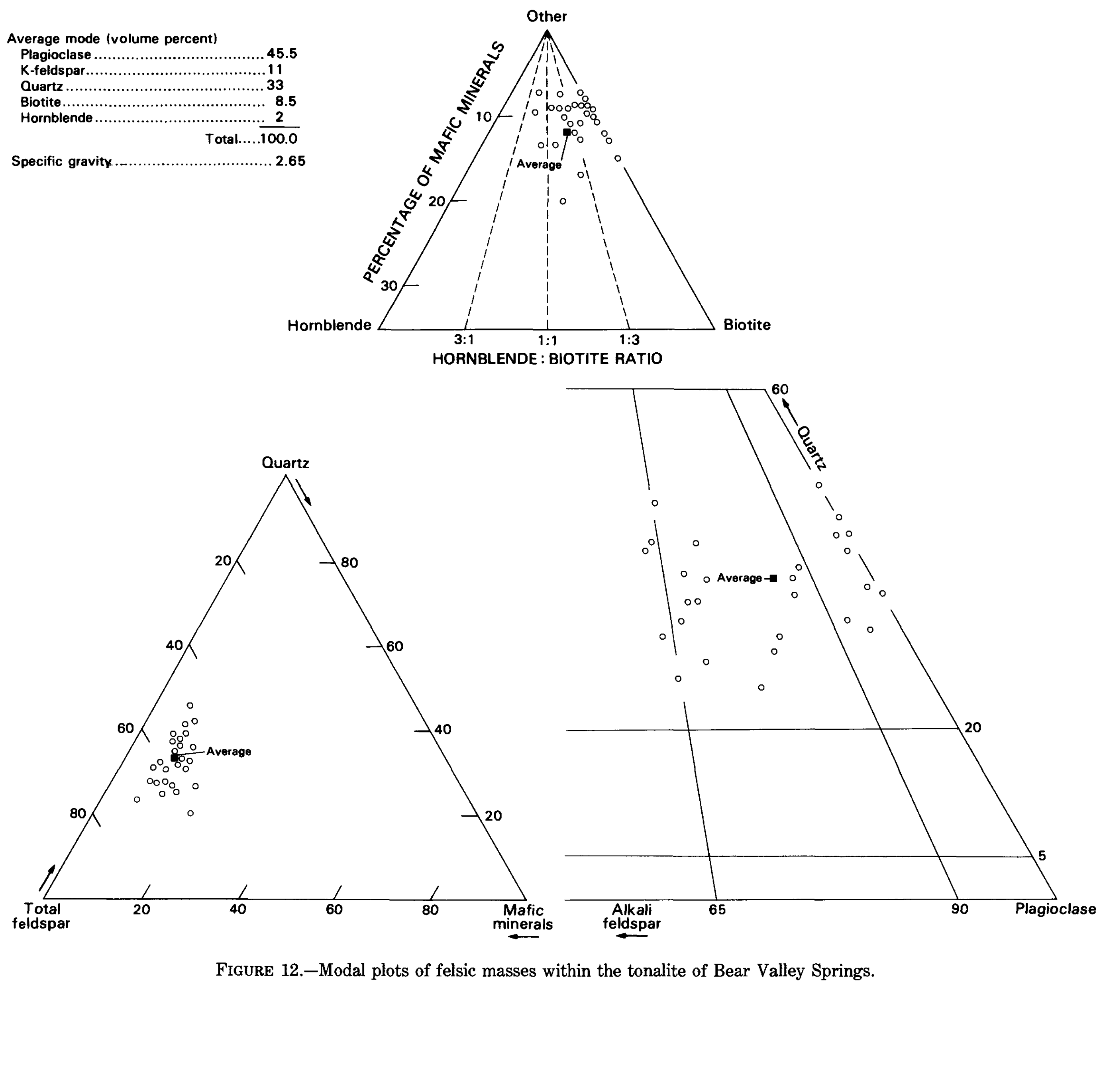


FIGURE 12.—Modal plots of felsic masses within the tonalite of Bear Valley Springs.

TABLE 6.—*Modes of felsic masses within the tonalite of Bear Valley Springs*

[All modes in volume percent. Others: G, garnet; O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3538	34.5	19	38.5	8	---	---	2.64
3539	48	7.5	31	8	5.5	---	2.70
3544B	48	12	25	15	---	---	2.67
3590C	38	15	38	7	2	---	2.62
3612	51.5	.5	39	6	3	---	2.68
3638	51.5	4.5	27	12	5	---	2.70
3648A	40	15	34	9	2	---	2.65
3653	50	1	36	13	---	---	2.63
3687	46	14	20	12	8	---	2.69
3689	58	3	28	8	3	<1(S)	2.67
3695	49	11	27.5	9.5	3	---	2.68
3697	47	7	33	6	7	<1(O), <1(S)	2.69
3704	47	---	46	3	4	---	2.66
3724A	50	1	39	7	3	---	2.64
3782	44	19	25	12	---	---	2.62
3783	49	6	35	10	---	---	2.61
3787	50	---	41	9	---	---	2.65
3788A	38	17	33	9	3	---	2.61
3794A	42	20	31	5	2	---	2.66
3815	62	---	28	10	---	---	2.68
3859A	34	20	37	3	6	---	2.65
3968A	55	1	33	10	---	1(G)	2.61
3970	32	16	42	9	---	---	2.68
4121	42	18	32	7	1	---	2.65
4122	42	18	32	5	3	---	2.61
4468	40	23	28	9	---	---	2.63
4469	45	24	24	7	---	---	2.62
Average--------	45.5	11	33	8.5	2	---	2.65
Standard deviation.	7.2	8.2	6.2	2.9	2.4	---	.03

Tweedy Creek (easternmost body with crosslined pattern on pl. 1) that contains angular, gray to pink, felsic volcanic fragments. This easternmost body, and the other altered areas as well, may represent vent breccias. Dibblee and Chesterman (1953) described a possibly comparable vent breccia containing gray to pink rhyolite to dacite fragments from a locality about 3 km south of the Walker Basin and about 14 km north of the Tweedy Creek locality.

TONALITE OF MOUNT ADELAIDE

SETTING

Three large masses of biotite tonalite intrude the Bear Valley Springs body in the northern part of the study area. The largest mass, north of the Walker Basin, covers an area of about 105 km^2 within the study area. The ovoid outcrop pattern suggests that this mass may have forced its way into the metasedimentary rocks that now bound it on the northeast and southwest. On the southwest side of the tonalite body, metasedimentary rocks form a septum, discontinuous to the southeast, between the Mount Adelaide body and the tonalite of Bear Valley Springs.

A second pluton of the Mount Adelaide rock type underlies an area of nearly 100 km^2 southwest of the Walker Basin. These rocks are bounded by the Breckenridge fault on the southeast and by a northeast-trending belt of mafic rocks and the Pampa Schist on the northwest. The parallelism of these two contacts suggests at least some sort of structural control on emplacement of the Mount Adelaide intrusive mass. Similar exposures of the Mount Adelaide body on California Highway 58 to the south are probably connected to the mass southwest of the Walker Basin beneath the alluvium. Also, similar granitic rocks were penetrated in wells drilled to basement in the Edison oil-field area to the southwest.

The third mass of the Mount Adelaide rock type to the north is much smaller (30-km^2 area) within the study area, rather duck shaped, and separated from the Bear Valley Springs unit by a septum of Pampa Schist along the duck's head and bill. Its grossly lozengelike shape suggests that this mass was forcibly emplaced into the area now underlain by the Bear Valley Springs body in a manner analogous to the emplacement of the larger mass northeast of the Walker Basin.

Petrography

The tonalite of Mount Adelaide (fig. 13; table 7) is probably the most distinctive and easily recognized granitic rock type in the map area. It contains euhedral biotite in six-sided crystal plates, as much as 1 cm across but generally smaller. Although euhedral biotite is not present in all outcrops of the Mount Adelaide bodies, its recurrence is a reassuring index in mapping this unit.

Well-formed biotite and hornblende crystals and subhedral plagioclase crystals give this rock type a hypautomorphic granular texture. The plagioclase is andesine in well-twinned crystals that in part show oscillatory zoning. K-feldspar is generally sparse in this unit, but the mass northeast of the Walker Basin contains more than 10 percent K-feldspar in some specimens. Quartz is abun-

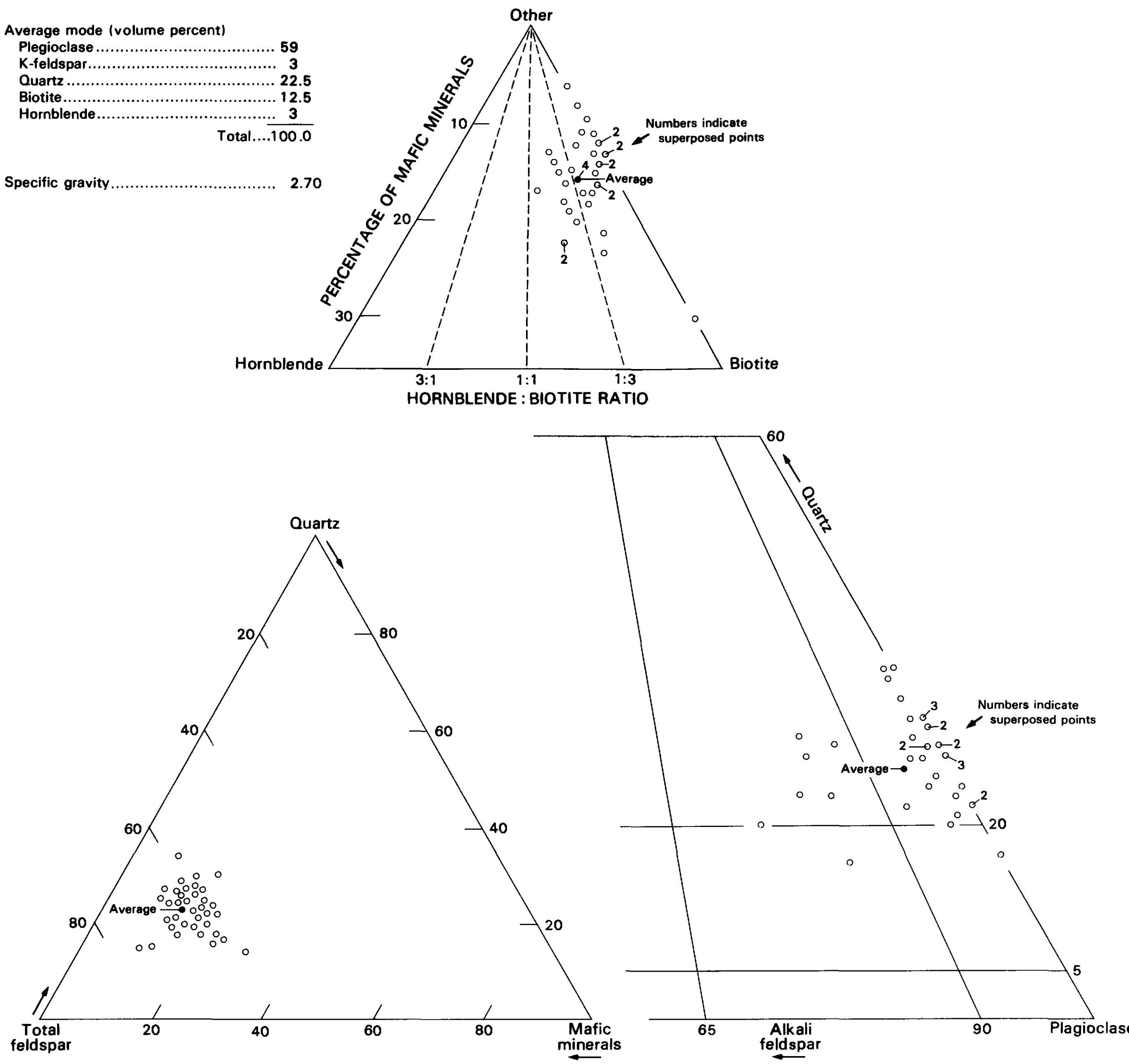


FIGURE 13.—Modal plots of the tonalite of Mount Adelaide.

TABLE 7.—*Modes of the tonalite of Mount Adelaide*

[All modes in volume percent; tr, trace. Others: 0, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3631	72	---	15	8	5	---	2.64
3631-1	67	---	19	9	5	---	2.73
3631-2	58	---	34	8	---	---	2.65
4137E	42	14	14	30	---	---	2.75
4144A	58	---	25	9	8	---	2.72
4145	59	<1	25.5	12	3.5	---	2.68
4145-2	52	1	30	14	3	---	2.66
4145-4	59	---	22	13	6	---	2.72
4148A	64	---	20	14	2	---	2.66
4189	62	1	24	12	1	---	2.70
4197	62	---	23	10	5	---	2.70
4220	63	---	24	13	---	---	2.70
4236	66	1	20	12	<1	1(0)	2.64
4238	59	1	27	12	1	---	2.70
4246	59	---	26	13	2	---	2.70
4253A	66	3	25	6	---	---	2.64
4253B	62	2	26.5	9.5	---	---	2.67
4265	51	10	23	11	5	---	2.69
4266	62	12.5	14.5	10	1	---	2.68
4273	54	13	20	10	2	1(S)	2.69
4275A	50	10	25	11	4	---	2.72
4275B	53	8	24	12	3	---	2.73
4282	60	2	20	12	6	---	2.71
4283	52	10	18	14	6	---	2.70
4285	58	2	22	14	2	2(S)	2.73
4287	65	2	18	13	1	1(S)	2.70
4294	59	2	16	14	8	1(S)	2.75
4304A	60	2	20	14	4	---	2.72
4308	55.5	4.5	17	14	8	1(S)	2.74
4315	60	tr	17	18	5	---	2.73
4319	56	1	22	17	4	---	2.73
4419	59	1	29	11	---	---	2.65
4420	56	1	30	13	---	---	2.69
4564	60	---	23	13	4	---	2.71
4567A	57	---	27	12	4	---	2.73
4569A	62	---	23	12	3	---	2.73
"Rock Pile"---------	61	<1	27	12	---	---	2.63
Average--------	59	3	22.5	12.5	3	<1	2.70
Standard deviation.	5.6	4.3	4.7	3.9	2.5	---	.04

dant in small to large grains that range from virtually unstrained to strongly undulatory and mosaicked. Biotite is generally deep brown to opaque in the two larger masses, but along California Highway 58 and in the smaller northwest mass it is in part moderate reddish brown. Hornblende is absent in some specimens but makes up a few percent of most samples; it is generally strongly pleochroic in shades of green and olive. The same thin sections that contain reddish-brown biotite also seem to contain colorless to pale-green hornblende. Metallic opaque-mineral grains are rather widely scattered in this mass.

Along the railroad line west of Caliente, along and east of Caliente-Bodfish Road near the north edge of the study area, and in several smaller patches, finer grained rocks are present that have much the same mineralogy and texture as the coarser, average Mount Adelaide rock types (fig. 14; table 8). Coarser, well-formed biotite crystals are found in many of these finer grained masses. Both west of Caliente and east of Caliente-Bodfish Road, near the north edge of the study area, there is a gradation between the coarser, "normal" Adelaide and the finer grained rocks. However, just north of the study area along Caliente-Bodfish Road, I have seen dikes of finer grained rocks cutting coarser grained, normal Mount Adelaide rocks. Some of the finer grained rocks are richer in K-feldspar and poorer in hornblende than are their coarser grained associates, features suggesting that they may be felsic derivatives of the normal Mount Adelaide type. Subsequent geologic mapping to the north of the study area has shown that the finer grained rocks are most probably younger bodies.

TONALITE OF HOFFMAN CANYON

Setting

The best exposures of the tonalite of Hoffman Canyon are along the canyon, a tributary to Jawbone Canyon in the northeastern part of the map area. This tonalite, together with associated gneissic and amphibolitic rocks, forms a large mafic inclusion or enclave in an area of notably felsic granitic rocks. The tonalite covers an area of some 95 km^2, and the gneissic and amphibolitic rocks cover an additional 30-km^2 area to the south near Jawbone Canyon. The limits of the tonalite mass to the north and west are somewhat uncertain, partly because of limited exposures. The contact to the south with the gneissic and amphibolitic rocks is a broad and arbitrary zone that lies between dominant tonalite on the north and dark amphibolitic and gneissic rocks on the south.

The Hoffman Canyon mass is enigmatic: Either it could be a separate hornblende-rich pluton that may or may not be correlative with the tonalite of Bear Valley Springs, or it may represent the granodiorite of Claraville that is

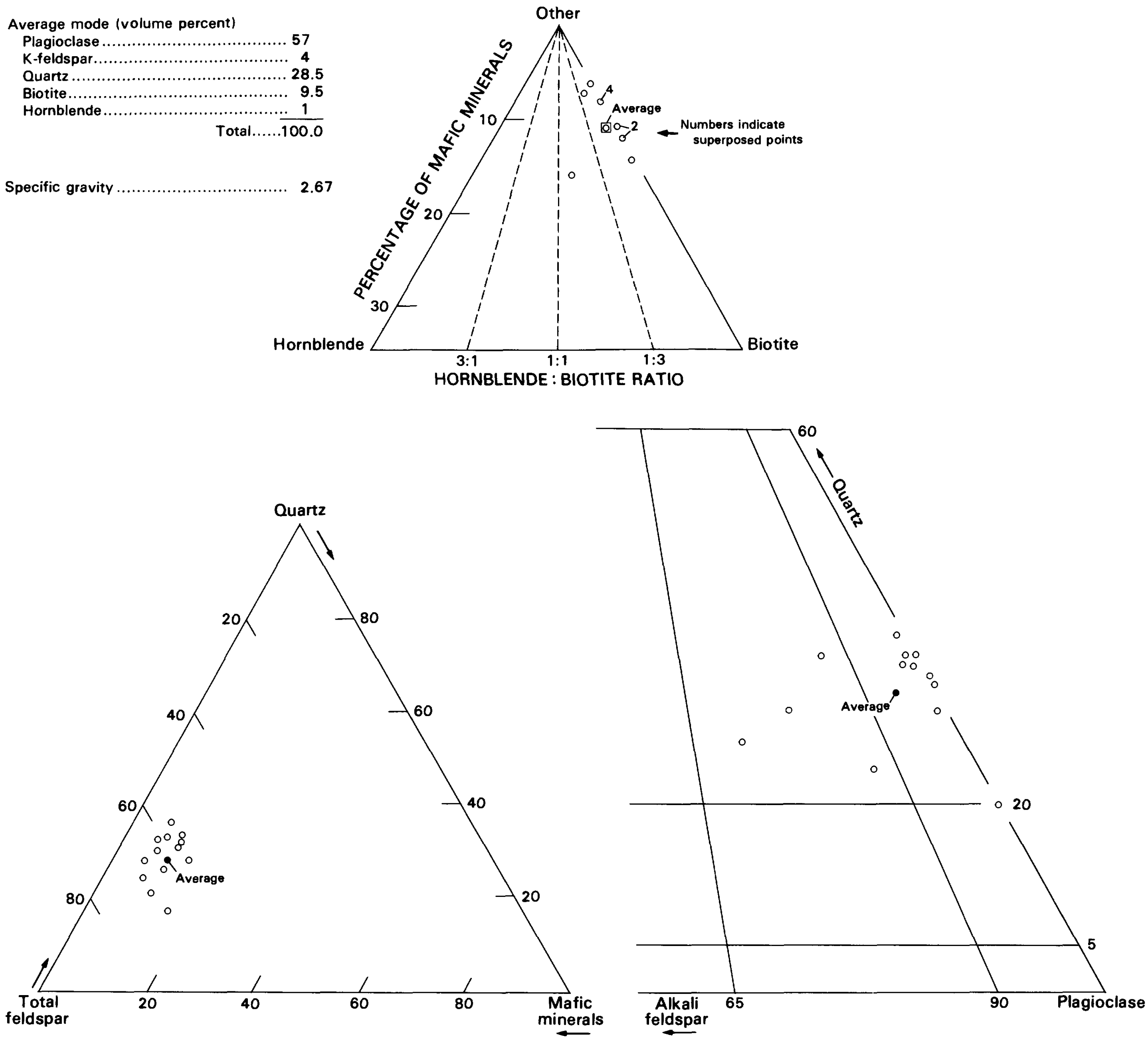


FIGURE 14.—Modal plots of the finer grained facies(?) of the tonalite of Mount Adelaide.

TABLE 8.—*Modes of the finer grained facies(?) of the tonalite of Mount Adelaide*

[All modes in volume percent. Others: O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3866	51	8	33	8	---	---	2.63
3866-1	62	---	30	8	---	---	2.70
3868A	61	1	27	11	---	---	2.67
38688	56	1	35	8	---	---	2.66
3868-1	58	1	33	8	---	---	2.69
4160	56	<1	32	11	<1	1(O)	2.65
4221	67	---	17	9	7	---	2.76
4248	57	<1	28.5	14	.5	---	2.65
4251	56	1	31	12	---	---	2.63
4268	52	14	28	6	---	---	2.66
4270	58	9	21	10	1	1(O,S)	2.71
4286	48	20	25	6	1	---	2.67
4421	56	1	31	12	---	---	2.67
Average--------	57	4	28.5	9.5	1	---	2.67
Standard deviation.	4.9	6.5	5.1	2.4	1.9	---	.04

contaminated by interaction with the gneissic and amphibolitic rocks of Jawbone Canyon. The choice is severely hampered by an absence of exposed contacts. I have placed the north contact with the Claraville mass arbitrarily between rocks containing abundant hornblende to the south and the hornblende-poor Claraville body on the north, but I have seen no dikes or other evidence to indicate that two intrusive masses are present. The south contact of the Hoffman Canyon unit with the gneissic and amphibolitic rocks is even more arbitrary, in a broad mixed zone of gneissic and amphibolitic rocks and various granitic rocks.

This problem cannot be resolved by reconnaissance mapping alone; further work is needed to determine whether, indeed, a Hoffman Canyon intrusive mass exists, or whether this unit is merely a contaminated facies of the Claraville mass. In any event, the dark gneissic and amphibolitic rocks of Jawbone Canyon seem to be somewhat similar to, if not correlative with, the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. The Hoffman Canyon mass contains gneissic zones, schlieren, and inclusion swarms that suggest the kind of contamination which is common in the tonalite of Bear Valley Springs. Whatever its genesis, the Hoffman Canyon mass is a mappable unit whose limits are generally as shown on plate 1; it may be a correlative of the Bear Valley Springs mass.

PETROGRAPHY

The outcrop appearance of the tonalite of Hoffman Canyon resembles that of the Bear Valley Springs mass. Both units contain abundant dark minerals with irregular anhedral crystals, and both units contain abundant dark inclusions. The tonalite of Hoffman Canyon is dominated by anhedral to subhedral, well-twinned andesine crystals. K-feldspar is untwinned and ranges in abundance from 0 to nearly 20 percent. Quartz is abundant and forms grains several millimeters across that are commonly mosaicked. Brown biotite and olive hornblende are abundant in clusters of anhedral grains. Clinopyroxene is present as cores in some hornblende crystals. Discrete muscovite crystals that look primary are found in some specimens. Lozenges of prehnite occur in some biotite plates, and blue tourmaline was noted in one specimen. Metallic opaque-mineral grains and sphene are the most common accessories, with minor grains of apatite and zircon.

Although the modal average for the tonalite of Hoffman Canyon contains more K-feldspar and has a higher biotite/hornblende ratio than the average for the Bear Valley Springs mass, the range of values for each mineral is nearly identical (fig. 15; table 9). Individual samples of the Hoffman Canyon unit are identical modally to those of the Bear Valley Springs unit, and the modal average of the two units is also similar.

The dark gneissic and amphibolitic belt south of the tonalite of Hoffman Canyon contains conspicuous dark to light, quartzofeldspathic gneissic layers that appear to be invaded by tonalite, and the gneiss is partly interleaved with tonalite. Along Jawbone Canyon are large areas of dark rocks that resemble the inclusion material common in the tonalite of Hoffman Canyon. It is not difficult to imagine the abundant inclusions being derived from the belt of dark gneissic and amphibolitic rocks. The dark rocks are dominated by fresh, well-twinned plagioclase of about An_{50} and moderate green hornblende. Brown biotite is less common, and minor quartz is present. Opaque material is common, both as irregular grains and as schillerlike inclusions in hornblende. Some of these rocks are weakly porphyritic and contain small subhedral plagioclase crystals. Less common are dark rocks spotted by clusters of green hornblende that may be relicts of volcanic phenocrysts. The dark gneissic and amphibolitic belt almost surely has a metaigneous parentage. Such parentage suggests a correlation with the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. So far, only one occurrence of the coarse-haloed garnets has been found, in a hybrid rock associated with metamorphic rocks near the west end of the Hoffman Canyon mass; that single occurrence contrasts with the widespread abundance of coarse-haloed garnets in the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains.

The tonalite of Hoffman Canyon generally is somewhat finer grained than much of the Bear Valley Springs mass.

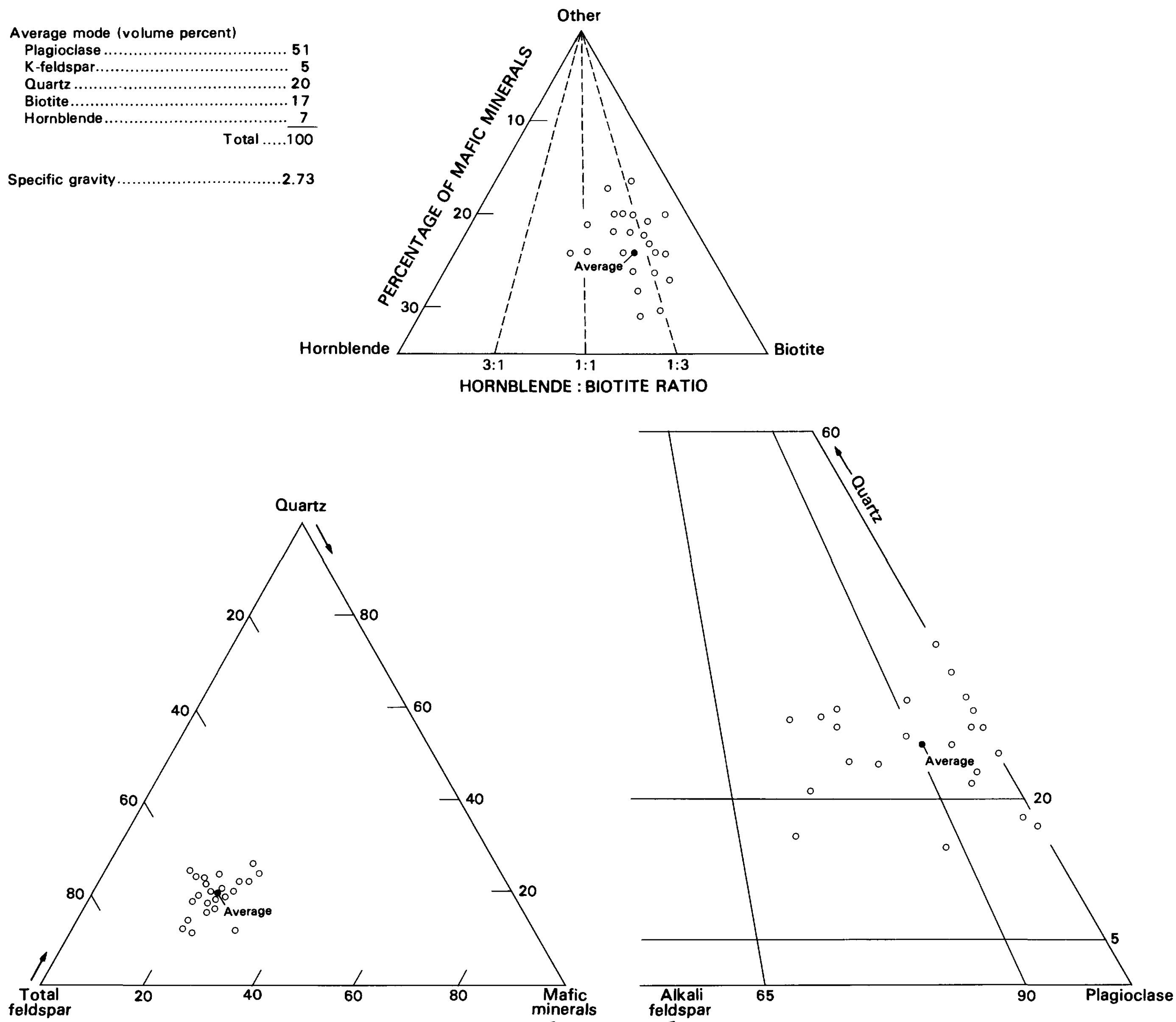


FIGURE 15.—Modal plots of the tonalite of Hoffman Canyon.

TABLE 9.—*Modes of the granodiorite of Hoffman Canyon*

[All modes in volume percent. tr, trace; n.d., not determined. Others: C, clinopyroxene; O, opaque minerals; S, sphene.

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3842A	59	3	18	15	5	---	2.74
3843A	45	12	23	15	5	---	2.70
3844	49	5	24	14	8	---	2.70
3845	48	11	25	13	3	---	2.70
3846A	46	11	22	17	4	---	2.73
3849A	58	7.5	11.5	18	5	---	2.77
4379	56	2	17	11	13	1(S)	2.78
4382	51	6	21	16	6	---	2.73
4390	52	3	19	18	8	---	2.73
4392	54	1	21	18	2	4(C)	2.72
4393	57	---	19	12.5	11	.5(O)	n.d.
4395	57	---	12	22	9	---	2.71
4398B	64	1	14	17	4	---	2.73
4400	49	9	18	16	8	---	2.73
4401	44	15	24	11	6	---	2.71
4509A	47	19	13	11	10	---	2.72
4509B	45	15	16	19	5	---	2.69
4536	51	<1	22	20	4	3(C)	n.d.
4542	49	---	22.5	19.5	9	---	n.d.
4554	53	---	20.5	20	6.5	---	2.78
4555A	46	tr	27	22	5	---	2.75
4555B	46	---	24	22	8	---	n.d.
4557	49	12	19	13	7	---	2.74
Average--------	51	5	20	17	7	---	2.73
Standard deviation.	5.4	5.9	4.3	3.6	2.7	---	.03

In addition, the foliation of the Hoffman Canyon mass, shown by aligned inclusions and gneissic layers, is gentle to near-horizontal (at least along Hoffman Canyon). These features suggest that the tonalite of Hoffman Canyon may represent a near-roof part of a granitic body. The original configuration of the Hoffman Canyon mass and its possible original connection with the presumably deeper Bear Valley Springs mass (based on coarser grain size and steep foliation) is now obliterated by younger granitic rocks.

GRANITE OF SADDLE SPRING ROAD

A small (8-km^2 area) granitic pluton lies between metasedimentary and metavolcanic rocks at the north edge of the study area northeast of the Walker Basin. These granitic rocks extend northward out of the study area and are also exposed within the belt of metasedimentary rocks and gneiss that adjoins the Saddle Spring Road body on the southwest. In outcrop, these rocks are generally fine grained, varicolored in shades of gray, and characterized by abundant small, rounded dark clots and inclusions, as much as a few centimeters long.

Thin-section study shows a strikingly bimodal texture, with ragged plagioclase crystals, as much as 3 mm long, in a much finer grained matrix, mainly of quartz and feldspar. Although this matrix may reflect a rehealed mortar texture, its general appearance is more reminiscent of the texture of a hypabyssal rock or a quenched groundmass. This texture may represent chilling at a pluton margin, although such chilling is unusual in this region.

Some of the larger plagioclase crystals are well twinned and weakly zoned in the sodic to intermediate andesine range. Grid-twinned K-feldspar occurs in the matrix and varies greatly in abundance. Quartz is abundant and commonly is mosaicked and in broken angular grains, a texture suggesting that the matrix may be cataclastic. Brown biotite and pale- to dark-green hornblende are common in shreddy aggregates and in large anhedral crystals. Concentrations of these dark minerals form the clots and small inclusions that are common in this pluton. Zircon, sphene, and apatite are present, but metallic opaque material is rare.

GRANODIORITE OF LEBEC

SETTING

Crowell (1952) first named the Lebec Quartz Monzonite for exposures on both sides of Grapevine Creek north of the Garlock fault near Lebec. This unit, here informally designated the granodiorite of Lebec, extends eastward from Crowell's original type area and is truncated by the Garlock fault; it also extends about 30 km westward, where it is cut off by the San Andreas fault. In its present form, the body covers an area of about 115 km^2. The Lebec mass has engulfed and included large volumes of metasedimentary rock (see subsection above entitled "Metasedimentary Rocks of Salt Creek"), which consists largely of quartzofeldspathic schist, impure quartzite, and calcareous rocks.

Petrography

The Lebec mass (fig. 16; table 10) consists of peppery biotite granodiorite that contains small, sparse mafic inclusions. Locally it is porphyritic and contains stubby, poikilitic K-feldspar crystals, as large as 15 mm across. These K-feldspar crystals range in size from large phenocrysts down to small interstitial grains giving the rock a seriate texture. Some outcrops contain distinctive round, blue grains, as large as 5 mm across. Scattered coarse biotite flakes and red-stained cores of hornblende crystals are also conspicuous in some outcrops. The Lebec mass is generally homogeneous and is free from obvious wallrock contamination, except locally.

In thin section, the most distinctive feature of the Lebec unit is the occurrence of euhedral to subhedral plagioclase crystals of sodic andesine that show well-developed oscillatory zoning. K-feldspar occurs locally in irregular interstitial grains but, more commonly, in discrete subhedral and euhedral poikilitic crystals that include all other minerals. Reddish-brown biotite, peppered with zircon-cored pleochroic haloes and apatite prisms, occurs in discrete books that are as large as the quartz and feldspar grains.

Hornblende, in part in euhedral prisms, is uncommon

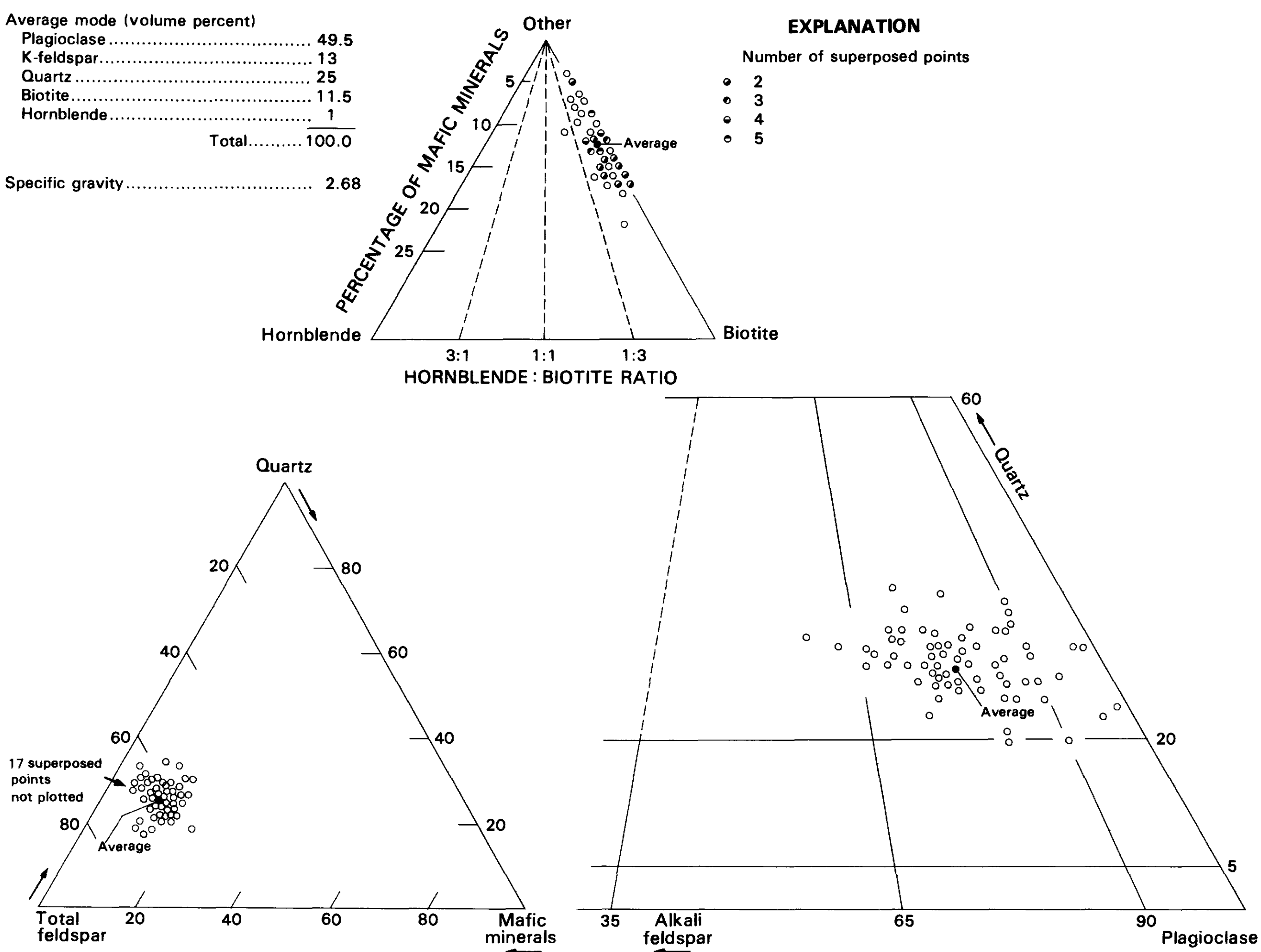


FIGURE 16.—Modal plots of the granodiorite of Lebec.

TABLE 10.—*Modes of the granodiorite of Lebec*

[All modes in volume percent; n.d., not determined. Others: A, allanite; C, clinopyroxene; M, muscovite; O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
643	50	19	20	11	---	<1(M)	2.69
660	49	15	25	11	---	---	2.62
673	48	15	25	11	1	<1(M)	2.67
674	48	16	21	14	1	---	2.72
679A	43	21	28	7	1	<1(M)	2.65
680	47	15	28	10	<1	<1(A,M)	2.68
682	44	22	29	5	---	<1(M)	2.64
683	53	6	27	14	---	<1(A)	2.67
686	45	18	26	10	1	<1(M)	2.68
687	49	12	27	11	1	<1(A)	2.68
688	51	12	22	15	---	<1(M)	2.71
690	48	15	27	8	2	---	2.67
691	34	26	28	11	1	---	2.66
692	48	14	26	11	1	<1(C)	2.67
693	49	10	27	14	---	---	2.68
694	47	14	23	13	3	<1(C)	2.71
695	47	16	25	10	2	---	2.65
696	41	19	24	13	3	<1(C,M)	2.68
697	49	12	26	13	---	<1(A)	2.68
698	41	17	27	13	2	---	2.68
699	57	4	24	15	---	<1(A)	2.69
700	49	13	22	14	2	<1(C,M)	2.69
701	49	6	30	14	1	---	2.63
702	52	9	24	12	1	2(C)	2.70
706	58	2	18	19	3	---	2.69
707	52	9	23	14	2	<1(C)	2.70
712	47	20	28	5	---	---	2.65
713	48	16	26	10	---	---	2.67
779	52	7	29	12	---	---	2.68
780	60	1	27	12	<1	<1(C), 1(M)	2.71
FM-1	47	12	24	17	<1	---	2.71
3047	56	6	26	10	2	---	2.70
3048	62	1	20	16	1	---	2.72
3053	49	12	22	16	1	<1(A)	2.70
3054	48	6	30	15	1	<1(S)	2.70
3056	45.5	11.5	34	9	---	---	n.d.
3069	41	21	25	10	3	<1(S)	2.66
3078	44	17	30	8	1	---	2.67
3079	50	14	24	10	2	<1(A)	2.68
3081	56.5	9	21	10	3.5	---	2.69
3086	43	18	30	9	---	<1(A,S)	2.65
3088	58	12	18	7	4	1(S)	2.70
3132	50	15	25	9	---	1(S)	2.65
3138A	32	24	36	8	---	---	2.64
3142	63	7	17	12	1	<1(A,O,S)	2.68
3145	40	15	33	12	---	---	2.63
3148	56	2	26	16	---	<1(A)	2.68
3162	46	14	27	11	2	<1(A)	2.66
3181	51.5	8	27	11.5	2	---	2.67
3185	51	7.5	28	12.5	1	<1(A)	2.68
3190	50	11	30	9	---	---	2.68
3193	47	13	27	12	1	<1(A)	2.68
3195	53	10	21	14	2	<1(A)	2.68
3197	45	11	26	17	1	---	2.65
3203	44.5	15	28.5	10	2	---	2.63
3204	52	9	24	13	2	---	2.67
3208	45	17	33	4	---	---	2.66
3211	44	19	30	6	1	<1(A)	2.64
3212	58	7	21	11.5	2.5	<1(A)	2.685
3213	56.5	11.5	18	13	1	---	2.70
3217	40	23.5	29	7.5	<1	---	2.65
3222	48	15	30	7	<1	<1(S)	2.66
3262	54	7	22	17	---	---	2.69
3263	52	10	25	12	1	---	2.69
3888	47	16	23	12	2	<1(A)	2.68
3890	45.5	17	23	11.5	3	---	2.68
3891	54	7	22	15	2	<1(A)	2.69
Average--------	49.5	13	25	11.5	1	---	2.68
Standard deviation.	7.3	6.9	4.7	3.6	1.1	---	.02

but widespread in these rocks; much of it is pale green, although locally it has the deep-olive-green shades typical of many granitic masses. The red cores in hornblende consist of iron-oxide-alteration material associated with lacy remnants of clinopyroxene. The Lebec mass also contains discrete (primary?) muscovite flakes in most specimens. The most distinctive accessory mineral is reddish-brown allanite, which occurs in euhedral zoned crystals, as large as 1 mm. In addition, zircon, apatite, and sphene are present in most sections, but metallic opaque-mineral grains are notably rare.

Alteration of these rocks varies considerably; some are virtually unaltered, whereas others contain completely chloritized biotite and strongly saussuritized plagioclase. Some of the biotite appears to have been bleached, with the resulting development of droplets of sphene and some opaque material, but no chlorite. Some of the more intense alteration is accompanied by shearing and granulation of the rocks, and some of the outcrops near the San Andreas fault show intense shearing and a conspicuous anastamosing fabric. However, it is surprising how fresh and structurally undisturbed are some outcrops that are relatively close to the fault, partly because of the preservation of lozenges of relatively fresh rock in otherwise-sheared outcrops.

Fine-Grained Facies

Sugary, fine-grained granitic rocks are found throughout the Lebec mass, but they are more common in its western part (pl. 1). From field relations, it was unclear whether or not these rocks are a fine-grained facies of the Lebec mass. Some fine-grained dikes intrusive into the Lebec mass are similar to the fine-grained rocks. Petrographic study and modal analysis (fig. 17; table 11) suggest that the fine-grained facies is a more felsic, later pulse and not a fine-grained equivalent of the Lebec mass.

The fine-grained facies contains considerably less biotite than the granodiorite of Lebec, and very rare hornblende. On average, the two facies also contain considerably different proportions of quartz and feldspar. Nevertheless, a near-overlap in modal percentages occurs between individual samples of the granodiorite of Lebec and the fine-grained facies. The fine-grained rocks probably represent a somewhat more differentiated and later pulse of the Lebec magma, although the compositional gradations of both units suggest a close relation between the two facies.

Felsic Marginal Facies

Locally within the Lebec mass and generally near the margin of the mass are felsic rocks (fig. 18; table 12) that have about the same grain size as the Lebec mass. These rocks present much the same dilemma as the fine-grained facies; that is, the relations of these felsic rocks to the average Lebec rocks are unclear. These felsic marginal rocks are best exposed near the north end of the San Emigdio fault; they are in an unusual setting to be a late-crystallizing part of the Lebec mass. Most probably, they represent a local late magmatic pulse, as do the fine-grained rocks, although, as yet, no evidence has been found that the felsic marginal rocks intrude the Lebec mass.

GRANODIORITE OF CLARAVILLE

Setting

The granodiorite of Claraville nearly equals the tonalite of Bear Valley Springs in size, covering an area of about 635 km^2. Within the study area, outcrops of the Claraville mass terminate against the Sierra Nevada fault on the east, the Garlock fault on the south, and a nearly continuous septum of metamorphic rocks on the west that separates the Claraville body from the tonalite of Bear Valley Springs. Early in the study of this area, three presumably separate granitic masses were mapped. These masses are in the Whiterock Creek area north of Tehachapi, in Lone Tree Canyon north of Mojave, and near Claraville northwest of Kelso Valley. With continued mapping, there was no obvious evidence for more than one intrusive mass. A separation can be made, however, between areas of dominantly porphyritic rocks and areas of nonporphyritic rocks (fig. 19) that contain somewhat more biotite and hornblende. Tentatively, these nonporphyritic and somewhat darker rocks are considered to be the Whiterock facies of the Claraville body, although they could, in fact, be a separate intrusive body.[2]

The Claraville body probably intrudes the tonalite of Bear Valley Springs, but the contact is either largely masked by metamorphic rocks or poorly exposed. The tonalite of Hoffman Canyon and the associated amphibolitic and gneissic rocks of Jawbone Canyon may be a large inclusion or enclave of the tonalite of Bear Valley Springs, and the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains that were engulfed by the Claraville pluton or the Hoffman Canyon mass could be a contaminated facies of the Claraville body, formed by interaction of magma with the gneissic and amphibolitic rocks of Jawbone Canyon.

West of Lone Tree Canyon, Dibblee (1959) noted that the rock is "* * * hydrothermally altered to hard coherent but cavernous-weathering masses." He interpreted these masses as part of the (Claraville) granitic mass in Lone Tree Canyon. Indeed, in some places the Claraville mass has its mafic minerals extensively leached and even destroyed, and the rocks are stained orange by iron oxides. However, most of the craggy, cavernous-weathering outcrops have a distinctive mineralogy that is richer in quartz and K-feldspar than those of the Claraville body, a difference suggesting that the cavernous-weathering rocks are younger intrusions into the Claraville mass. In

[2]Geologic mapping north of lat 35°30′ N., since the preparation of this report, has shown that the granodiorite of Claraville is the southern part of a large porphyritic pluton which I have named the "granodiorite of Castle Rock." Probably, this body extends as far north as lat 36°20′ N., where it is called the granite of White Mountain (du Bray and Dellinger, 1981).

TABLE 11.—*Modes of the fine-grained facies(?) of the granodiorite of Lebec*

[All modes in volume percent]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Specific gravity
721	42	31	23	4	---	2.63
722	41	31	22	6	---	2.65
3115	33	26	39	2	---	2.62
3118	48	13	34	5	---	2.66
3119	42	21	34	3	<1	2.63
3125	30.5	30	37	2.5	<1	2.60
3127	44	22	31	3	---	2.62
3136	38	26	32	4	---	2.62
3164	48	20	28	4	---	2.655
3167	43	25	27	5	<1	2.64
3169	46	21	28	5	---	2.65
3186	31	38	29	2	---	2.61
3225	45	18	33	4	---	2.61
Average--------	41	25	30	4	---	2.63
Standard deviation.	6.0	6.6	5.0	1.3	---	.02

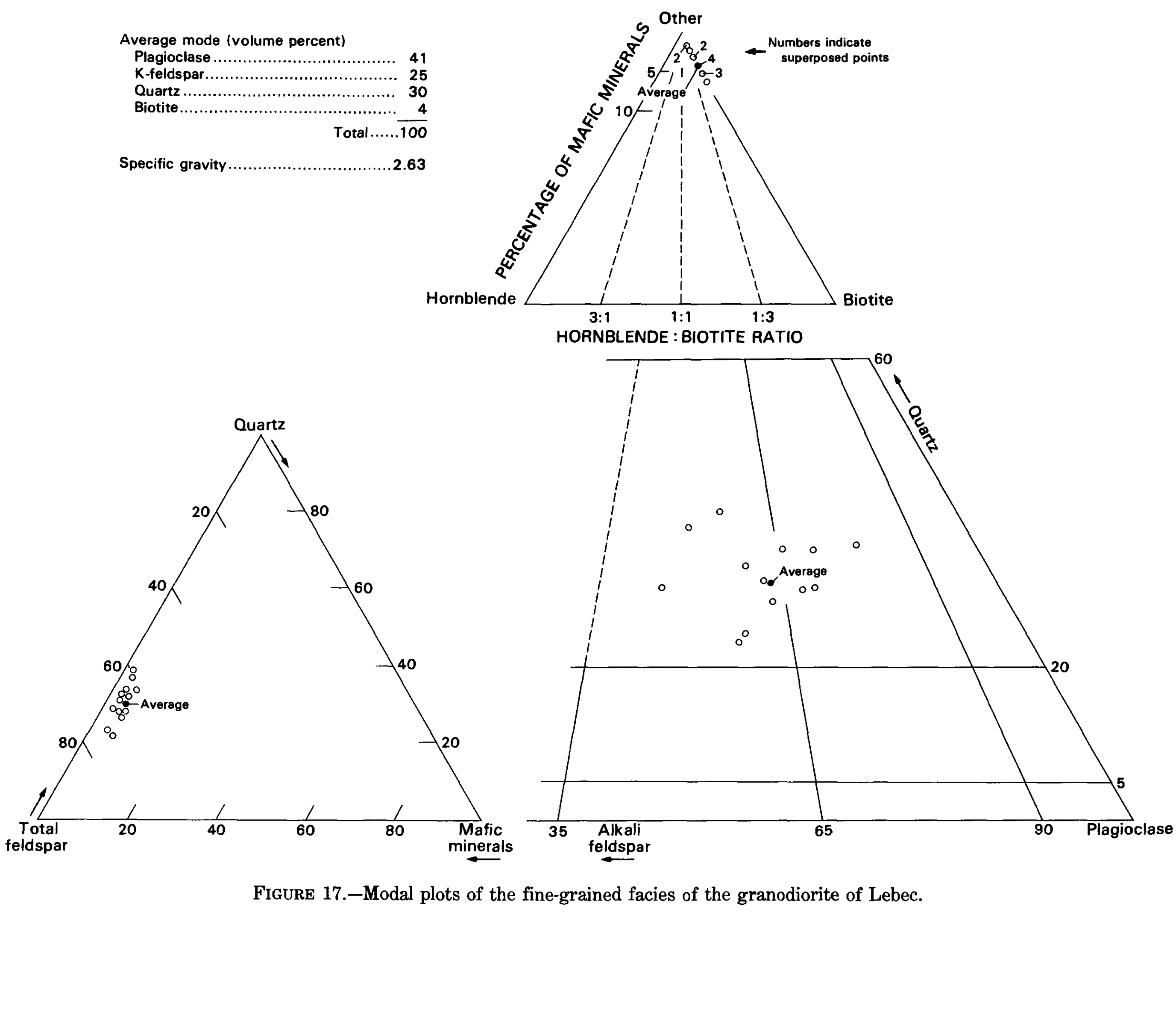


FIGURE 17.—Modal plots of the fine-grained facies of the granodiorite of Lebec.

TABLE 12.—*Modes of the felsic marginal facies(?) of the granodiorite of Lebec*

[All modes in volume percent. Others: M, muscovite]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Others	Specific gravity
678	33	28	34	5	<1(M)	2.64
3010	33	22	42	3	---	2.63
3012A	31	28	37	4	---	2.62
3138A	32	24	36	8	---	2.64
3879	29.5	31	36	3.5	---	2.59
3880	29	34	31	6	---	2.60
Average--------	31	28	36	5	---	2.62
Standard deviation.	1.7	4.4	3.6	1.9	---	.02

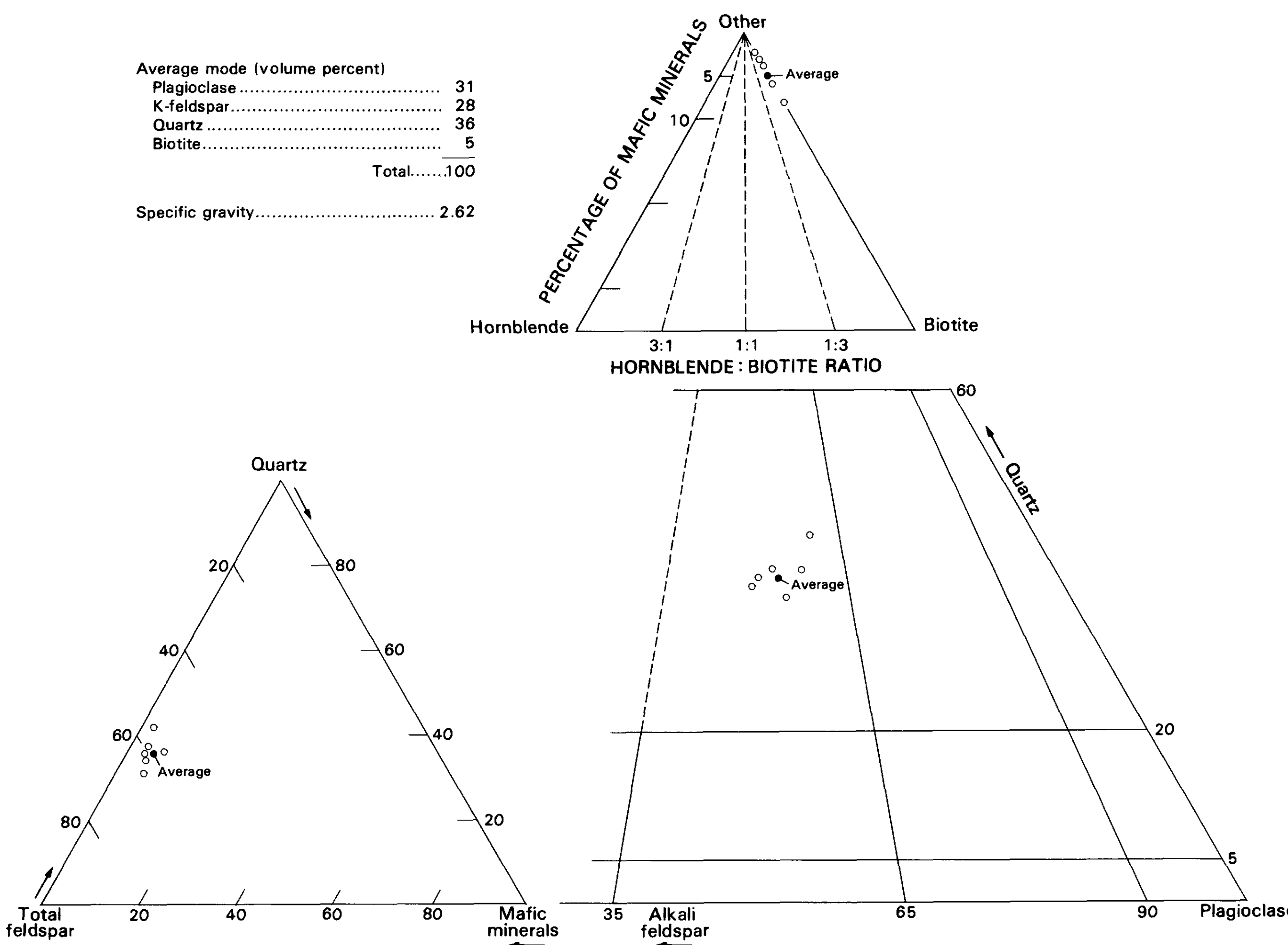


FIGURE 18.—Modal plots of the felsic marginal facies(?) of the granodiorite of Lebec.

at least two localities, craggy orange rocks intrude the Claraville mass. These younger plugs, which are extensively altered, are presumably part of the granite of the Bishop Ranch.

Petrography

The Claraville mass (fig. 20; table 13) is freshest and best exposed north and west of Kelso Valley. It has a peppery texture characterized by a liberal sprinkling of small biotite crystals and lesser hornblende. Scattered ellipsoidal, dark, fine-grained inclusions, as much as a few centimeters long, are also characteristic. Subhedral phenocrysts of pink K-feldspar, as much as 4 cm long, are particularly conspicuous in the northern part of the body. A few specimens contain scattered euhedral hornblende crystals, as much as 1 cm long. The Claraville mass generally has a xenomorphic granular texture.

Tabular anhedral to euhedral plagioclase grains form a major part of the rock; they are well twinned and exhibit oscillatory zoning in the range of sodic to intermediate andesine. K-feldspar varies greatly in abundance; it occurs in anhedral untwinned grains that in part are interstitial. As previously noted, some of the K-feldspar forms subhedral to euhedral phenocrysts. Quartz is generally abundant and mostly occurs in anhedral grains. Strain in quartz is indicated by undulatory grains, mosaicked masses, and local granulation. Biotite is generally dark brown or olive; some is reddish brown. Chloritization of biotite is widespread in this mass, and in some specimens no original biotite remains. Dark-green hornblende, though not everywhere present, is a recurring accessory mineral. What appears to be primary muscovite occurs in discrete flakes that make up as much as 1 percent of some specimens. Metallic opaque-mineral grains and sphene are abundant in these rocks; allanite, zircon, and apatite are also present.

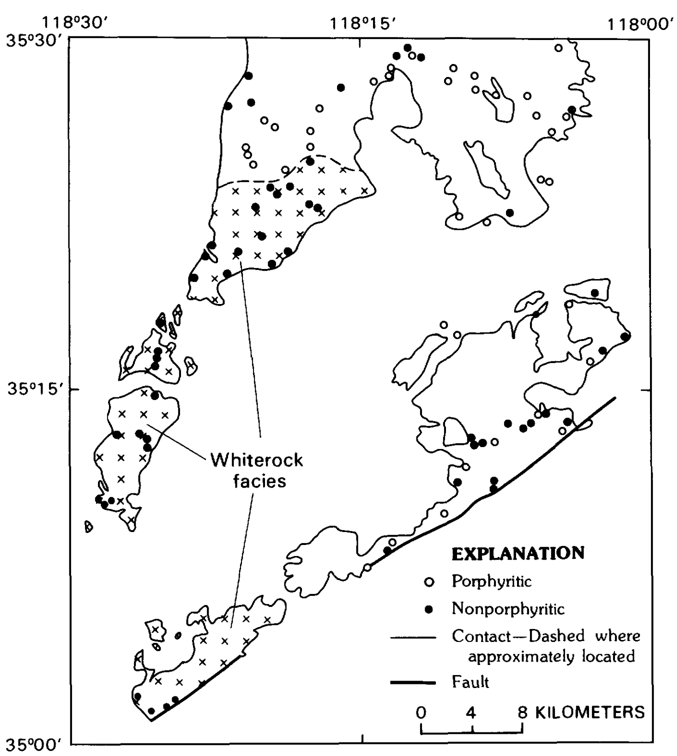


FIGURE 19.—Tentative distribution of the Whiterock facies of the granodiorite of Claraville on the basis of the distribution of porphyritic and nonporphyritic samples.

K-feldspar in the Claraville pluton does not have an even distribution. A north-south line drawn at about long 118°10′ W. separates most of the specimens that contain 20 percent or more K-feldspar to the east of the line, from most of those that contain less than 20 percent K-feldspar to the west. This uneven distribution may reflect the Whiterock facies to the west. Some specimens west of the line contain 24 to 28 percent K-feldspar, and some east of the line contain only 2 to 5 percent K-feldspar, but generally K-feldspar is more abundant to the east. The Whiterock facies is petrographically much like the rest of the Claraville body, but the Whiterock facies contains more dark minerals and generally less K-feldspar (fig. 21; table 14). Both the main Claraville body and the Whiterock facies, however, vary considerably in mineral content.

Orange-weathering, hydrothermally altered rocks of the Claraville mass are most common near Lone Tree Canyon and are similar in appearance to outcrops of the craggy weathering granite of Bishop Ranch. Dark minerals have been completely replaced by pseudomorphs composed of chlorite, epidote, calcite, and sphene globules. In some rocks, only iron oxide clots remain as evidence of former dark minerals. Clouded, clayey to strongly saussuritized plagioclase is also common. Sphene is altered to leucoxene. Some of the altered rocks are strongly cataclastic.

In Jawbone Canyon, particularly near Hoffman Canyon, alteration of the granitic rocks is common. Previously, these rocks had been considered to be brecciated rocks in a north-south-trending shear zone. Their appearance and distribution, however, suggest that they were, instead, hydrothermally altered, either during intrusion of the granite of the Bishop Ranch or, more likely, as the result of nearby Tertiary volcanic activity.

Strongly hydrothermally altered rocks of the Claraville mass are also found along the Garlock fault. These dark rocks are in part cataclastic. They contain the same alteration products as the orange-weathering, hydrothermally altered rocks discussed above. I note that some of these rocks along the Garlock fault strongly resemble the green, altered granitic rocks that are present in a thick retrograde zone between upper-plate granitic rocks and lower-plate Rand Schist that marks a strongly compressional

TABLE 13.—*Modes of the granodiorite of Claraville*

[All modes in volume percent; n.d., not determined. Others: O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3848	43	26	24	7	---	---	2.64
3849B	52	9	29	10	---	---	2.67
4049FL	50	17	23	10	1	---	2.67
4051	55	7	25	10	1	1(O), 1(S)	2.68
4052	49	11	26	8	4	1(O), 1(S)	2.70
4059A	47	15	26	10	1	1(S)	2.66
4059B	42	18	32	7	<1	1(O)	2.66
4064	53	17	21	9	---	---	2.66
4065A	57	5	25	11	1	1(O,S)	2.71
4067	48	10	27	13	1	1(O)	2.68
4070	54	9	25	10	1	1(O)	2.66
4072	51	20	21.5	5.5	---	1(O), 1(S)	n.d.
4073	43.5	21	29	4.5	1	1(O)	2.60
4079	39	22	30	7	1	1(O)	2.64
4080	49	17	28	5	---	1(O), 1(S)	2.65
4086	44	20	26	6	2	1(O), 1(S)	2.67
4087	50	19	25	5	1	---	2.65
4089A	42.5	25	29	3.5	---	---	2.62
4092	44	20	30	5	---	1(O), 1(S)	2.64
4093	48	15	24	10	2	1(O,S)	2.68
4093-1A	47	14	28	10	1	---	2.67
4094	44	23	26	6	1	---	2.66
4095B	48	12	22	17	1	---	2.67
4131FL	52	17	22	7	1	1(O,S)	2.66
4328	50	15	21	12	1	1(S)	2.68
4335	51	10	22	13	4	---	2.70
4336	54	16	18	10	2	<1(S)	2.71
4341	52	9	24	13	2	---	2.70
4343A	61	5	15	13	6	---	2.71
4345	59	9	18	11	3	---	2.69
4346	49	15	22	12	2	---	2.68
4348	60	6	21	11	2	---	2.68
4351	57	7	20	13	3	---	2.70
4352	43	24	26	7	<1	---	2.65
4353	51	18	19	10	2	<1(S)	2.71
4354	50	11	27	10	2	---	2.70
4356	61	8	14	11	4	2(S)	2.70
4359	60	6	13	17	2	2(S)	2.69
4360	56	12	23	8	---	1(S)	2.66
4363A	50	18	20	10	2	---	2.71
4365	51	16	18	10	4	1(S)	2.72
4384	55	3	28	13	---	1(O)	2.70
4386	62	3	14	14	6	1(S)	2.71
4387A	48	24	19	9	---	---	2.65
4387B	57	7	15	13	7	1(S)	2.69
4398A	45	15	29	11	---	---	2.64
4398C	54	8	27	10	1	---	2.68
4402	48	16	25	9	1	1(S)	2.69
4403	60	3	15	17	4	1(S)	2.74
4404	55	2	22	17	3	1(S)	2.69
4406	43	16	31	9	<1	1(S)	2.66
4409	51	10	28	9	1	1(S)	2.71
4411A	59	8	21	12	---	---	2.65
4413	40	29	27	4	---	---	2.61
4415	64	1	12	16	7	---	2.77
4417	59	2	18	13	7	1(S)	2.73
4470B	37	30	27	6	---	---	2.65
4470C	37	22	34	7	---	---	2.64
4472	54	10	25	10	---	1(S)	2.69
4473A	59	6	25	10	---	---	2.76
4477A	39	33	24	4	---	---	2.63
4485A	46	26	24	4	---	---	2.63
[1]4486	32.5	32.5	33	3	---	---	2.60
4487A	42	25	28	5	---	---	2.61
4489	55	8	20	17	---	---	2.72
4496B	51	7	29	13	---	---	2.69
4497A	53	2	28	17	---	---	2.72
4498C	57	4	25	14	---	---	2.71

TABLE 13.—*Modes of the granodiorite of Claraville*—Continued

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
4500	44	12	34	10	---	---	2.68
4505	39	23	32	6	---	---	2.64
4508	59	2	26	13	---	---	2.70
4518	61	8	26	10	---	---	2.65
4519	46	12	27	15	---	---	2.69
4528	41	27	25	7	---	---	2.66
4562	41	23	28	8	---	---	2.61
Average--------	50	15	23	10	1.5	.5	2.68
Standard deviation.	7.1	9.0	6.9	3.6	1.9	---	.04

[1]Altered rock.

thrust-fault zone in the western Rand Mountains (40 km east of Kelso Valley).

Just north of Kelso Valley is a large patch of coarse-grained, dark rocks that are distinct from and partly separable from those typical of the Claraville body. These darker rocks are distinguishable by more abundant dark minerals, particularly hornblende, and by mafic schlieren and inclusions. A small body of gabbroic rocks is present in this darker mass just north of Kelso Valley. On Piute Mountain Road west of Kelso Valley Road, a sharp contact is exposed between the darker, coarser grained rocks and typical peppery Claraville, but to the southeast along

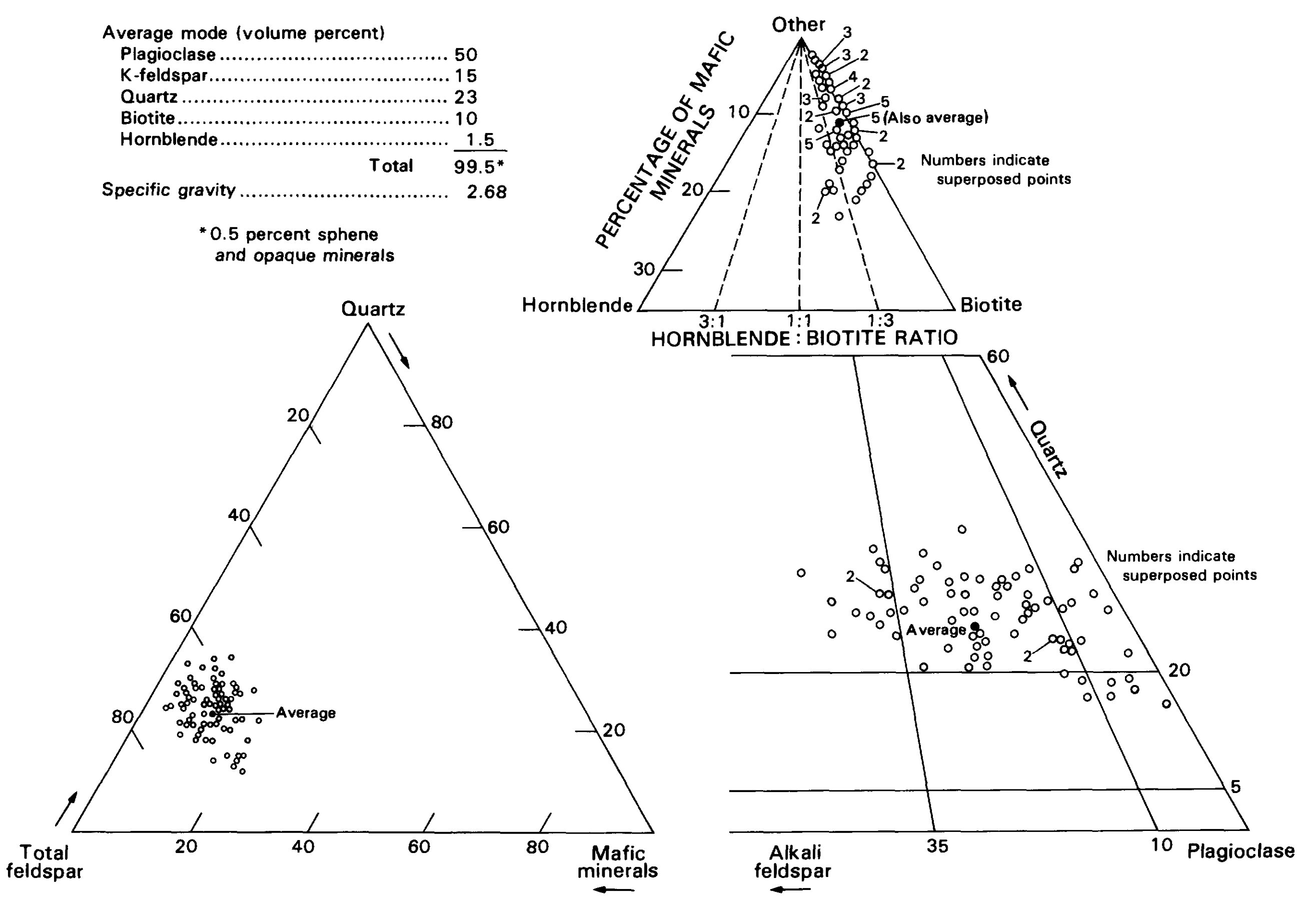


FIGURE 20.—Modal plots of the granodiorite of Claraville.

a road to Butterbredt Canyon the contact is gradational, and its location is uncertain. At first, I suspected that these coarse-grained, darker rocks are not a separate intrusive mass but are local areas of contamination of the Claraville mass by gabbroic rocks similar to those just north of Kelso Valley. Recent mapping north of the area of plate 1 suggests that the darker, coarser grained rocks are part of a separate and extensive mass in the Lake Isabella area.

GRANODIORITE OF SORRELL PEAK

Setting

The crestline of Sorrell Peak west of Kelso Valley is underlain by coarse-grained, yellow- to orange-weathering granitic rocks, much aplite-alaskite-pegmatite, and possible inclusions of the Claraville mass. Four additional small plugs(?) of the Sorrell Peak rock type are present to the south. The total outcrop area of this rock type is less than 5 km^2. About 2 km north of Sorrell Peak and about 2 km to the southwest, the Sorrell Peak mass sharply contacts the Claraville body. Although age relations are not obvious at these intrusive contacts, the rock distribution suggests that the coarser grained Sorrell Peak unit intrudes the Claraville body. These coarse-grained plugs appear to be somewhat more differentiated than the Claraville rock type. The association with abundant aplite-alaskite-pegmatite suggests that the coarse-grained plugs may be a late pulse of the Claraville mass.

Petrography

The Sorrell Peak and associated small masses (fig. 22; table 15) are locally somewhat altered and iron stained but are generally fresh. Coarse, anhedral grains of oligoclase-andesine, poorly twinned K-feldspar, and quartz dominate. Anhedral grains and clusters of light-olive to opaque biotite are common. The biotite is sprinkled with tiny crystals of apatite and zircon, and some anhedral

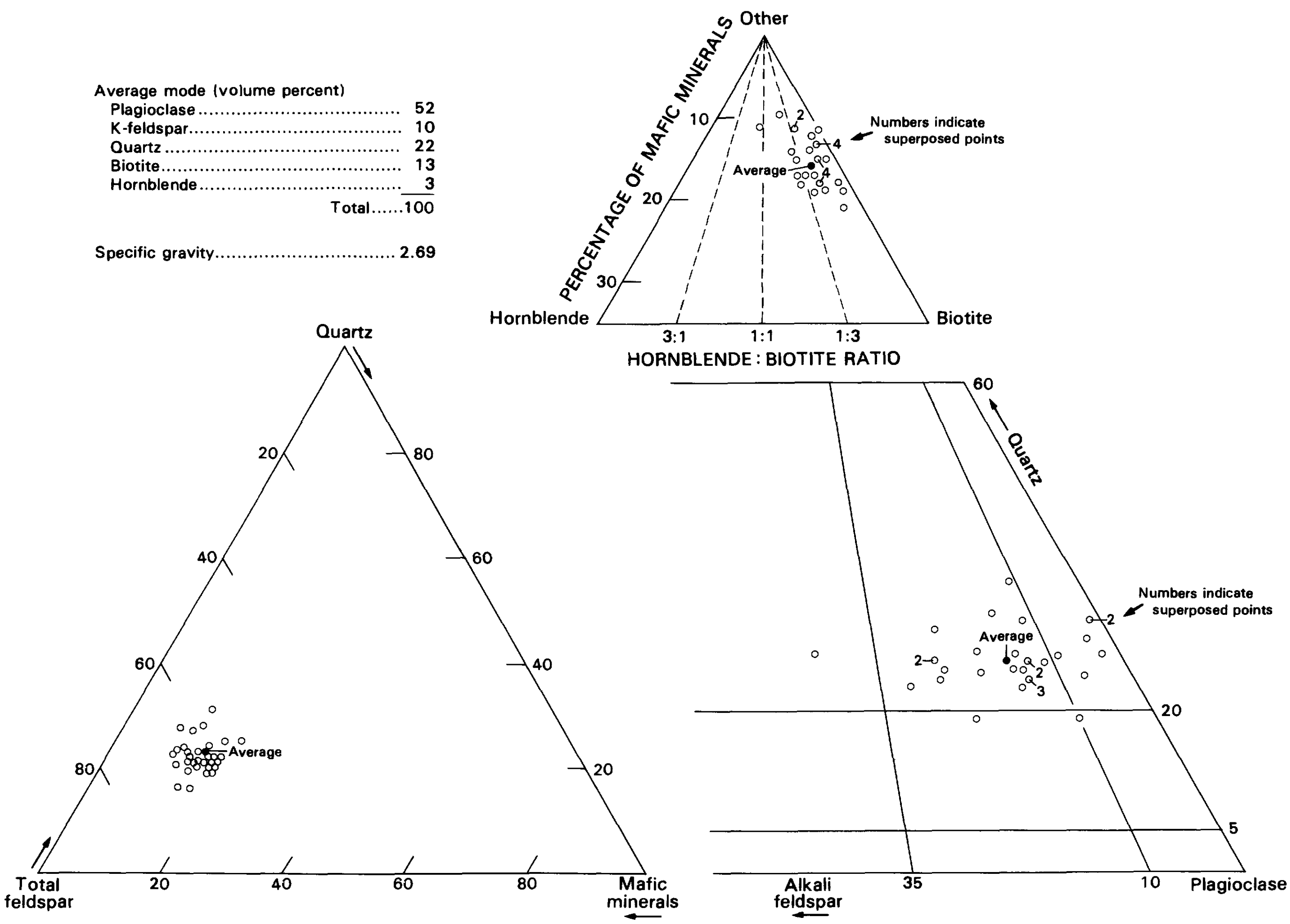


FIGURE 21.—Modal plots of the Whiterock facies of the granodiorite of Claraville.

TABLE 14.—*Modes of the Whiterock facies of the granodiorite of Claraville*

[All modes in volume percent; n.d., not determined. Others: O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
3735A	57	7	22	12	1	1(S)	2.64
3737	60	4	20	11	4	<1(O), 1(S)	2.64
3738A	53	9	20	15	3	<1(S)	2.65
3739A	36	28	24	5	6	1(S)	2.63
3805A	60	1	27	12	---	---	2.70
3806A	58	2	24	13	2	<1(O), 1(S)	2.74
4102A	56.5	9	23.5	9	2	---	2.67
4103A	49.5	9.5	28	12	1	---	2.68
4104B	47	16	27.5	6.5	3	---	2.67
4119F1	57	9	23	9	2	---	2.70
4126	55	1	25	15	3	1(S)	2.70
4309	53	9	23	13	2	<1(S)	2.69
4310	43	20	19	13	4	1(S)	2.70
4312	53	9	21	14	3	<1(S)	2.69
4314	51	16	16	12	5	<1(S)	2.70
4370	53	11	23	12	1	<1(S)	2.70
4371	57	2	22	15	4	<1(S)	2.72
4374	56	9	21	12	2	---	2.69
4376A	50	6	31	12	1	<1(S)	2.69
4447	45	16	21	15	3	---	2.70
4448	47	16	21	14	1	1(S)	2.72
4449	45	16	19	18	1	1(S)	2.71
4454	55	5	22	17	1	---	2.73
4467	51	13	21	13	2	---	2.69
4531A	44	16.5	21.5	15	3	---	2.70
4534	51	10	20	16	3	---	2.73
4538	48	12	22	13	5	---	2.73
4541	48	6	25	19	2	---	2.71
4565-1	57	10	20.5	11.5	1	<1(S)	n.d.
4568-1	55.5	10	19.5	13	2	<1(S)	n.d.
4569A	62	7	16	10	4	1(S)	n.d.
Average--------	52	10	22	13	3	<1	2.69
Standard deviation.	5.9	6.0	3.3	3.0	1.4	---	.03

grains of sphene. About half of the specimens examined contain small amounts of olive hornblende. Dark-brown, zoned allanite crystals are also present, as are sparse grains of primary(?) muscovite. The five small plutons of the granodiorite of Sorrell Peak are undoubtedly correlative. In addition, some of the felsic rocks in the Bear Valley Springs mass, particularly the small body east of Caliente along the Caliente Creek Road, may also be related to the Sorrell Peak type.

GRANITE OF TEHACHAPI AIRPORT

SETTING

The small isolated hill on the north side of the Tehachapi Airport exposes hummocky, bouldery slopes of tan felsic rocks that contain distinctive small accessory pink garnets. Similar rocks are exposed discontinuously for about 25 km to the north-northeast of the Tehachapi Airport hill, in a belt as much as 5 km wide. These felsic rocks appear to intrude the center of the large Claraville mass, but all exposed contacts of the Tehachapi Airport mass are with metasedimentary rocks of the Keene area. The northernmost outcrops have domed up the adjacent metasedimentary wallrocks, a relation suggesting that the present exposures are near the original roof of the Tehachapi Airport mass.

PETROGRAPHY

The Tehachapi Airport body (fig. 23; table 16) is modally similar to the granite of Bishop Ranch. The two units are distinguishable only by the invariable presence of accessory pink garnets in the Tehachapi Airport mass. The Tehachapi Airport rocks are also much less altered.

The Tehachapi Airport mass is dominated by grid-twinned K-feldspar. Coarse grains of quartz and thinly twinned oligoclase are also abundant. Brown biotite occurs only as scattered flakes in this pluton. Some biotite crystals contain zircon inclusions with pleochroic haloes. Small primary muscovite crystals are also present. Pink

TABLE 15.—*Modes of the granodiorite of Sorrell Peak*

[All modes in volume percent. tr, trace; n.d., not determined]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Specific gravity
4363B	35	26	31	8	---	2.63
4366	38	19	33	9	1	2.66
4368	34	30	30	6	---	2.64
4369	38	26	26	9	1	2.65
4372	40	22	28	10	---	2.61
4373	43	19	26	12	---	2.66
4376B	40	23	28	9	tr	2.66
4377	42	19	23	13	3	2.68
4381	39	22	24	14	1	2.66
4567	46.5	18	24.5	10	1	n.d.
4570	46	18	25	11	---	n.d.
Average--------	40	22	27	10	1	2.65
Standard deviation.	4.0	4.0	3.2	2.3	.9	.02

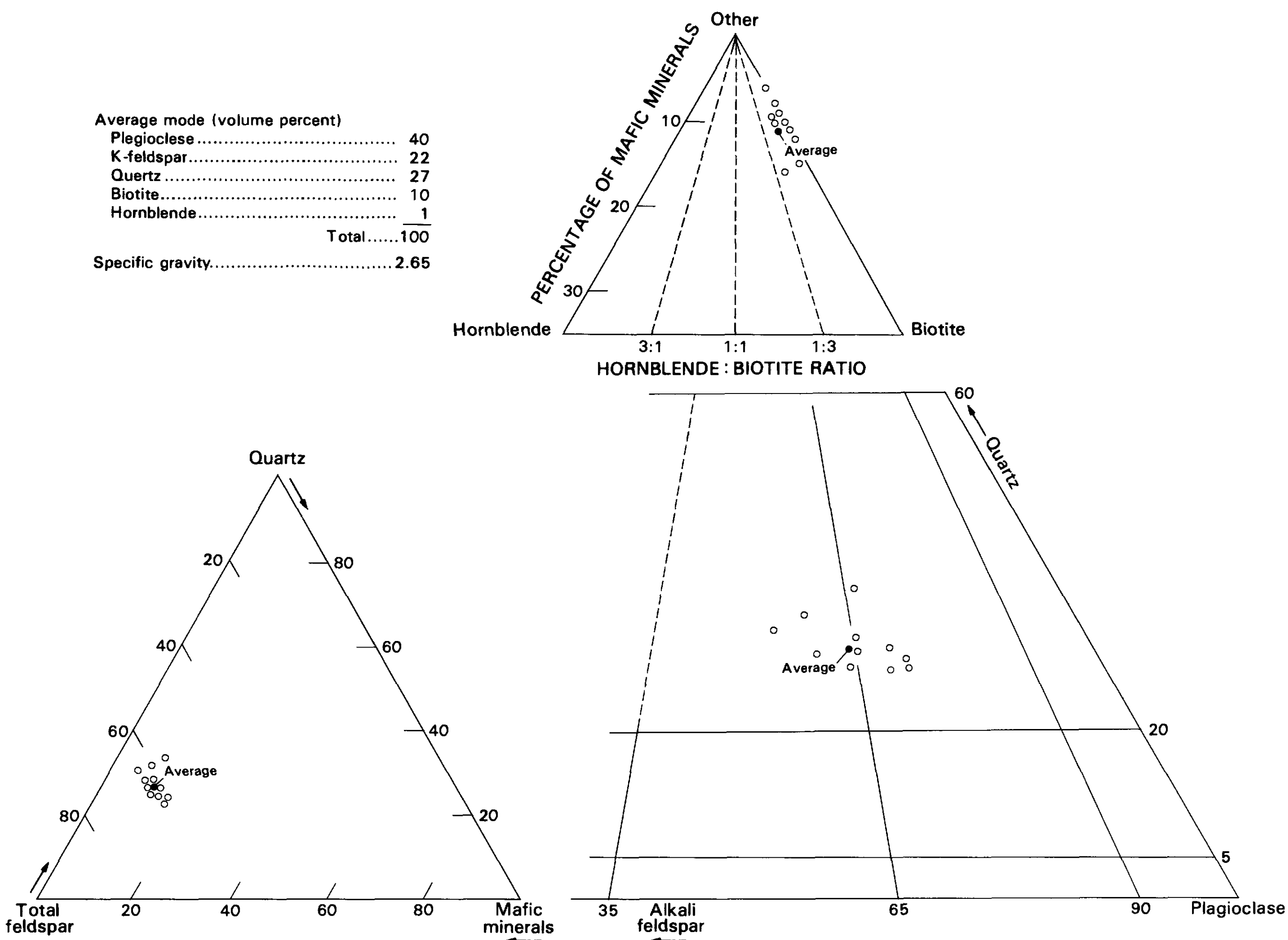


FIGURE 22.—Modal plots of the granodiorite of Sorrell Peak.

TABLE 16.—*Modes of the granite of Tehachapi Airport*

[All modes in volume percent]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Garnet	Specific gravity
3804A	26	40	32	---	2	2.55
4097	31	32	35	2	<1	2.61
4098	28.5	38.5	30.5	2	.5	2.61
4100A	22	42.5	33	1.5	1	2.60
4101	25	38	35	2	<1	2.58
4106	27	36	34.5	1.5	1	2.59
Average--------	26.5	38	33	1.5	1	2.59
Standard deviation.	3.1	3.6	1.8	.8	---	.02

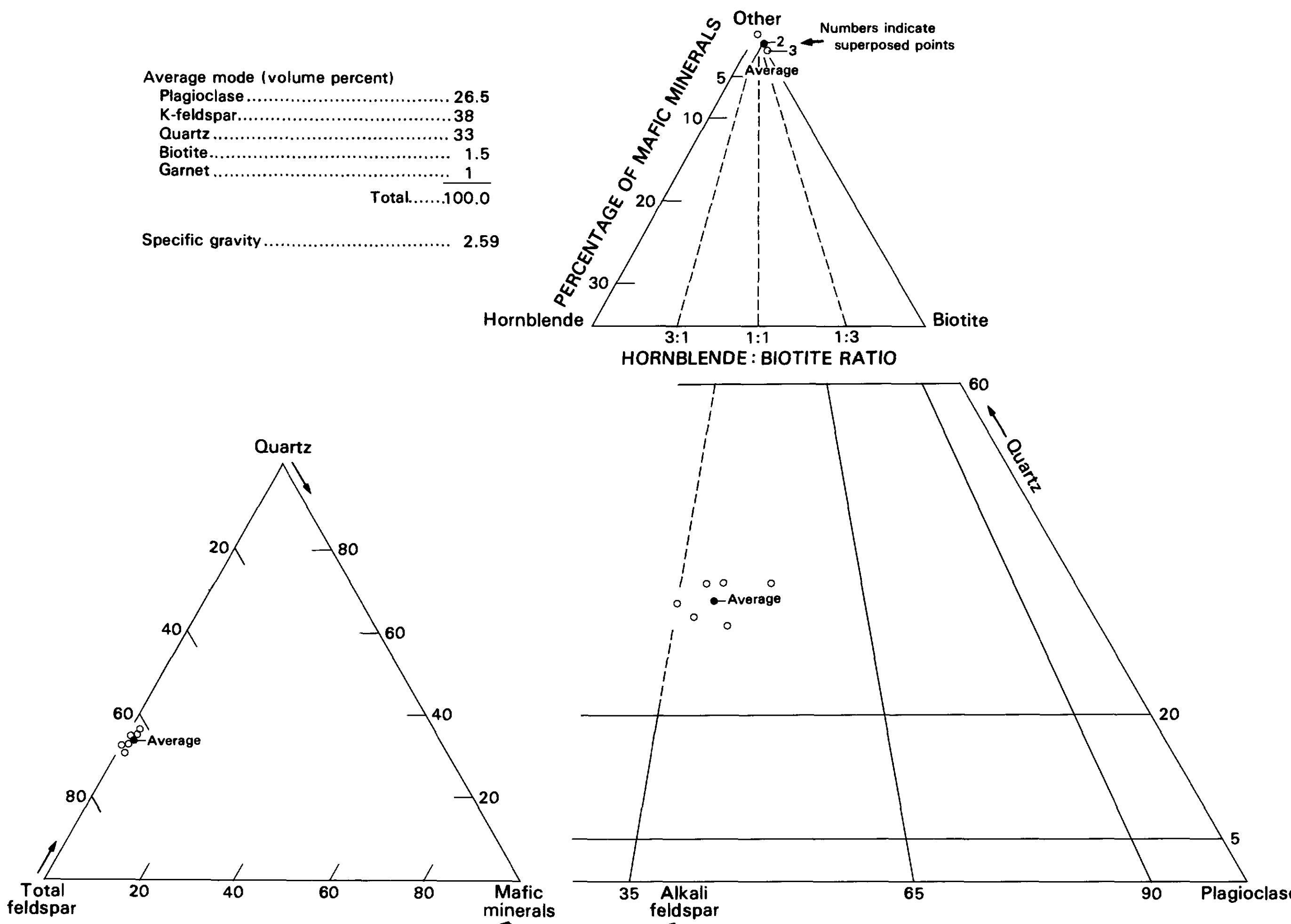


FIGURE 23.—Modal plots of the granite of Tehachapi Airport.

garnet occurs in tiny rounded crystals that are scattered but widespread.

GRANITE OF LONE TREE CANYON

Setting

A small plug of fine-grained biotite granite, about 5 km^2 in area, intrudes the Claraville mass at the mouth of Lone Tree Canyon. North of this small plug, additional small plugs and dikes intrude the Claraville mass but are too small to show at the scale of the map (pl. 1). These fine-grained rocks are particularly abundant in the northeast corner of the Claraville mass and continue northward out of the study area. Similar rocks are exposed along an east-west-trending ridge along the south side of California Highway 178 just east of Onyx. The Onyx locality is about 21 km north of where Kelso Valley Road leaves the study area, a relation suggesting that significant amounts of this fine-grained rock type are present north of the study area.

Petrography

In general, the rocks of Lone Tree Canyon (fig. 24; table 17) have a fine-grained, sugary to aplitic texture. Plagioclase is generally subhedral to euhedral and in some specimens forms small phenocrysts as large as 5 mm across. K-feldspar and quartz also occur as rare megacrysts, as large as 1 cm across. The plagioclase is generally zoned in the range calcic oligoclase to sodic andesine; some calcic zones are accentuated by alteration products such as sericite. Both K-feldspar and quartz occur in irregular to rounded grains; the quartz in particular seems to be droplike and to accentuate the sugary texture. Small, anhedral, brown biotite grains are scattered throughout the rocks. Some of the biotite is fresh, but in some specimens it is completely replaced by chlorite and iron oxides. Green hornblende is only locally and sparsely present. Metallic opaque-mineral grains are abundant in some of these felsic rocks. Allanite is also present, as well as sphene, zircon, and apatite.

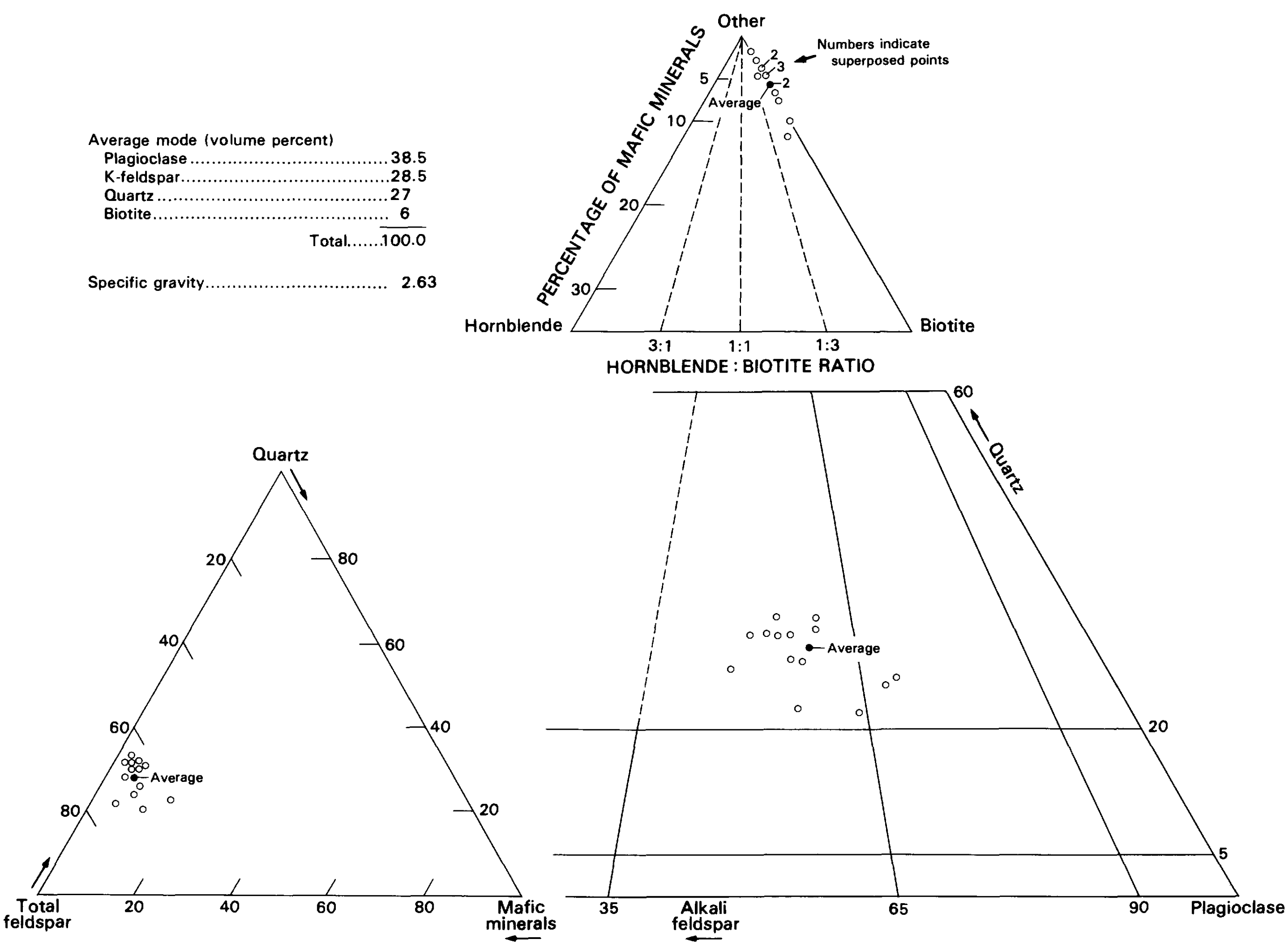


FIGURE 24.—Modal plots of the granite of Lone Tree Canyon.

TABLE 17.—*Modes of the granite of Lone Tree Canyon*

[All modes in volume percent]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Opaque minerals	Specific gravity
4076	34	31.5	29.5	4	1	<1	2.62
4077	35	30	29	6	<1	<1	2.63
4088	36.5	30	29.5	4	<1	<1	2.64
4090	47	22	23	8	---	<1	2.65
4091	38	30	27	4	---	1	2.64
4330	44	25	19	11	1	---	2.65
4339	47	19	24	10	<1	---	2.63
4470A	39	25	30	6	---	---	2.66
4478B	42	32	21	5	---	---	2.63
4481	35	30	32	3	---	---	2.62
4557A	33	34	31	2	---	---	2.60
4558	38	26	31	5	---	---	2.62
4563A	32	36	25	7	---	---	2.62
[1]Onyx-1	39	29	27	5	---	---	2.65
Average--------	38.5	28.5	27	6	---	---	2.63
Standard deviation.	4.8	4.6	4.0	2.6	---	---	.02

[1]North of study area.

Two specimens of fine-grained rock in the Claraville mass west of Kelso Valley contain more dark minerals (10 and 12 percent) than the other fine-grained specimens, and both contain hornblende. Modes of these rocks resemble those of the Claraville mass, and so these rocks may represent local quenched areas in the Claraville body rather than felsic later differentiates. Rocks intermediate in grain size between the Claraville and Lone Tree types occur in the northeastern part of the study area. In individual outcrops, these two units can be difficult to distinguish. I suspect that, at least in part, these finer grained rocks grade into the Claraville pluton with no sharply mappable contact.

GRANITE OF BRUSH MOUNTAIN

SETTING

At the extreme west end of the basement outcrop of the San Emigdio Mountains, an irregular mass of coarse-grained granite, the granite of Brush Mountain, underlies an area of about 8 km^2. This granite is surprisingly fresh in part and generally is homogeneous. It is truncated by the San Andreas fault.

East of the Brush Mountain mass, several bodies are present that are modally similar to the Brush Mountain mass, but they all are altered to form yellow- to orange-stained craggy outcrops. Although these bodies are almost wholly within the granodiorite of Lebec, they appear to intrude the mafic metamorphic rocks and the Rand Schist on the north flank of the San Emigdio Mountains. Dikes of the coarse-grained Brush Mountain rock type cut the Lebec mass, and this relation, together with the form of some of the bodies, leaves no doubt that the Brush Mountain mass is a young rock body intruding the Lebec mass. The bodies east of the Brush Mountain mass total about 20 km^2 in the outcrop area. The easternmost body east of Grapevine Canyon was named the "School Canyon Granite" by Crowell (1952). I believe that all the felsic masses from School Canyon westward to Brush Mountain form part of a single related granitic unit.

PETROGRAPHY

The following description deals chiefly with the relatively fresh westernmost body of the Brush Mountain type. The eastern masses have the same mineral content; they differ chiefly in that their biotite is partially to wholly altered to chlorite, muscovite, and opaque material, and their plagioclase is strongly clouded with alteration products.

The plagioclase is strikingly subhedral, well twinned, and generally about An_{30}. K-feldspar is strongly grid twinned, and most specimens show some patchy perthite. Some of the K-feldspar is also in subhedral grains, but most is anhedral. Quartz ranges in size from small round grains to large interstitial grains. Biotite, where fresh, is brown to locally reddish brown. Some muscovite occurs in well-formed books that may be primary, but most appears to be a secondary-alteration product of plagioclase and biotite. Zircon, as the cores of pleochroic haloes, and apatite are common inclusions in the biotite crystals. Opaque material is generally red, and most appears to be a product of biotite alteration.

The granite of Brush Mountain, which contains almost equal proportions of plagioclase, K-feldspar, and quartz, plots in the low-melting trough in the center of the modal triangle (fig. 25; table 18). Because such coarse-grained rocks are a common end member of many granitic suites, they are of only limited value for correlation.

GRANITE OF BISHOP RANCH

Setting

The granite of Bishop Ranch forms distinctive craggy, orange to salmon-pink outcrops. The two largest bodies of this granite are north of Jawbone Canyon and in the upper part of Lone Tree Canyon. An elongate body on the southeast side of Kelso Valley may be correlative with the other two masses, but I have examined that elongate body only locally along the road on the southeast side of Kelso Valley. Biotite and plagioclase in these granite bodies are hydrothermally altered to some extent. The distinction between the pink-colored granite of Bishop Ranch and hydrothermally altered rocks of the Claraville mass has been a continuing problem throughout the mapping of this region. The pink granite is characterized by an abundance of K-feldspar and by minor dark minerals (biotite). The hydrothermally altered rocks of the Claraville mass generally contain relicts of more abundant dark minerals and much less K-feldspar. However, without an examination of many outcrops, it is easy to confuse these two units, and so my reconnaissance mapping is suspect in some areas. The pink granite closely resembles the craggy, orange-weathering masses of the granite of Brush Mountain near Grapevine Canyon and to the west in the San Emigdio Mountains.

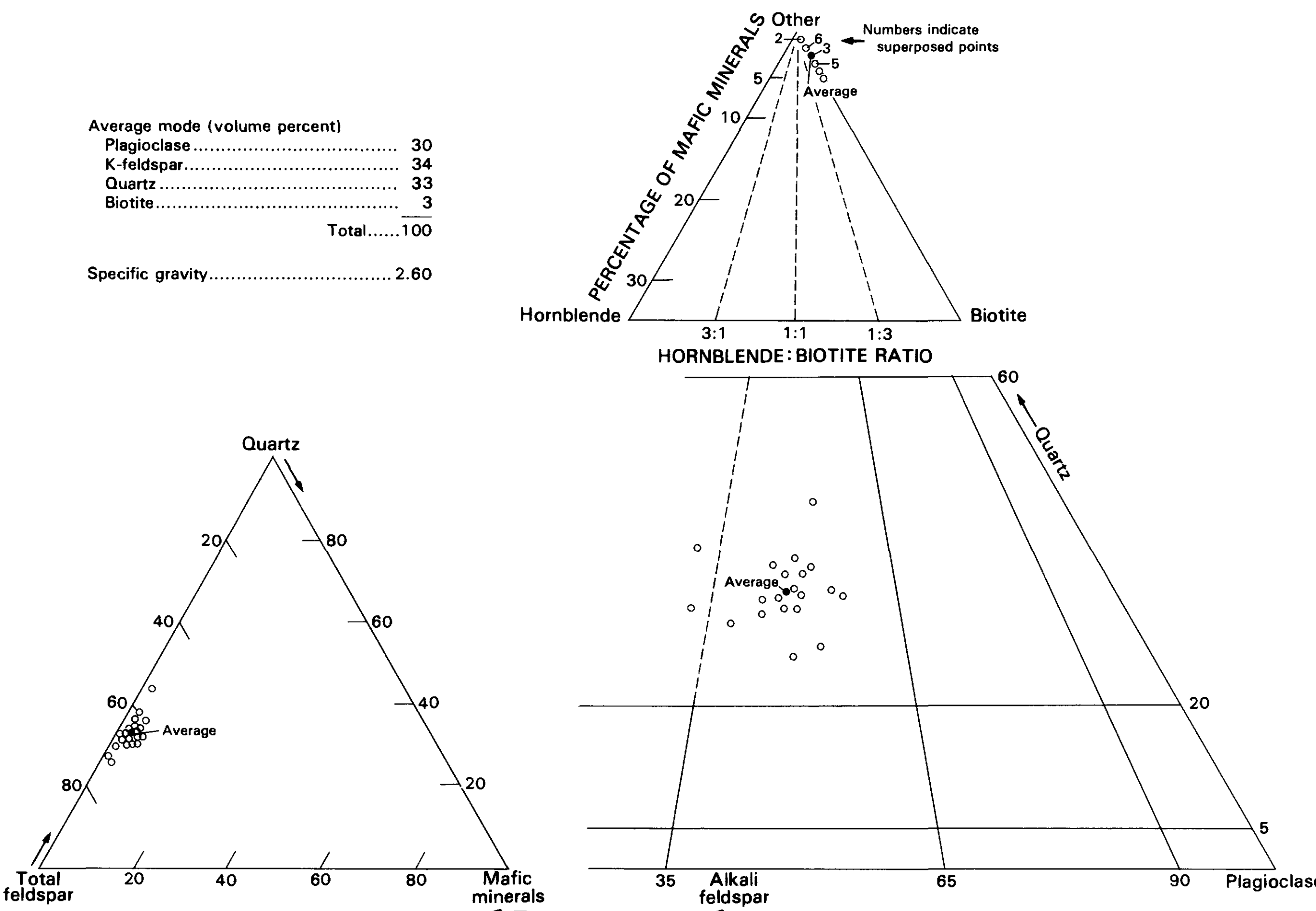


FIGURE 25.—Modal plots of the granite of Brush Mountain.

TABLE 18.—*Modes of the granite of Brush Mountain*

[All modes in volume percent; n.d., not determined. A, altered to chlorite]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Specific gravity
646	29	38	31	2	2.59
676	31	31	35	3	2.58
681	27	42	30	1	2.60
710	32	33	32	3	2.59
[1]726	28.5	31	36.5	4	2.62
3005	32	33	31	4	2.62
3035	35	37	26	2(A)	2.59
3037	28	26	44	2	2.62
3046	36	29	33	2(A)	2.60
3087A	38	34	27	1(A)	2.60
3089	34	28	32	6(A)	2.62
3102A	29	33	35	3	2.62
3110	30	35	31	4	2.62
3113	31	31	34	4	2.61
3130A	21	44	31	4	2.59
3146	19	41	38	2	2.60
3221	30	34	33	3	2.61
3881	28	36	31	5	n.d.
3882	28	34	36	2	2.56
3887	30	35	32	3	2.57
Average--------	30	34	33	3	2.60
Standard deviation.	4.5	4.6	3.9	1.3	.02

[1]Composite of samples 723 through 727.

PETROGRAPHY

The following description is based on a study of only four thin sections from the granite of Bishop Ranch. The rock is dominated by grid-twinned K-feldspar (fig. 26; table 19), and the mass southeast of Kelso Valley contains unusual wavy microperthite(?). Plagioclase is generally clouded, and although twinning is partly preserved, much is obscured by alteration, and so it is difficult to determine the anorthite content—some is oligoclase or sodic andesine. Quartz is also very abundant in this unit and shows mortar structure in some samples. Scattered brown biotite flakes are prèserved in some rocks, but most biotite is completely altered to red opaque material. Some of these rocks contain what appears to be primary muscovite. Also present are small scattered opaque-mineral grains and leucoxene pseudomorphs after sphene.

PLUTONIC ROCKS IN THE SUBSURFACE OF THE SOUTHEASTERN SAN JOAQUIN VALLEY

Several oil wells in the southeastern San Joaquin Valley have penetrated plutonic rocks. Although I have not had access to basement core material, the California Academy of Sciences loaned me several thin sections that were used by May and Hewitt (1948) in a previous study of basement rocks of the San Joaquin Valley. In addition, I had access to several drill logs that provided very brief (generally one word) descriptions of the basement rocks (J.A. Bartow, written commun., 1978). My data were previously published (Ross, 1979).

DIORITE AND GABBRO

Five drill holes penetrated mafic plutonic rocks, all at about the latitude of outcrops of the Pampa Schist. The hole nearest to surface outcrops (lat 35°27′ N., long 118°54′ W.) penetrated diorite, according to the drill log. The next hole to the west (lat 35°27′ N., long 119°09′ W.) is described in the drill log as penetrating "* * * greenish dark gabbro with 40 percent plagioclase and 60 percent augite, hornblende, and biotite."

Three holes farther west, at about lat 35°27′ N., long 119°15′ W., penetrated gabbroic rocks. Thin sections are available from these three drill holes. In the larger diorite and gabbro area on the map (pl. 1), the rocks feature bladed to stubby, well-formed crystals of sodic labradorite that are relatively fresh and sharply twinned, but fractured. In part, the bladed labradorite crystals are set in large, pale clinopyroxene crystals that give the rock an ophitic texture. The clinopyroxene is much altered to fibrous, pale-green amphibole. Chlorite and epidote are also present.

The remaining hole, farthest west from the surface outcrops, penetrated a sugary, gabbroic rock type composed chiefly of fresh, well-twinned sodic labradorite and pale-brown clinopyroxene. The rock also contains abundant pale-green chlorite, pale-green amphibole in veinlets, and abundant scattered opaque-mineral grains and sphene. From the texture, it is difficult to decide whether this rock is plutonic or metamorphic. Its composition and the nearby presence of definite gabbroic rocks of similar composition (described in the preceding paragraph) suggest that it is related to the gabbro.

From these limited data, I suggest that the five gabbro samples are related to the greenschist and amphibolite described from the subsurface. The gabbro and those rocks are distinguishable arbitrarily on the basis of the premise that the gabbros retains more of its original plutonic features. The Pampa Schist, the gabbro bodies associated with it, and the subsurface greenschist, amphibolite, and gabbro are probably all parts of a single related terrane.

HORNBLENDE-BIOTITE TONALITE

Along the north margin of the study area (long 118°51′ W.), one well penetrated granitic rock composed of moderate-green hornblende; fresh, sharply twinned andesine; coarse, virtually unstrained quartz; and chlorite (penninite) after biotite. A thin section of this rock is tiny and fragmented; it is impossible to determine that the sample does not represent regolith on top of the basement.

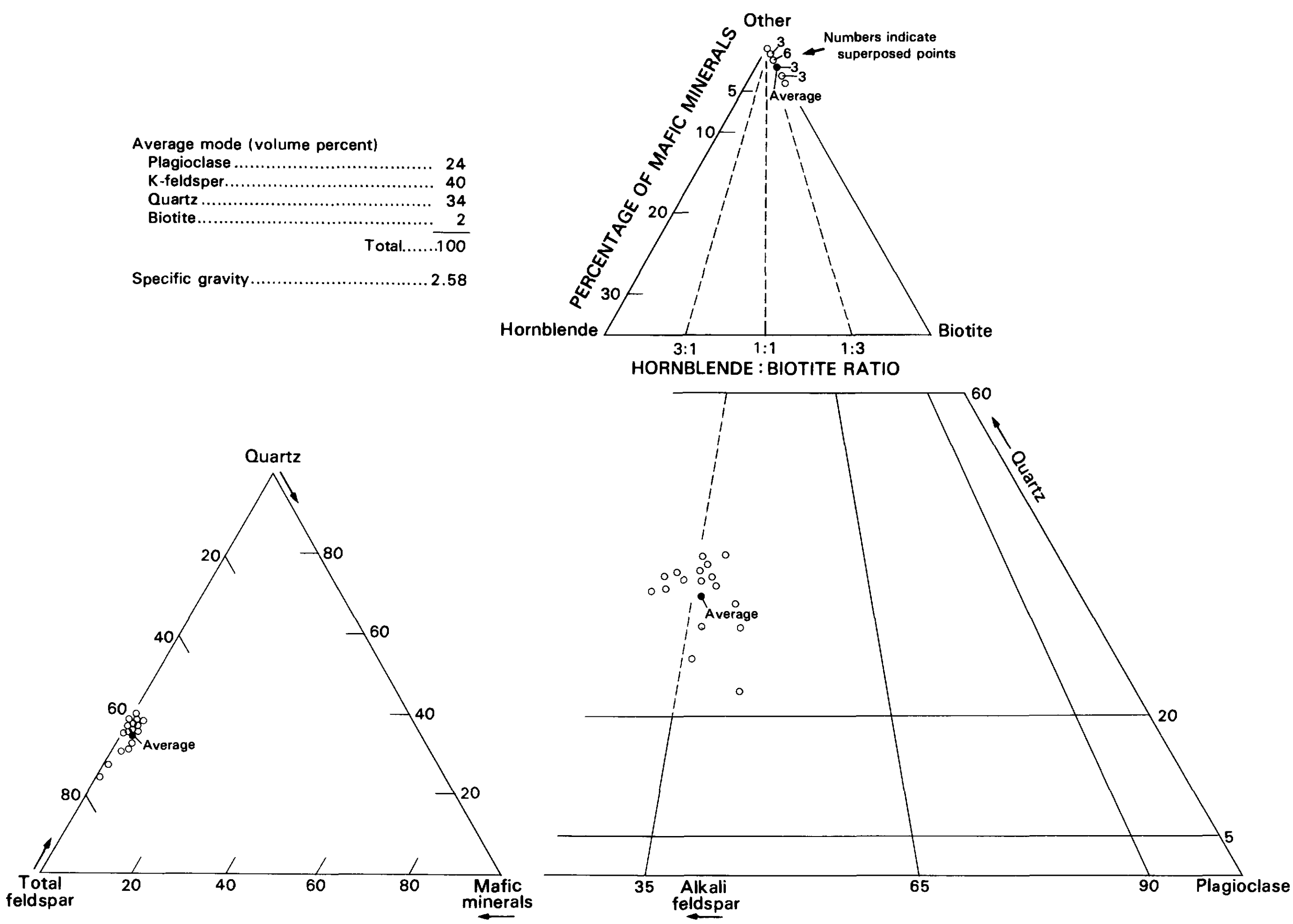


FIGURE 26.—Modal plots of the granite of Bishop Ranch.

TABLE 19.—*Modes of the granite of Bishop Ranch*

[All modes in volume percent]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Opaque minerals	Specific gravity
3798A	33	42	23	1	1	2.54
3799	22	39	39	<1	---	2.49
3843B	18	45	35	2	---	2.60
3851	23	39	36	2	---	2.55
4053	25	42	30	3	---	2.61
4053FL	29	38	30	3	---	2.59
4068	20	43	35	2	---	2.61
4075A	20.5	42	36.5	.5	.5	2.61
4085B	25	38	36	1	---	2.58
4411B	27	36	32	4	1	2.61
4414	22	39	37	1	1	2.59
4473C	24	38	37	.5	.5	2.60
4474	19	43	37	1	---	2.55
4475	27	45	27	1	---	2.59
4476	21	38	38	3	---	2.60
4478A	24	36	39	.5	.5	2.61
4563C	21	41	36	1	1	2.61
Average--------	24	40	34	2	<1	2.58
Standard deviation.	3.9	2.9	4.5	1.1	---	.03

However, the location of the hole and the general mineralogy of the thin section suggest that the hole either penetrated the tonalite of Bear Valley Springs or bottomed in regolith very close to inplace tonalite.

Two additional holes (lat 35°17′ N., long 118°46′ W.) penetrated hornblende-bearing granitic rocks (samples 90 and 102 of Ross, 1979) within the large subsurface area of biotite tonalite that adjoins rocks of the Mount Adelaide mass along California Highway 58. The thin sections of these two samples show abundant well-twinned intermediate andesine, abundant quartz, pale- to moderate-green hornblende, and brown biotite. From mineralogy alone, these rocks could be assigned to the Bear Valley Springs mass, but the biotite and hornblende crystals are generally separate and discrete. This texture is much more characteristic of the Mount Adelaide rock type than of the (more messy) Bear Valley Springs rock type. Therefore, these two samples are considered to be parts of the Mount Adelaide mass.

BIOTITE TONALITE

An isolated bouldery outcrop (the Rock Pile) about 9 km northeast of Arvin and 12 drill holes provide data on two biotite tonalite bodies that underlie part of the San Joaquin Valley in the study area. The 12 thin sections available from the drill-hole basement material are all remarkably similar. They consist chiefly of well-twinned and locally impressively oscillatory-zoned andesine in generally subhedral crystals, very minor K-feldspar, and abundant quartz. The thin sections also contain discrete, in part equant crystals of brown biotite and traces of green hornblende. Present as accessory minerals in most sections are metallic opaque minerals, sphene, apatite, and zircon; allanite is present locally. Some of the sections show strong saussuritization of the plagioclase and chloritization of the biotite. These thin sections probably represent samples of the tonalite of Mount Adelaide; the discrete biotite crystals are the most diagnostic feature.

Three drill holes near the north boundary of the study area at about long 119° W. penetrated biotite tonalite that is also believed to correlate with the tonalite of Mount Adelaide. The subsurface extent of this body is partly delimited to the east and south by wells that penetrate other rocks, but its extent to the west and north is unknown.

The other mass of biotite tonalite (south of the Edison fault) is more clearly delimited. Outcrops of the Bear Valley Springs mass limit its extent to the east and south, and drill holes to metamorphic rocks further limit its extent to the south and west, south of the Edison fault. No extrapolation has been attempted to the north of the Edison fault because only one subsurface data point in biotite tonalite is available north of the fault, but most probably there is a continuous subcrop to connect with the large outcrop area of the Mount Adelaide mass to the northeast.

GRANITIC ROCKS, UNDIVIDED

South of the White Wolf fault, several wells penetrated granitic basement; however, drill logs with commonly one-word descriptions of rock types provide the only data for these holes. The basement material is referred to merely as "granite" or "granodiorite." Cuttings from one well in the largest mass of "granitic and metamorphic rocks undivided" (20 km south of Arvin) are described as "gneissic, greenish black coarse mica." Therefore, this large mass may contain elements of the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains, which crop out extensively only a few kilometers to the east and south.

North of the White Wolf fault (8 km north-northeast of Bakersfield), a drill hole penetrated rocks referred to in the drill log as "granitoid, gneissic, micaceous." That drill hole lies just south of the northernmost mass of biotite tonalite that is presumably correlative with the Mount Adelaide mass. The Mount Adelaide body is not gneissic, and so it seems most this well probably penetrated some representative of the amphibolite-greenschist terrane.

UNITS SOUTH OF THE GARLOCK FAULT

GRANODIORITE OF GATO-MONTES

Setting

Outcrops of the granodiorite of Gato-Montes extend as a narrow belt from Cameron Road westward to the San Andreas fault, where they are truncated. The Gato-Montes body is also truncated on the north by the Garlock fault. The total outcrop area is about 115 km^2. The Gato-Montes mass is intruded by the granite of Tejon Lookout throughout the belt of outcrop, and by smaller bodies of the granite of Bean Canyon near the east end of the outcrop belt. The relation between the Gato-Montes mass and the granodiorite of Cameron to the east is presently unknown. The Gato-Montes mass has a striking physical resemblance to the previously described Lebec mass. Their possible correlation is described below in the subsection "Possible Correlation of the Granodiorites of Lebec and Gato-Montes."

Fine-grained, gray aplitic rocks occur locally as dikes in the Gato-Montes mass. Similar rocks, whose relations to the Gato-Montes body are uncertain, are also found as small masses (too small to show on pl. 1). These fine-grained rocks closely resemble the fine-grained facies of the granodiorite of Lebec.

Petrography

The granodiorite of Gato-Montes (fig. 27; table 20) is a fine- to medium-grained granitic rock with a peppery texture that contains scattered small, ovoid mafic inclusions. It is locally porphyritic and contains K-feldspar phenocrysts, as much as 2 cm long. In some outcrops, rounded quartz grains, as much as 1 cm across, are conspicuous. Red-cored hornblende crystals are present in some exposures.

Plagioclase in this mass is generally well twinned; it forms in subhedral crystals that in part have oscillatory zoning in the sodic to intermediate andesine range. K-feldspar is generally untwinned to weakly grid twinned and ranges in size from small interstitial grains to discrete, blocky subhedral crystals, as large as 2 cm in diameter. Quartz ranges in size from small interstitial grains to discrete crystals, as much as 1 cm across, generally smaller. Biotite, which is generally brown and less commonly red, generally forms small, partly subhedral crystals. Hornblende is much less abundant than biotite, forms irregular to euhedral crystals, is generally olive to deep green, and locally bears lacy cores of remnant clinopyroxene stained red by iron oxides. Opaque minerals and sphene, particularly the former, are sparse, but zircon, apatite, and allanite are present in all specimens.

GRANODIORITE OF CAMERON

Setting

The granodiorite of Cameron extends along the south side of the Garlock fault eastward from Cameron Road for about 16 km. Its east and south sides are covered by younger sedimentary material. The presumed contact with the Gato-Montes mass on the west is in an alluviated valley, and no relations can be determined. The elongate belt of outcrop truncated against the Garlock fault on the north covers an area of about 23 km^2. Salmon-pink K-feldspar crystals are a striking feature of some outcrops.

Petrography

The Cameron mass (fig. 28; table 21) is rather coarse grained; many crystals range in size from 5 to 10 mm across. It therefore markedly contrasts with the finer grained Claraville mass to the north across the Garlock fault and with the finer grained Gato-Montes rocks to the west. Nevertheless, the modal averages of both the Claraville and Gato-Montes masses are very close to the modal average of the Cameron mass.

Plagioclase in the Cameron mass is finely twinned, in part with poorly developed oscillatory zoning. Many crystals are subhedral. Refractive-index determinations in oils suggest that the plagioclase is sodic andesine (approx An_{35}). K-feldspar is abundant in strongly grid twinned, generally large crystals that locally show string perthite. In contrast, the adjacent Gato-Montes and Claraville masses generally have untwinned or very weakly grid twinned K-feldspar. X-ray-diffraction patterns of strongly twinned K-feldspar from three Cameron samples and untwinned to weakly twinned K-feldspar from six samples of the Claraville body from outcrops near the Cameron mass show (with one exception) higher 2θ values for peaks 060 and $\bar{2}04$ for the Cameron samples (fig. 29). These limited data suggest that a difference in the structural state of the K-feldspar, reflected in the twinning, may be a correlation tool, but further work is needed.

Quartz in the Cameron mass occurs as large anhedral grains that are only weakly strained and exhibit minor undulatory extinction and mosaicking. Brown to opaque biotite is widespread and abundant. Green hornblende is less abundant than biotite but is present throughout the mass. Muscovite, apatite, sphene, zircon, and opaque minerals are present in trace amounts. Most of the accessory grains are included in biotite flakes.

GRANITE OF TEJON LOOKOUT

Setting

The granite of Tejon Lookout was first described and named by Crowell (1952) for exposures in the Lebec quadrangle. The granite mass extends from a point near the junction of the San Andreas and Garlock faults northeastward to Little Oak Canyon, a distance of about 35 km. A smaller mass of the same rock type is exposed farther east, about 12 km southeast of Tehachapi. The total outcrop area of the Tejon Lookout mass is about 95 km^2.

The easternmost body and the eastern part of the western elongate body contain many small masses of metamorphic rocks assigned to the Bean Canyon Formation. These masses appear to be inclusions and small pendants that are at least partly keeled down into the granitic rocks. However, the two largest areas of metamorphic rocks in the western body, which have unusual ameboid shapes and appear to lie on top of the granite, suggest that the pluton there has a relatively flat roof. Wiese (1950) noted that the contact between the granitic and metamorphic rocks "* * * is nearly a plane, sloping quite evenly southeast at an angle of 6° to 7°." The larger mass of metamorphic rocks (present mostly in the Lebec quadrangle) contains four small intrusive stocks or cupolas

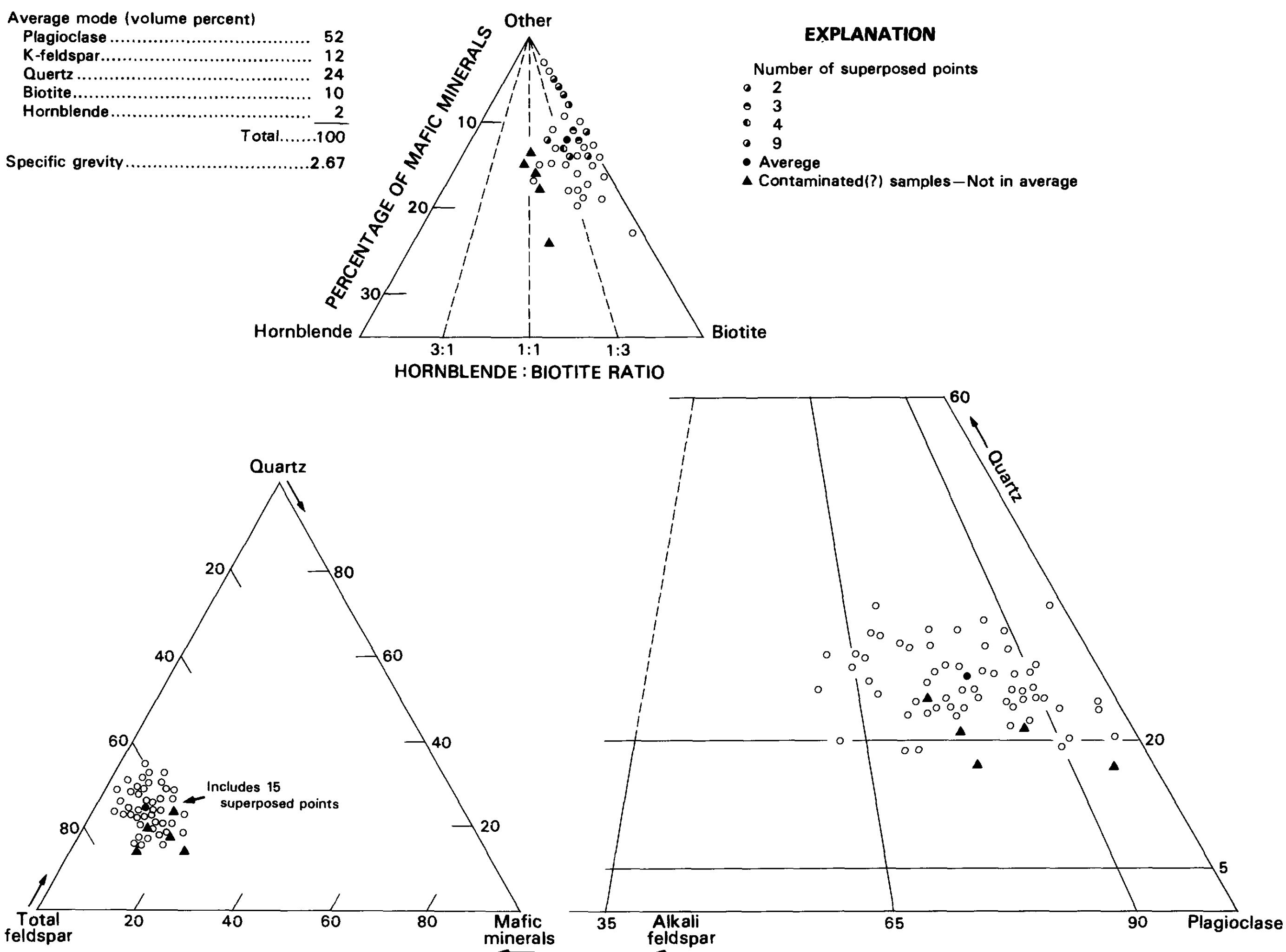


FIGURE 27.—Modal plots of the granodiorite of Gato-Montes.

TABLE 20.—*Modes of the granodiorite of Gato-Montes*

[All modes in volume percent; n.d., not determined. Others: A, allanite; C, clinopyroxene; M, muscovite; O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Others	Specific gravity
662	50	9	28	10	3	---	2.68
663	57	8	22	10	3	---	2.68
664	53	14	23	10	---	---	2.61
667A	44	21	24	10	1	---	n.d.
3310C	51	15	30	4	---	---	n.d.
3316	53	12	23	8	4	<1(C)	2.69
3317	53	15	20	8	4	---	2.67
3322	52	13	27	8	---	---	2.66
3340	44	13	28	8.5	6.5	---	2.70
3356	57	6	25	11	1	<1(C), <1(S)	2.68
3357	60	5	21	13	1	---	2.70
3371B	52	8	32	8	---	---	2.66
3372	53	6	29	11	1	---	2.68
3373	58	7	23	11	1	<1(C), <1(M)	2.69
3377	51	11	31	7	---	---	2.66
3389	43	19	31	7	---	---	2.64
3390	63	6	28	3	---	---	2.64
3481A	52	10	24	10	3	1(S)	2.71
3483A	62	2	20	16	<1	<1(M), <1(S)	2.72
3492	53	12	22	9	4	<1(S)	2.69
3498	51	17	21	8	3	<1(S)	2.68
3499	63	7	17	12	1	<1(S)	2.70
3502	56	9	21	10	3	1(S)	2.70
3505	60	8	24	8	<1	<1(M)	2.66
3508	40	24	28	8	---	---	2.64
3515	42	22	28	8	<1	<1(S)	2.66
3517	63	2	17	13	5	<1(S)	2.71
3521	43	15	26	13	3	<1(S)	2.69
3523	49	9	22	15	5	---	2.71
3534B	52	15	23	9	.5	.5(S)	2.69
3733A	56	2	18	22	1	<1(A), 1(S)	2.71
3734A	56	9	21	13	1	<1(O), <1(S)	2.66
3734B	55	9	22	12	2	---	2.67
3740	57	7	21	13	1	1(S)	2.625
3745	52	14	20.5	13	---	.5(S)	2.67
3746	41	22.5	25.5	10	1	---	2.65
3747A	55.5	10	19	9.5	5	<1(O), 1(S)	2.69
[1]3754	47	16	20.5	8.5	8	---	2.65
3756	53	19	23	5	---	<1(S)	2.61
3756-1	42	28	25	5	---	---	2.70
3757-4A	42	27	17	14	(?)	---	2.69
3757-6	50	20	16	11	3	---	2.69
3762B(?)	56	1	32	11	---	---	2.69
3763B(?)	56	8	18	14	4	---	2.72
3766(?)	62	6	17	14	1	---	2.69
3818	52	7	23	9	8	1(S)	2.71
3819	52	7	26	11	4	<1(S)	2.69
3820	46	17	29	8	---	<1(A)	2.63
3829A	42	18	34	6	<1	---	2.64
3829B	51	18	23	8	---	<1(S)	2.65
3830	54	16	22	8	---	---	2.65
[1]4004A	58	9	19	7	7	---	2.69
[1]4004C	60	3	13	10	---	<1(C)	2.76
4005A	45.5	14	23	14	3.5	---	2.71
4006	54.5	6.5	20	17	2	---	2.71
4013	43	19	29	8	---	1(S)	2.62
4014	49	15	25	11	---	---	2.69
4018	45	21	28	6	---	---	2.63
[1]4026	50	15	17	10	8	---	2.69
[1]4027	55	16	14	7	8	---	2.70
4028	44	21	23	10	2	---	2.66
4029	47	19	20	11	3	---	2.67
4031	46	19	16	15	4	---	2.71
4034	49	15	23	10	1	2(S)	2.68
Average--------	52	12	24	10	2	---	2.67
Standard deviation.	6.0	5.9	4.8	3.3	1.9	---	.03

[1]Sample contaminated(?); not used in calculation of averages.

TABLE 21.—*Modes of the granodiorite of Cameron*

[All modes in volume percent]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Hornblende	Specific gravity
3773	50	6	26	15	3	2.67
4048	51	15	21	10	3	2.64
4047	46	19	28	5	2	2.65
4048	49	14	23	9	5	2.66
4061	45	26	23	5	1	2.63
4062A	43	23	23	10	1	2.66
Average--------	47	17	24	9	3	2.65
Standard deviation.	3.5	7.1	2.5	3.7	1.5	.01

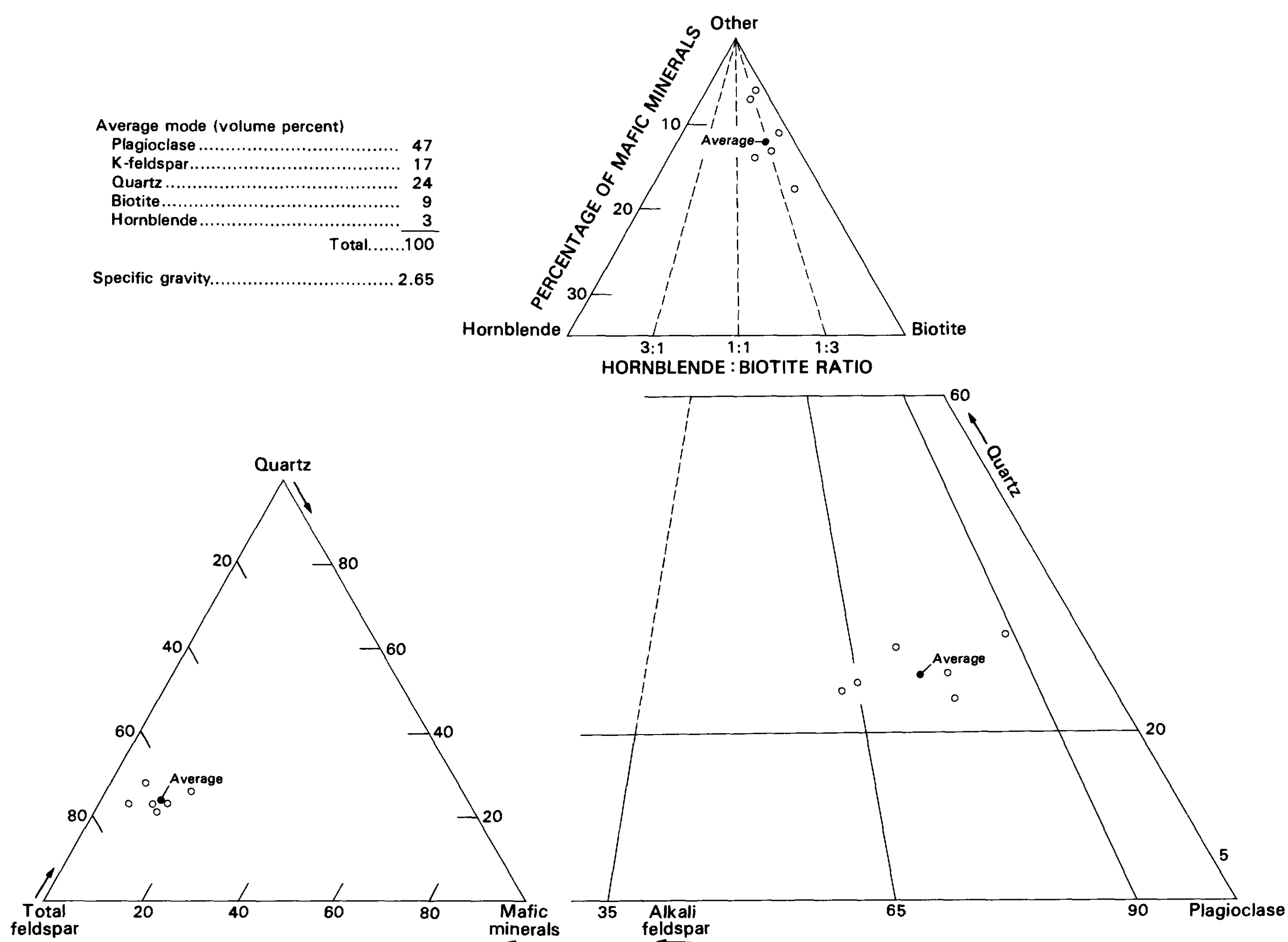


FIGURE 28.—Modal plots of the granodiorite of Cameron.

of the granite of Tejon Lookout, a feature indicating that the contact is not completely planar. Nevertheless, the overall parallelism of the contact of the granitic and metamorphic rocks with the topography is highly unusual and suggests a structural break. Thermal metamorphism along the surface between the granitic and metamorphic rocks, however, indicates an intrusive contact rather than a fault, although Wiese (1950) noted that the surface could represent a pregranite thrust fault along which granite intruded.

Petrography

The granite of Tejon Lookout (fig. 30; table 22) is a medium- to coarse-grained rock type that locally forms prominent, craggy, orange- to yellow-weathering outcrops. Through much of its outcrop area, however, it is poorly exposed or deeply weathered. The coarse-grained granite apparently disintegrates easily into yellowish grus, a characteristic that is distinctive for identifying the area of exposure but allows for very few fresh samples for study. The grain size varies; the coarsest quartz and K-feldspar crystals are as large as 1 cm across, but most grains in most specimens do not exceed 5 mm across. The three small granite stocks or cupolas in the eastern part of the large mass of metamorphic rocks in the Lebec quadrangle are notably finer grained and have a hypabyssal texture that features large crystals of quartz and feldspar set in an aplitic groundmass. Similar textures have been seen locally in the main granite mass. Wiese (1950) noted this texture "* * * in the marginal parts of the main granite intrusive, 10 to 200 feet from the overlying metamorphic rocks." He described the marginal rocks as "* * * porphyritic, consisting of phenocrysts of feldspar and quartz as large as a centimeter across in a microgranitic matrix of quartz, feldspar, and biotite."

In thin section, the dominant feature of the Tejon Lookout mass is the abundance of coarse grains of patchy to stringy perthite in which the K-feldspar is strongly grid twinned. All of the perthitic material was counted as K-feldspar in the slab modes, and so the modal average is overly weighted toward K-feldspar. Plagioclase is generally very finely twinned and in part weakly zoned; its composition is about An_{30}. Both the plagioclase and the K-feldspar crystals are partly subhedral. Quartz ranges in size from small globules to coarse grains, as much as 1 cm across, that are locally granulated. Small amounts of widely scattered brown biotite in small shreds and discrete crystals are present throughout the mass. Primary(?) muscovite is present in trace amounts. Crystals of zircon and apatite chiefly occur as inclusions in the biotite. Metallic opaque-mineral grains are absent in most specimens.

The granite of Tejon Lookout is one more example in the study area of a coarse-grained, low-melting-trough granite that contains salmon-pink K-feldspar and weathers to yellow and orange tints. The Tejon Mountain body is physically and compositionally similar to the granite of Brush Mountain, and the two bodies may be correlative.

GRANITE OF BEAN CANYON

Between Tylerhorse and Bean Canyons on the south side of the Garlock fault are five irregular felsic masses that cover no more than 8 km^2 in aggregate area. The rocks of these yellow-weathering bodies vary considerably in grain size and partly consist of aplite-alaskite-pegmatite. These felsic masses are composed of about equal amounts of sodic plagioclase, K-feldspar, and quartz; biotite is absent or present only in small amounts (fig. 31; table 23).

The positions of the plutons composed of the granite of Bean Canyon, which fill the gap between the main areas of outcrop of the granite of Tejon Lookout, are probably more than coincidental. Locally, both the texture and mineralogy of the granite of Bean Canyon are similar to those of the granite of Tejon Lookout—but only very locally. The limited data available indicate that the granite of Bean Canyon may well be a more felsic differentiate of the Tejon Lookout mass.

MAFIC IGNEOUS ROCKS IN (AND ASSOCIATED WITH) THE BEAN CANYON FORMATION

Mafic rocks, referred to collectively as "hornblende diorite and gabbro" by Dibblee (1967), are scattered

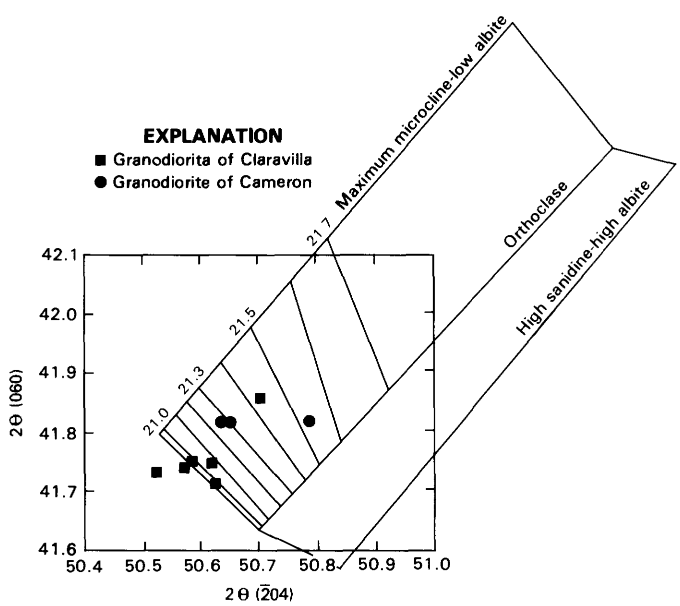


FIGURE 29.—Structural state of K-feldspar for selected samples of the granodiorites of Claraville and Cameron.

through the basement terrane south of the Garlock fault. Similarly named rocks north of the Garlock fault are in large part hornblende-rich metamorphic rocks. It thus seems worthwhile to briefly summarize the data from these mafic rocks south of the Garlock fault, as well as from the hornblende-bearing rocks of the Bean Canyon Formation.

A much-altered and retrograde ultramafic rock, rich in colorless acicular amphibole that is studded with olivine, is exposed in Bean Canyon (fig. 2). G.A. Davis (written commun., 1977) suggested that this serpentinized peridotite is a fragment of disrupted ophiolite. West of Cottonwood Creek (fig. 2), another possible ultramafic remnant is present that is altered to a mat of hornblende and epidote, with traces of plagioclase and biotite. East of Cottonwood Creek, a remnant of truly intrusive gabbro was found in a mixed area of largely contaminated granodiorite. The gabbro contains abundant pale-brown to colorless, in part euhedral hornblende crystals, as large as 3 mm across. Also present are smaller, in part euhedral clinopyroxene crystals, plagioclase (approx An_{50}), and minor K-feldspar. These are the only localities south of the Garlock fault where I have noted unequivocal mafic and ultramafic magmatic rocks. Some areas previously mapped as hornblende diorite and gabbro have proved to be largely dark (contaminated?) granodiorite, but I have not examined all these mafic rocks, and some may be gabbro.

Metavolcanic rocks are a minor component of the Bean Canyon Formation along its entire outcrop belt. Most are readily recognizable in thin section by the preservation of phenocrysts (or pyroclasts) of plagioclase, as long as 4 mm, and, less commonly, of quartz. Aggregates of brown biotite and green hornblende also mimic former larger crystals in some layers. All these metavolcanic rocks now have a dense hornfelsic matrix. Original compositions range from rhyolite or quartz latite to andesite or basalt. Some of these metavolcanic rocks are rich in hornblende and are now amphibolites, but all are fine grained, and none bear resemblance to the coarser grained, high-grade metamorphic rocks north of the Garlock fault.

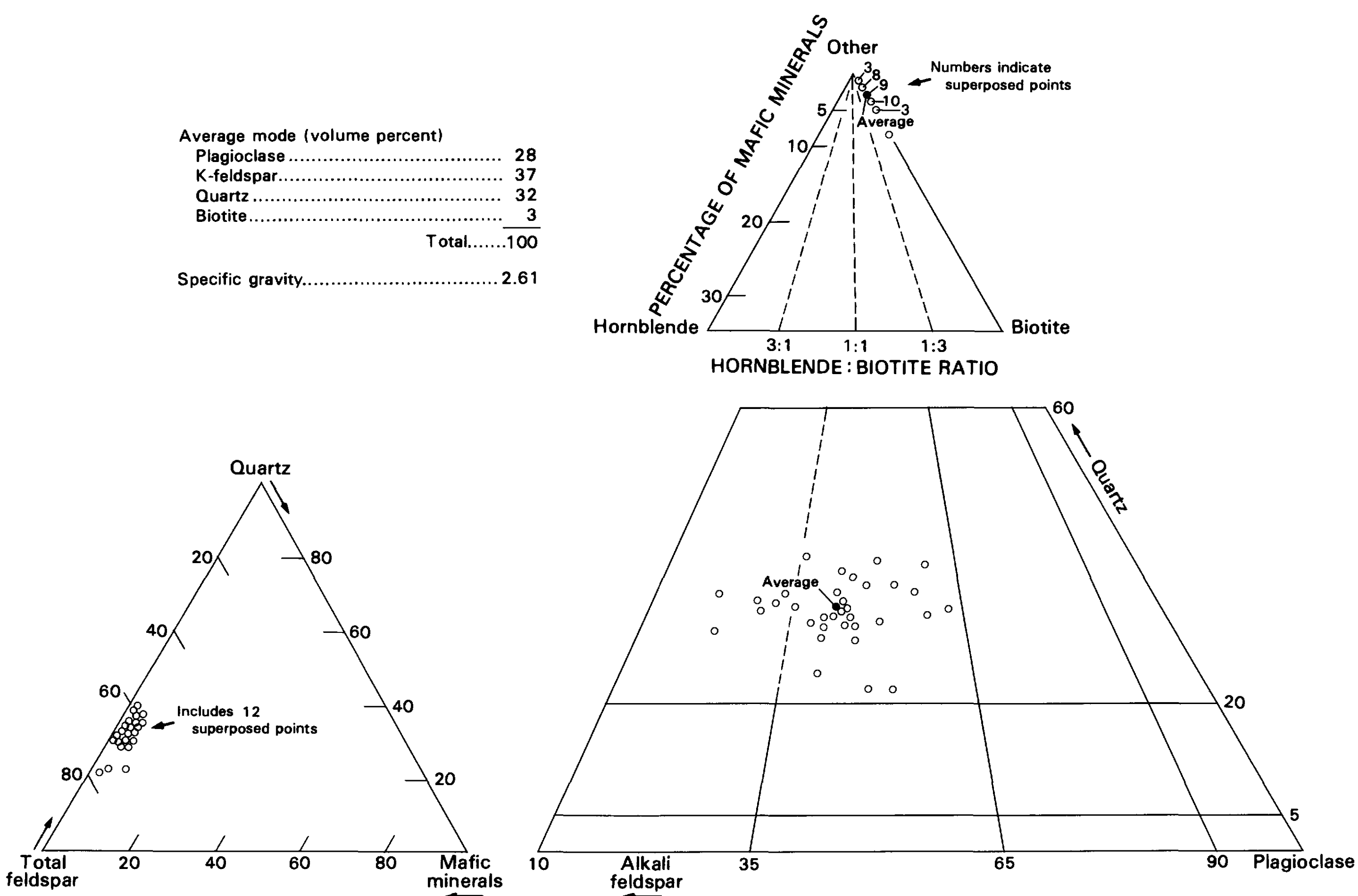


FIGURE 30.—Modal plots of the granite of Tejon Lookout.

TABLE 22.—*Modes of the granite of Tehon Lookout*

[All modes in volume percent. Others: O, opaque minerals; S, sphene]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Others	Specific gravity
647	27	39	30	4	---	2.59
666	38	27	31	4	---	2.64
667B	28	36	32	4	---	2.67
3307A	24	42	32	2	---	2.61
3308	31	35	30	4	---	2.59
3309	22	42	34	2	---	2.59
3313A	32	36	28	4	---	2.60
3315	28	40	28	4	---	2.60
3323	21	44	33	2	---	2.59
3329	19	46	32	3	---	2.60
3330	16	52	29	3	---	2.60
3339	22	37	39	2	---	2.60
3401	40	24	32	4	---	2.60
3457A	31	37	31	1	---	2.57
3458	30	35	30	5	---	2.61
3465	18	46	33	3	---	2.62
3466	31	32	35	2	---	2.62
3467	39	35	22	4	---	2.62
3469A	33	29	35	3	---	2.60
3473	26	41	30	3	---	2.60
3475	34	25	37	5	---	2.60
3476A	36	27	34	3	---	2.62
3480	34	33	30	3	---	2.58
3493	26	34	37	3	---	2.61
3495	28	33	35	4	---	2.62
3509	28	36	33	3	---	2.61
3512	28	37	31	4	---	2.60
3514	28	36	34	2	---	2.58
3741	29	41	22	8	<1(O), <1(S)	2.63
3752A	14	50.5	34	1	.5(O)	2.60
3755	37	39	21	2	1(O)	2.60
3763A	30	30	38	2	---	2.60
3769	27	38	30	5	---	2.61
3771	30	36	33	1	---	2.59
Average--------	28	37	32	3	---	2.61
Standard deviation.	6.3	6.6	4.1	1.4	---	.02

MISCELLANEOUS TOPICS (METAMORPHIC AND PLUTONIC ROCKS)

DISTRIBUTION OF SELECTED MINERALS AND THEIR SIGNIFICANCE

The following discussion focuses on the distribution of selected minerals in the southernmost Sierra Nevada that may have specific value as indicators of metamorphic grade, crustal depth, or other petrologic significance. Identified localities, based on thin-section identification, are shown on index maps for each mineral. These index maps do not provide an exhaustive catalog of the occurrences of these minerals in the study area, but they should provide a fairly representative sample because some 800 thin sections of metamorphic and plutonic rocks were examined in this study.

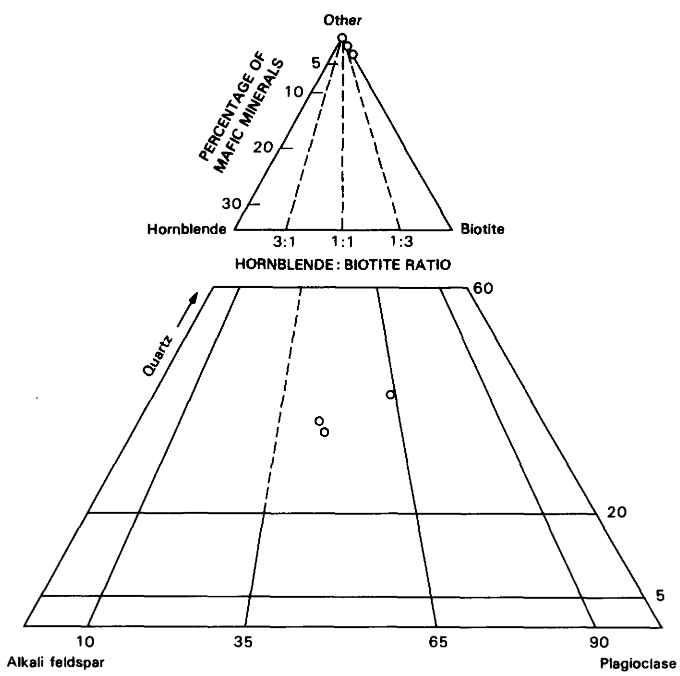
HYPERSTHENE[3]

Hypersthene in distinctly pleochroic (pink to green) fresh grains occurs in three settings in the southernmost Sierra Nevada (fig. 32). First, some hypersthene is in undoubtedly granoblastic, high-grade metamorphic rocks. In part, these rocks are somewhat retrograded, and the hypersthene has a pale amphibole reaction rim against plagioclase and dark-colored primary hornblende. In other samples, fresh hypersthene has clean, sharp contacts with both hornblende and plagioclase, suggestive of preservation of a high-grade equilibrium assemblage. Second,

[3]Detailed investigations by Sams (1986) have suggested that most of the hypersthene in the study area is igneous rather than metamorphic in origin. However, the highly enriched $\delta^{18}O$ values (max 15.0 permil SMOW) in some hypersthene-bearing samples suggest that some of the rocks referred to in this report are metamorphic granulite with a sedimentary source (Ross, 1983c).

hypersthene also occurs as remnants in undoubtedly retrograde gabbro and ultramafic magmatic rocks. Third, hypersthene occurs in rocks that look magmatic. Modal data on some of these rocks are shown in figure 33 and listed in table 24. Much of the hypersthene in these rocks is fresh and sharply contacts adjacent minerals or has only a thin reaction rim of pale amphibole. In the Cummings belt, the hypersthene is in tonalitic rocks that are closely associated with granulite. The hypersthene occurrences near the north edge of the study area (in the Pampa body and near the Live Oak body) resemble the hypersthene in the tonalite in the Cummings belt, but the setting suggests that the hypersthene here either may be a primary accessory mineral or may have been inherited from nearby mafic and ultramafic plutonic rocks. Hurlbut (1933) attributed the presence of hypersthene and augite in mafic inclusions in the southern California batholith to contamination from gabbro.

I have been repeatedly perplexed by tonalitic rocks that have magmatic features (sharply twinned and zoned plagioclase and hypautomorphic texture) but contain fresh hypersthene in equilibrium with adjacent hornblende and



Average mode (volume percent)

Plagioclase	34
K-feldspar	28
Quartz	37
Biotite	1
Total	100
Specific gravity	2.60

FIGURE 31.—Modal plots of the granite of Bean Canyon.

TABLE 23.—*Modes of the granite of Bean Canyon*

[All modes in volume percent. M, muscovite]

Sample	Plagioclase	K-feldspar	Quartz	Biotite	Specific gravity
3828	37	21	42	1(M)	2.58
4016	30	35	34	1	2.61
4032	34	27	36	3	2.60
Average---	34	28	37	1	2.60

plagioclase crystals. These rocks have seemed to me to be transitional between magmatic tonalite and metamorphic granulite. Warren Hamilton (written commun., 1983) suggested that the hypersthene tonalites are magmatic rocks which crystallized under granulite-facies conditions and that the hypersthene tonalite reflects crystallization under conditions of lower water content and higher pressure and temperature relative to the shallower, normal granitic rocks of the Sierra Nevada batholith. However, Saleeby (1977) and Saleeby and Sharp (1980) discussed hypersthene-bearing tonalites from the west side of the central Sierra Nevada batholith (Durrell, 1940; Macdonald, 1941) as part of a suture-filling suite of rocks, ranging from cumulate olivine gabbro to biotite-hornblende tonalite. To my knowledge, no coarse amphibolitic gneiss or hypersthene-bearing granulitic rocks have been reported from the area that Saleeby (1977) discussed. I note that the tonalite of Bear Valley Springs, which is in a transitional position between oceanic and continental crust, on the basis of initial strontium-isotopic data, could well be a tonalitic end member of Saleeby's (1977) suture-filling suite. Two things set the southernmost Sierra Nevada rocks apart from the more northern areas: (1) the close association of hypersthene-bearing magmatic tonalite and unequivocal hypersthene granulite (for example, in the Cummings belt); and (2) inclusions of gneiss, granulite, and associated rocks that are found throughout the outcrop area of the tonalite of Bear Valley Springs.

In May 1983, I visited one of the bodies of hypersthene-bearing rock in the central Sierra Nevada—the Academy pluton, first described by Macdonald (1941) and later examined and studied in detail by Mack and others (1979). My cursory observations (and reading the report of Mack and others, 1979) left no doubt that the hypersthene-bearing rocks of the Academy pluton are truly magmatic rocks. The presence of discrete mafic minerals, the rarity of ovoid mafic inclusions, and the absence of gneissic patches in the Academy outcrops that I examined also suggest a cleaner pluton of a higher crustal level than the tonalite of Bear Valley Springs appears to represent.

Hypersthene in tonalite can originate in different geologic environments, and in the southernmost Sierra

Nevada the choice is not everywhere obvious. Possibly, the two origins of the magmatic hypersthene just mentioned can be somewhat melded together. A suture-filling transitional tonalite could bring up and include fragments of deeper, batholithic-root rocks. Thus, the local hypersthene-bearing spots in the tonalite of Bear Valley Springs reflect its deeper origin, whereas most of this possibly suture-filling rock now reflects a somewhat higher crustal level.

Hypersthene is spread over what seems to be a north-south-trending belt encompassing the eastern part of the San Emigdio-Tehachapi terrane and several of the mafic patches engulfed in the tonalite of Bear Valley Springs (fig. 32). I have identified hypersthene in only 25 of the several hundred thin sections that have been examined of presumably suitable host rocks (mafic rocks, metamorphic rocks, and tonalite). Hypersthene is easily identifiable in thin section, and golden-brown hypersthene crystals can generally be identified in hand specimens after a little practice. Thus, hypersthene is not apt to be overlooked, and its relative sparseness in the southernmost Sierra Nevada is probably real. Some caution should be exercised, however, in extrapolating from these sparse hypersthene occurrences to conclusions about metamorphic grade for the entire area.

COARSE-HALOED RED GARNET

Distinctive anhedral to euhedral garnets, as large as 10 cm across, are widespread in the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains (pl. 1). These crystals commonly are accentuated in dark matrix rocks by a bleached halo, from a few millimeters to 1 cm thick, that is virtually devoid of dark minerals.

These distinctive garnets were first described from boulders in alluvial fans and in a streambed about 18 km southwest of Cummings Valley (Murdoch, 1939; the locality that Murdoch visited is the one with 6-cm garnets

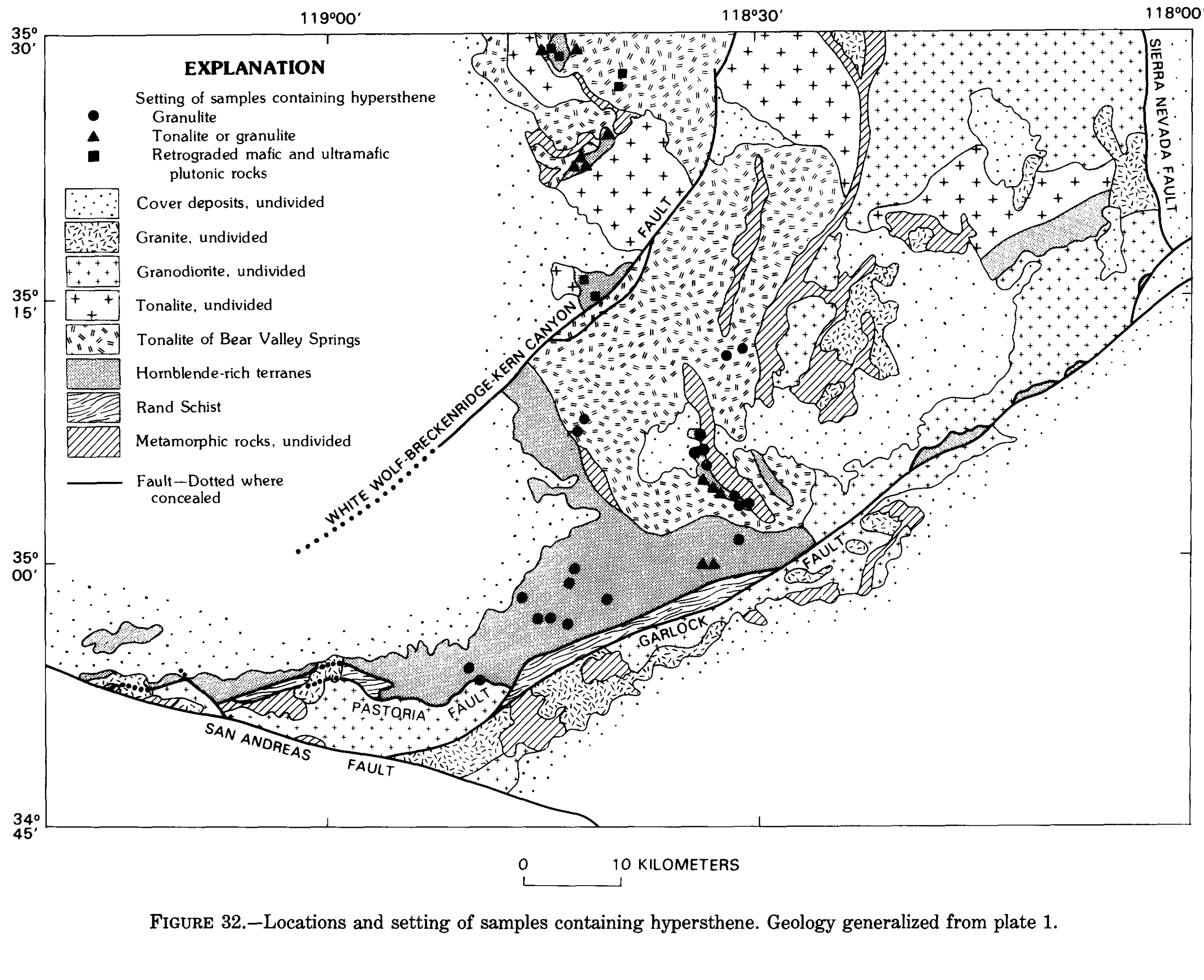


FIGURE 32.—Locations and setting of samples containing hypersthene. Geology generalized from plate 1.

noted on pl. 1). Wiese (1950) noted several garnet-bearing outcrops in the Neenach 15-minute quadrangle but did not specifically locate them. A chemical and petrographic study of the garnets and their matrix material was made by Schürmann (1938).

My field studies have shown that coarse-haloed garnets are widespread in the mafic metamorphic rocks, and I am sure that there are more localities than my cursory studies discovered. Coarse-haloed garnets are also present in the tonalite of Bear Valley Springs and in hybrid metamorphic rocks associated with the tonalite, and there are also two occurrences in gneissic, hybrid-appearing rocks near the contact of the metasedimentary rocks of the Keene area with the tonalite of Hoffman Canyon.

Most of the garnets occur as isolated, white-rimmed eyes in a tonalitic, dioritic, or amphibolitic matrix, but some garnets are clustered within a common white halo. The garnets also occur as strings of beads in white veins; generally, these crystals are no larger than 1 cm across. Almost all the garnets are markedly poikilitic and enclose crystals of plagioclase and opaque-mineral grains. Some lacy patches of garnet suggest that the process of development was arrested before more euhedral crystals formed. Nearly all the garnets are anhedral and irregular, but the less common euhedral crystals are especially noticeable. I collected one euhedral dodecahedral crystal, 9 mm across, whose edges are modified by trapezohedron faces. Most of the larger euhedral garnets I have seen have this form, but some of the smaller (2–3 cm) euhedral crystals are solely trapezohedrons. Although some garnets are relatively fresh and a rich dark red, most are extensively fractured, weathered, and otherwise altered to rusty-red crystals that are far from gem quality.

Schürmann (1938) reported chemical analyses of garnets from boulders in the alluvial deposits where garnets were first noted by Murdoch (1939). Schürmann also analyzed "coarse-grained diorite," "finer-grained gabbro" (matrix rocks of the garnets), and hornblende. Quartz diorite and tonalite are host rocks for the garnets at the alluvial locality, but Schürmann did not analyze these host rocks. It is difficult to know what matrix rocks Schürmann analyzed from the data of table 25, but the modes of the

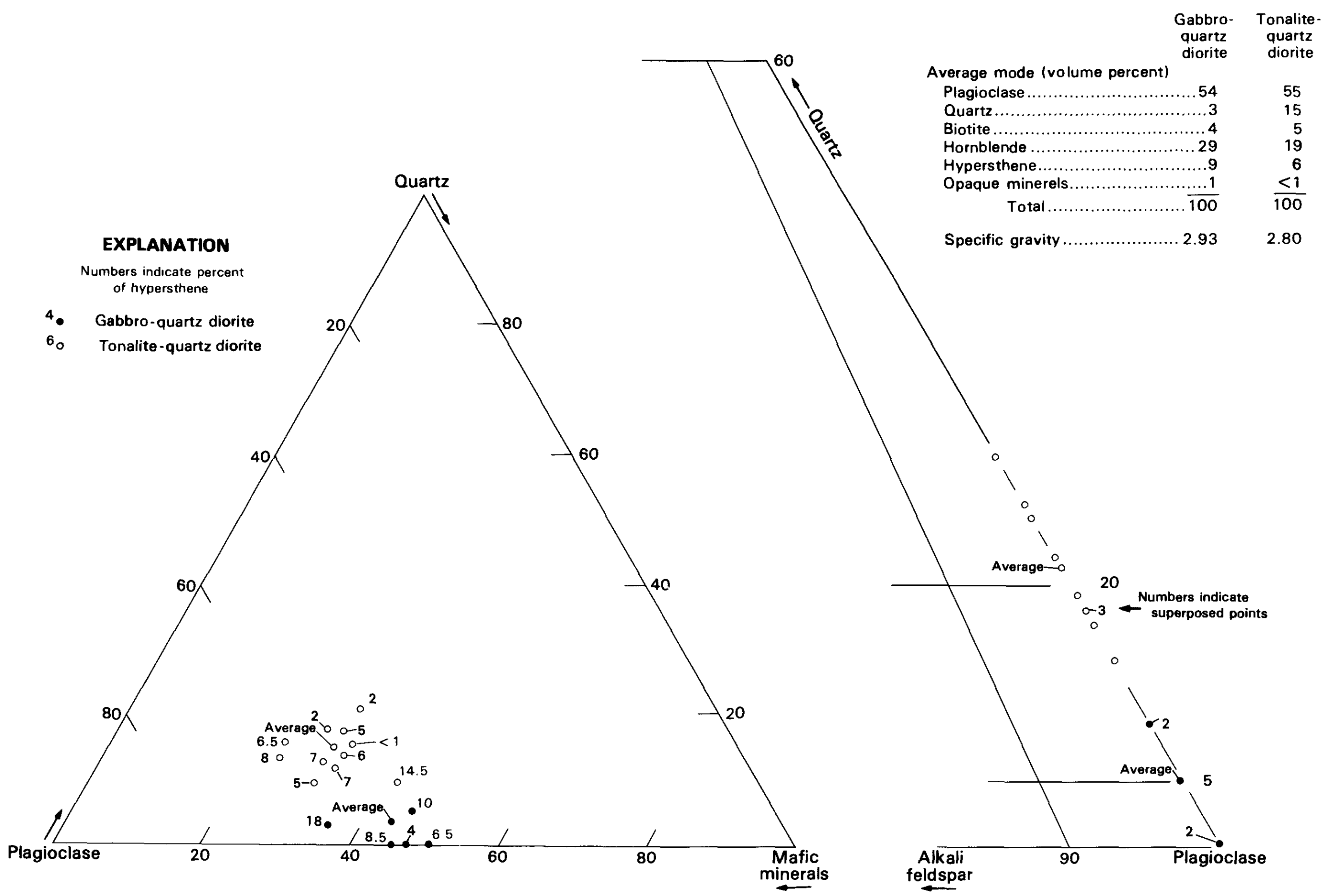


FIGURE 33.—Modal plots of hypersthene-bearing rocks.

TABLE 24.—*Modes of the hypersthene-bearing plutonic rocks*

[All modes in volume percent. n.d., not determined]

Sample	Plagioclase	Quartz	Biotite	Hornblende	Hypersthene	Opaque minerals	Specific gravity
			Tonalite to quartz diorite				
3352C	57	13	---	23	7	---	2.83
3353	52	18	---	25	5	---	2.82
3440	56	12	7	18	7	---	2.82
3441	54	18	8	18	2	---	n.d.
3593	61	16	<1	16.5	6.5	---	2.78
3594	60	10	5	19	5	1	2.82
3595	48	21	11	18	2	<1	2.81
3596A	52	16	11	21	<1	---	2.80
3702	54	14	---	26	6	---	2.80
3999C	49	10	10	16.5	14.5	---	2.79
4208A	62	14	4	12	8	---	2.70
Average--------	55	15	5	19	6	<1	2.80
Standard deviation.	4.7	3.5	4.6	4.1	3.5	---	.04
			Gabbro to quartz diorite				
3560	50	---	1	41.5	6.5	1	2.99
3778A	55	---	---	35.5	8.5	1	n.d.
3796	53	5	11	27	4	---	n.d.
4428B	50	5	4	30	10	1	2.91
4429	62	3	3	13	18	1	2.89
Average--------	54	3	4	29	9	1	2.93
Standard deviation.	5.0	2.5	4.3	10.7	5.3	.5	---

matrix rocks are compatible with amphibolite from the mafic metamorphic terrane. Chemical analyses of the host rocks indicate that they are unusually high in Al_2O_3 and low in K_2O in comparison with average diorite and gabbro (Nockolds, 1954). Some ophiolitic gabbros are high in Al_2O_3 and notably low in K_2O (Ross, 1972, p. 64; Bailey and Blake, 1974, p. 642). This composition would agree with the proposed oceanic affinity of the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains.

In an effort to get some modern chemical data on these garnets, microprobe studies were made on seven samples by L.C. Calk. Three samples (3302, 3938C, and 3949, fig. 34) are from the main body of the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains, one sample (3931) is from the alluvial-boulder locality first noted by Murdoch (1939), two samples (4044, 4060) are from slivers of dark amphibolitic gneiss along the Garlock fault zone, and one sample (3969) is from a granitic clast in the Bena Gravel of Miocene age. The granitic clast was most probably derived from the tonalite of Bear Valley Springs. Sample 3938C is probably from the source area of the alluvial-boulder locality (3931). Each chemical analysis in table 26 is an average of four to seven microprobe analyses of a single garnet. To test possible zoning in the garnets, the amount of each oxide for each individual determination is plotted in figure 35. Modest zoning is suggested by some of the oxide values of sample 3302.

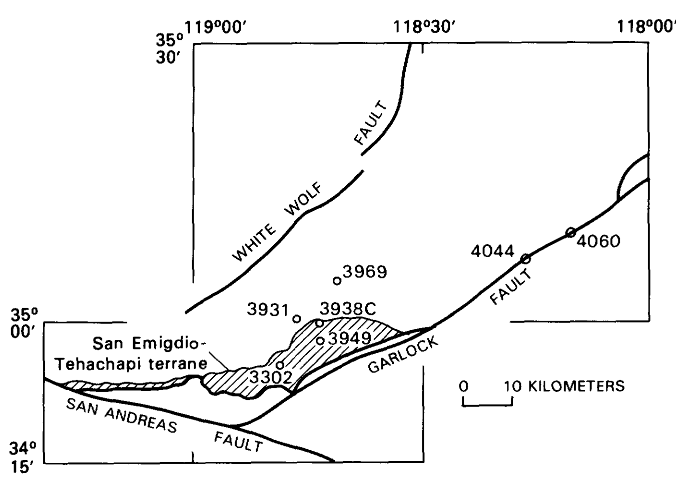


FIGURE 34.—Locations and sample numbers of garnets analyzed by electron microprobe.

TABLE 25.—*Chemical and peatrographic data on coarse-haloed garnets and their host rocks near sample locality 3931*

[All values in weight percent. Abstracted from Schürmann (1938)]

	Pure garnet[1]	Garnet crystal with inclusions of feldspar and ore[2]	Hornblende[3]	Plagioclase[4]	Diorite[5]	Gabbro[6]
SiO_2---------	36.70	37.04	41.55	---	46.16	43.60
Al_2O_3--------	20.83	20.72	16.21	---	24.15	18.70
Fe_2O_3--------	1.18	1.41	3.94	---	4.36	3.41
FeO----------	26.96	25.28	18.45	---	6.11	9.60
MnO----------	.98	1.12	.23	---	.09	.31
MgO----------	4.84	4.41	8.01	---	4.53	5.09
CaO----------	7.12	8.12	9.22	---	9.43	14.47
Na_2O---------	---	.12	.13	---	2.26	2.12
K_2O----------	---	.06	---	---	.18	.16
H_2O^+---------	.13	.31	1.00	---	.51	.60
H_2O^----------	---	---	---	---	.04	.06
CO_2----------	---	---	---	---	.04	.02
TiO_2---------	1.45	1.35	1.42	---	1.60	1.86
P_2O_5---------	.20	.40	---	---	.29	.25
Total---	100.39	100.34	100.16	---	99.75	100.25

[1]End-member molecules (in volume percent) almandine, 59; pyrope, 19, grossularite, 17; andradite, 3; spessartite, 2. Index of refraction, 1.79.

[2]Thin-section mode (in volume percent): garnet, 95; feldspar, 3.5; apatite and ore, 1.5. Specific gravity: pycnometer (powder), 3.98; chip, 4.085; X-ray (Röntgen) analysis, 4.074.

[3]Specific gravity, 3.23; index of refraction, 1.658-1.66. Distinctly pleochroic: α, bright yellow; β, olive green; γ, dark blue green. Maximum extinction angle, 21° (generally 17°-20°).

[4]An_{70-75}. $\underline{n}_g$, 1.571; $\underline{n}_p$, 1.5597. $2\underline{V}$, 85° to nearly 90° (+). Specific gravity: suspended in bromoform, 2.72; pycnometer, 2.692 (cor).

[5]Thin-section mode (in volume percent): plagioclase, 56; hornblende, 40; ore, 3; apatite, 1. Specific gravity, 2.95.

[6]Thin-section mode (in volume percent): plagioclase, 51; hornblende, 44; ore, 4; apatite, 1. Specific gravity, 3.04 (estimated from mode).

Sample 4060 shows erratic oxide values; the other garnets are relatively homogeneous.

Comparison of the analysis of sample 3931 with an analysis from the same general area reported by Schürmann (1938) and reproduced in table 25 shows that, in general, the two analyses compare favorably; however, there are differences, particularly in the amounts of FeO, CaO, and TiO_2. When end-member molecules are plotted on an almandine-pyrope-spessartite triangular diagram (fig. 36), Schürmann's analysis and three of the four new analyses cluster closely. One of the new samples analyzed (4044) is significantly lower in FeO and higher in MgO than the other three; this difference is reflected in the end-member plot, which shows it to be relatively richer in pyrope and lower in almandine than the other samples. Although this sample is from a sliver in the Garlock fault zone considerably east of the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains, the analyses are strikingly similar. Another of the new garnet analyses (sample 4060) is also from a sliver in the Garlock fault zone, but east of sample 4044. The analysis of sample 4060 is strikingly similar to the old analysis by Schürmann (1938)—both are relatively rich in CaO and, thus, in the grossularite end-member molecule. The garnet from the clast in the Bena Gravel (sample 3969) is notably higher in the spessartite end member and lower in pyrope and grossularite than are the samples from the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. This composition reflects the much higher Mn content and the lower MgO and CaO contents in sample 3969 relative to other samples, and suggests a possible difference between the metamorphic garnets of the gneiss, amphi-

TABLE 26.—*Chemical and end-member-molecular data on coarse-haloed garnets*

[Microprobe chemical analyses in weight percent; analyst, L.C. Calk. End-member-molecules in volume percent, based on the method of Rickwood (1968); samples also contain minor uvarovite and schorlomite]

Sample	3302	3931	3938C	3949	3969		4044	4060
					[1]1-5	[1]6-9		
Chemical analyses								
SiO_2	37.65	37.82	37.81	37.56	36.70	36.74	38.45	38.29
Al_2O_3	21.03	21.04	20.98	21.01	20.78	21.06	21.43	21.41
[2]Fe_2O_3	.47	1.00	.66	.58	.54	.11	.73	.56
FeO	28.46	29.02	28.69	29.12	26.60	28.86	21.88	26.11
MgO	4.55	5.74	5.12	5.28	2.61	3.40	8.42	5.47
CaO	5.07	3.94	4.01	4.03	2.33	2.14	6.46	7.08
TiO_2	.20	.13	.13	.13	.07	.06	.12	.11
MnO	2.02	1.26	2.27	1.34	10.00	6.69	.98	1.00
Cr_2O_3	.05	.05	.05	.04	.05	.06	.06	.03
Total	99.50	100.00	99.72	99.09	99.68	99.12	98.53	100.06
End-member molecules								
Andradite	1.02	2.71	1.70	1.48	1.52	0.27	1.91	1.43
Pyrope	18.00	22.55	20.19	20.94	10.60	13.79	32.62	21.21
Spessartite	4.55	2.80	5.07	3.03	23.06	15.42	2.16	2.17
Grossularite	12.85	8.02	9.25	9.63	4.94	5.70	15.67	18.08
Almandine	63.00	63.51	63.37	64.54	59.57	64.57	47.25	56.81

[1]Averages of multiple analyses on individual garnet crystals (see fig. 36).

[2]Fe_2O_3 content estimated by successive approximations of end-member molecules.

bolite, and granulite of San Emigdio-Tehachapi Mountains and the igneous accessory garnets of the tonalite of Bear Valley Springs.

BROWN HORNBLENDE

Hornblende that is pleochroic in various shades of light brown to olive brown is widespread in the amphibolite and mafic gneiss of the San Emigdio-Tehachapi Mountains and in the closely related Cameron and Cinco areas (fig. 37). Brown hornblende is also common in and near the Caliente, Pampa, and Live Oak mafic bodies north of the White Wolf fault. In part, these occurrences are from retrograde mafic and ultramafic plutonic rocks, but brown hornblende is also found in definitely metamorphic amphibolite in these same areas. In addition, there are scattered occurrences of olive brown hornblende in normal granitic rocks.

In contrast, the typical hornblende color in thin section of granitic rocks of the southernmost Sierra Nevada is bright green, grassy green, or olive green. Such green hornblende also occurs in the San Emigdio-Tehachapi mafic terrane, particularly in a broad transition zone with the tonalite of Bear Valley Springs.

Brown hornblende in metamorphic rocks is generally an index to relatively high grade conditions, as suggested by the following workers. Howie (1955) noted that hornblende in the granulite facies of the charnockitic rocks of Madras, India, has a characteristic olive-brown color. Greenish-brown or brown hornblende is associated with only the highest grade zone in a region of progressive regional metamorphism in the central Abukuma Plateau, Japan (Shido and Miyashiro, 1959). Eskola (1952) observed that in Lapland, common green hornblende is a reliable criterion of the amphibolite facies, in contrast to the typical greenish-brown hornblende of the lower granulite facies. Hamilton (1981) cited brown hornblende as a useful "index mineral" for the lower granulite facies. These data suggest that the wider distribution of brown hornblende (fig. 37) relative to hypersthene (fig. 32), par-

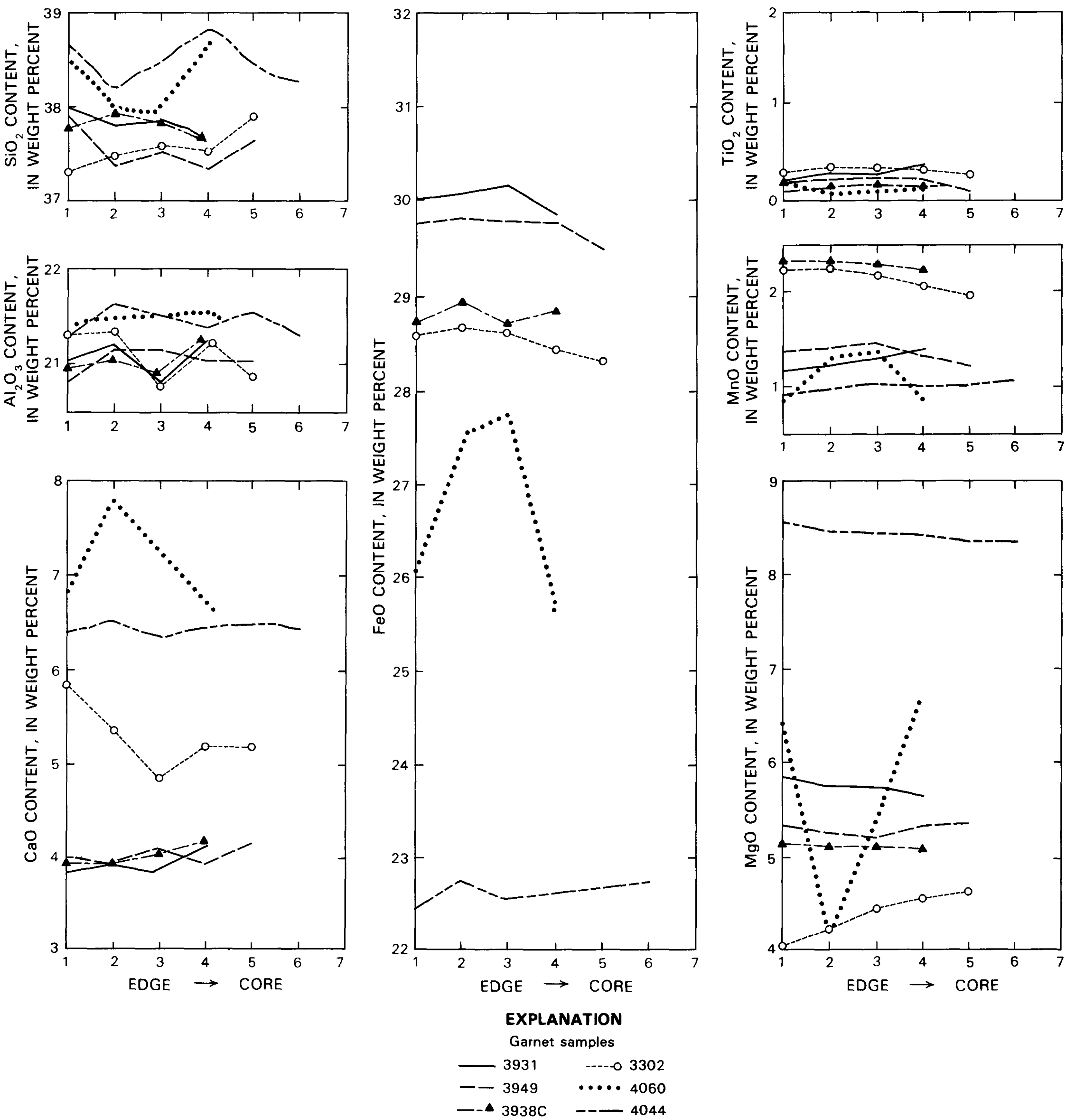


FIGURE 35.—Compositional data for selected garnets to show possible zoning. Numbers at base of graphs indicate relative locations of electron-microprobe analyses from edge to core of crystals. Analyst, L.C. Calk.

ticularly westward into the Sierra Nevada tail, points to a more extensive area of granulite-facies metamorphism in the San Emigdio-Tehachapi mafic terrane than the distribution of hypersthene-bearing samples alone would indicate.

REDDISH-BROWN BIOTITE

Reddish-brown biotite is widespread in the granodiorite of Lebec south of the Pastoria fault (fig. 38) and less common, but widespread, in the presumably correlative granodiorite of Gato-Montes to the east. Elsewhere in the granitic terrane, only scattered specimens contain distinctive reddish-brown biotite. Throughout the granitic terrane (except for the Lebec body), dark-brown to opaque interference colors are characteristic of biotite. Most of the biotite in granofels and felsic gneiss of the San Emigdio-Tehachapi terrane is distinctly reddish brown. Thus, reddish-brown biotite apparently is common in both granitic and biotite-bearing metamorphic rocks in the Sierra Nevada tail, but rare elsewhere.

Reddish-brown biotite generally indicates a relatively high TiO_2 content (Deer and others, 1965). Engel and Engel (1960) noted a systematic color change from greenish brown through reddish brown to deep reddish black, and a corresponding increase in TiO_2 content, with increasing regional metamorphism in the Adirondack Mountains, N.Y. John (1981) recorded the presence of reddish-brown biotite with "higher Ti contents in higher grade assemblages" in pelitic metamorphic rocks at the north end of the Gabilan Range, Calif. He furthermore noted that all the prograde mineral assemblages are above the second sillimanite isograd of Evans and Guidotti (1966). Warren Hamilton (written commun., 1982) suggested that red biotite in a metamorphic rock indicates highest amphibolite and granulite facies (at or higher than the "second sillimanite isograd"), and that red biotite contains half as much combined water as ordinary green and brown biotite, a feature restricting red biotite to high-grade rocks.

Reddish-brown biotite in metamorphic rocks thus appears to be a good index mineral for fairly high grade conditions. Reddish-brown biotite in granitic rocks probably means at least a relatively high TiO_2 content and, where abundant, as in the granodiorite of Lebec (fig. 38), may reflect some unusual (deep?) environment.

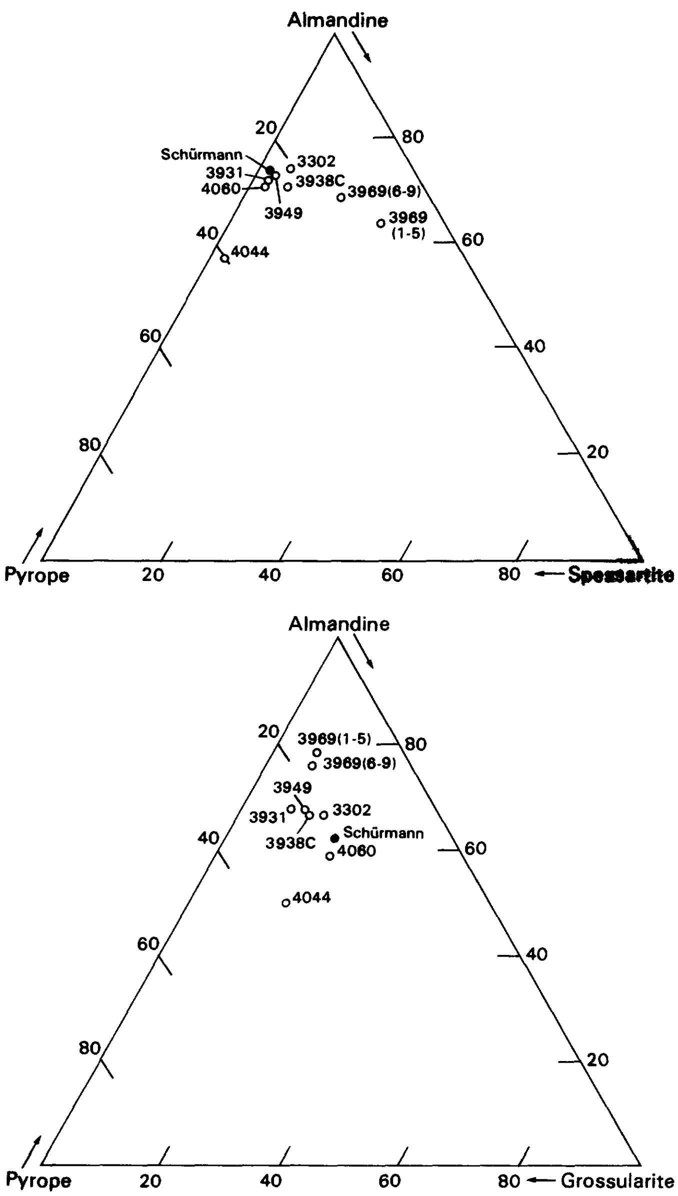


FIGURE 36.—End-member molecules for selected garnet samples (normalized to 100 mol percent). Numbers in parentheses after sample 3969 indicate multiple electron-microprobe analyses on two grains.

MUSCOVITE

Occurrences of discrete, relatively coarse muscovite crystals in granitic rocks are plotted on figure 39. Occurrences of definitely secondary white mica are not plotted. Nevertheless, one person's clean, discrete, primary muscovite crystal is, to someone else, a coarsely reconstituted, secondary muscovite crystal. Most of the units in which I have found coarse-grained, discrete muscovite also contain at least some hornblende and sphene, though not necessarily in the same sample. Also, at most 1 percent of coarse-grained muscovite is present in any one sample.

Coarse, discrete crystals of muscovite are relatively widespread in the granodiorite of Lebec but are limited to one small area in the presumably correlative granodiorite of Gato-Montes. North of the Pastoria and Garlock faults, coarse-grained muscovite is sparse, but it is relatively concentrated in the northeastern part of the study area.

Normative corundum is commonly cited as a clue to peraluminous, muscovite-bearing rocks. As much as 2 to 2.5 percent of normative corundum is present in some chemical analyses of granitic rocks of the southernmost Sierra Nevada. Uncertainties in analytical values of Na_2O, K_2O, CaO, and Al_2O_3 suggest that such amounts of normative corundum are unreliable guides to primary muscovite. Whether the mica is primary or secondary, I have been repeatedly impressed by the coarse discrete flakes of muscovite and the clean interlayers of muscovite and biotite in these rocks. These minerals are not simple alteration products. If they are, indeed, secondary, they at least represent considerable reconstitution of material that to me suggests something hotter, deeper, and (or) higher grade than typical deuteric or hydrothermal late granitic processes that produce the normal sericitic alteration.

I suggest that the relative abundance of coarse-grained muscovite in the granodiorite of Lebec, coupled with the abundant reddish-brown biotite in the same unit, may be significant. The granitic rocks in the Sierra tail south of the Pastoria fault zone may reflect a somewhat deeper crustal level than the similar-looking granodiorite and granite bodies to the north and east.

Miller and Bradfish (1980) referred to experimental work by Luth and others (1964) and Day (1973) which suggested that "ideal muscovite" should not be stable in granitic rocks at pressures of less than about 0.3 GPa.[4] Primary muscovite would thus seem to suggest crustal depths of at least 10 km. Miller and Bradfish (1980) pointed out, however, that plutonic muscovite is "nonideal" in composition, may have a larger stability field, and may be stable at lower pressures and shallower crustal depths. Thus, even if muscovite in granitic rocks is primary, there is still some question of what it means

[4] 1 gigapascal (GPa) = 10 kilobars.

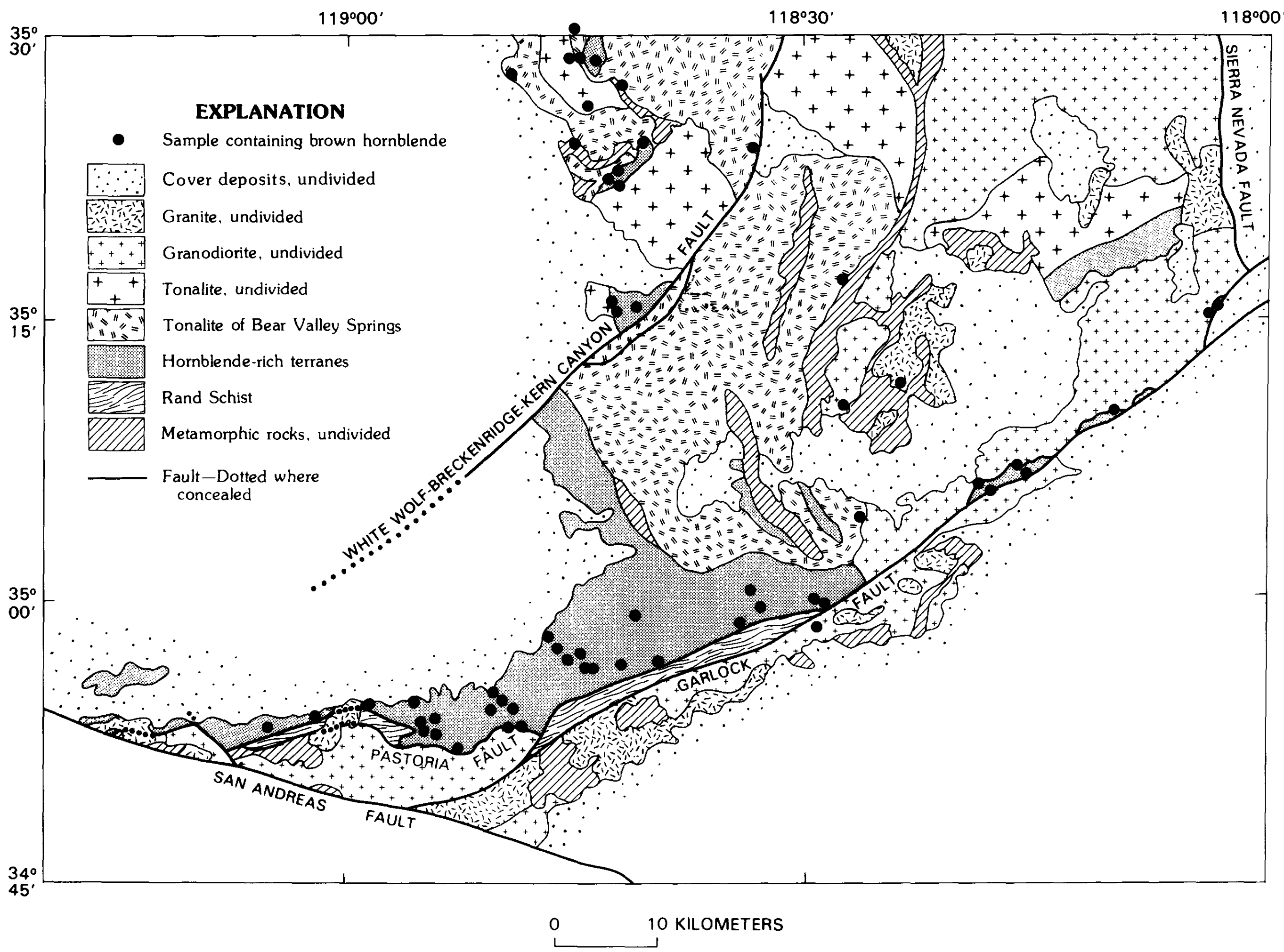


FIGURE 37.—Locations and setting of samples containing brown hornblende. Geology generalized from plate 1.

in terms of crustal depth. Nevertheless, one observation remains—primary-looking muscovite appears to be more common in the southernmost Sierra Nevada than in the rest of the batholith.

SILLIMANITE AND ANDALUSITE

Sillimanite both as coarse prismatic crystals and in fibrolite bundles is common and widespread (fig. 40) in the metasedimentary rocks included in the granodiorite of Lebec (south of the Pastoria fault). Sillimanite is also widespread north of the Garlock fault in several metasedimentary pendants. I note that sillimanite is particularly abundant immediately north of the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains.

Andalusite is much less widespread and is abundant only in the Pampa Schist (north-central part of the study area), where it occurs in coarse conspicuous crystals that are partly chiastolitic. Coarse-grained andalusite also is locally abundant in dark metasedimentary layers in the Bean Canyon Formation, south of the Garlock fault. In part, the more limited occurrence of andalusite may be a compositional control. Black, dense hornfelsic rocks that seem to be the perfect host for andalusite (and, particularly, chiastolite) are generally limited to the Pampa and Bean Canyon units.

The occurrences of sillimanite and andalusite are mutually exclusive except near the northern part of the study area, where the two aluminosilicates are closely associated in the Pampa Schist.

Noteworthy is the absence of sillimanite in the dark metamorphic rocks of the San Emigdio-Tehachapi Mountains, and the abundance of sillimanite just north of the limits of those dark rocks. This is not solely a compositional control because some of the biotite-bearing granofels in the dark metamorphic rocks is compositionally nearly identical to sillimanite-bearing layers in metasedimentary rocks to the north.

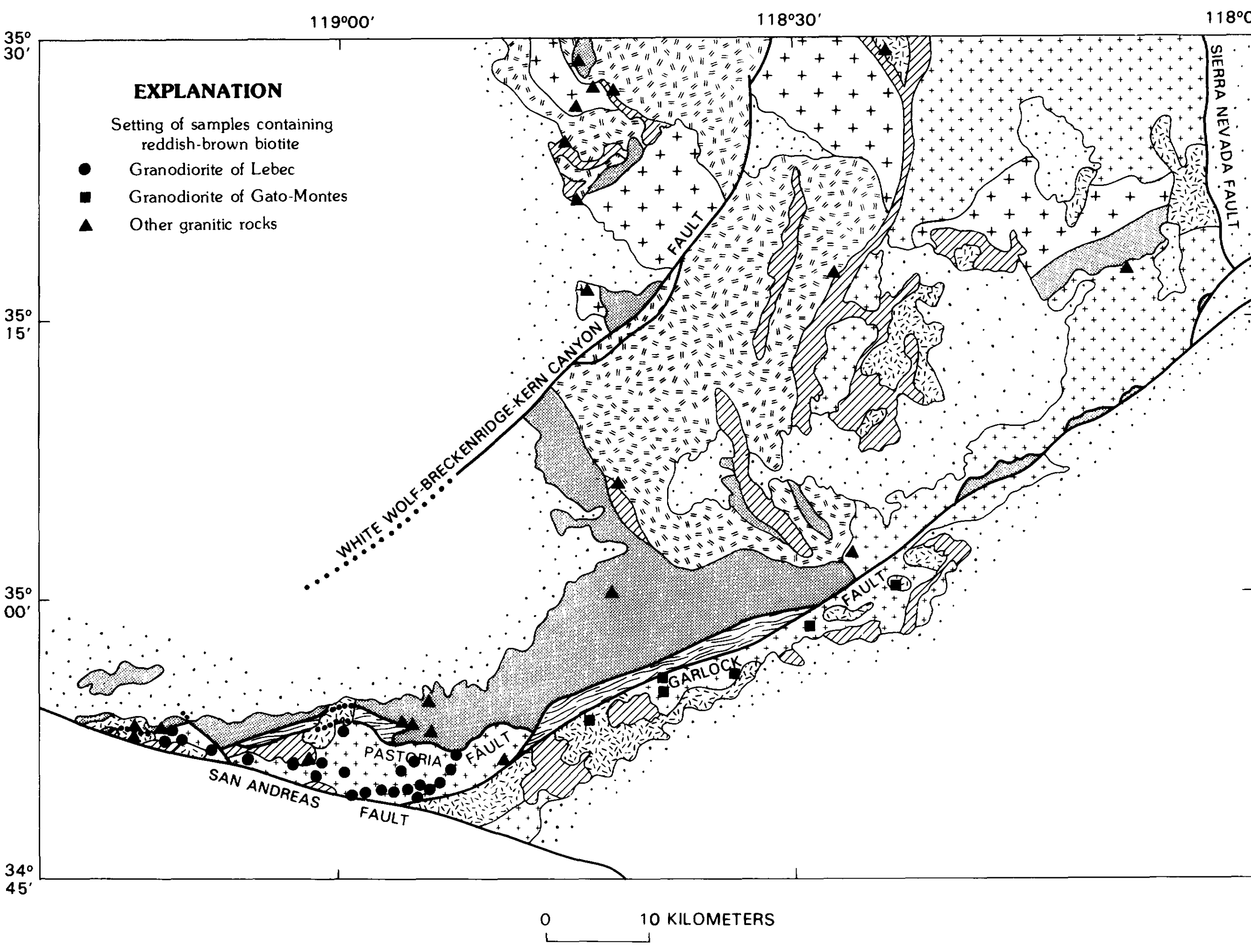


FIGURE 38.—Locations and setting of samples of granitic rocks that contain reddish-brown biotite. Geology same as in figure 37.

The Mountain Park terrane, south of Tehachapi, contains sillimanite-bearing layers, marble, and mica schist, but also amphibolite, labradorite, and brown hornblende that contains coarse poikilitic red garnets with bleached haloes. This terrane may be a transition between an area where sillimanite is stable to the north and a higher grade(?) terrane to the south and west where it is unstable. A similar possible transition zone is suspected along the margin of the San Emigdio-Tehachapi terrane, just south of the Comanche body, where sillimanite is common and associated with coarse-haloed red garnets.

Present distribution of the aluminosilicates indicates that sillimanite and andalusite occur together near the north margin of the study area (where some andalusite is replaced by sillimanite). To the south is a broad belt where sillimanite alone is present; then, rather abruptly, sillimanite disappears and is absent in the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains. Crude isograds could be drawn to reflect these associations, but because the data are sparse, I leave the placement of such boundaries to the discretion of individual readers.

I note that the granodiorite of Lebec (fig. 40) contains abundant sillimanite in its included metamorphic rocks, whereas the presumably correlative granodiorite of Gato-Montes contains no sillimanite and only andalusite in its pendant rocks. Here, once again, the Lebec unit differs from its granite and granodiorite relatives and neighbors.

Holdaway (1971) stated that "The Al_2SiO_5 phase diagram is perhaps the most studied and least well defined silicate phase diagram." This statement pretty well sums up the long-term uncertainty about the stability fields of andalusite and sillimanite and about what their presence or coexistence means in terms of pressure-temperature conditions and, thus, crustal depth. A review of aluminum silicate polymorphs (Ribbe, 1980) cited Richardson and others (1969) and Holdaway (1971) as the most current and reliable experimental phase diagrams

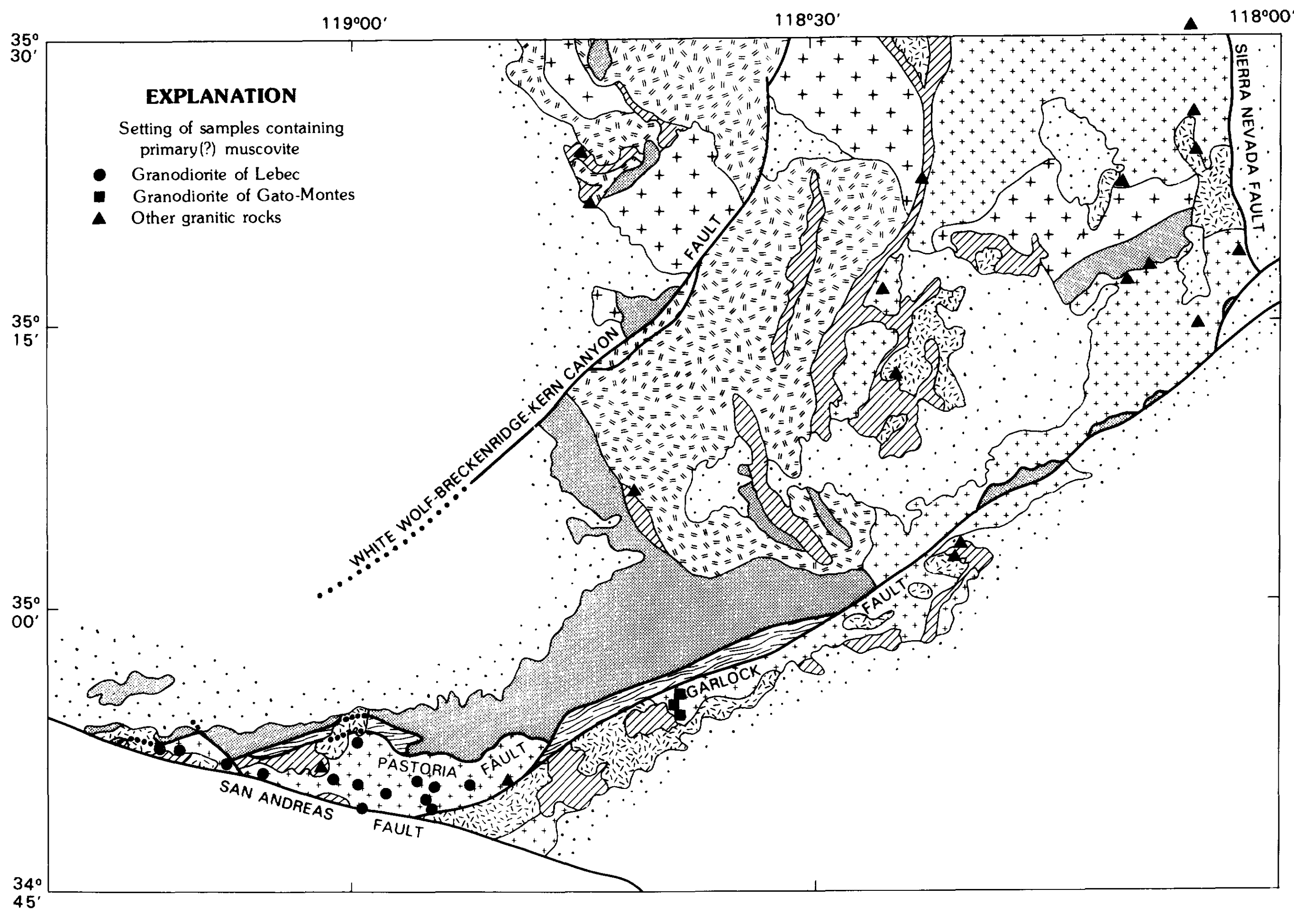


FIGURE 39.—Locations and setting of samples of granitic rocks that contain primary(?) muscovite. Geology same as in figure 37.

for the aluminosilicates and their elusive triple point. Holdaway (1971) placed the triple point at 0.376±0.030 GPa and 501±20 °C. He pointed out that the formation of andalusite indicates a depth of cover of less than 13 km. Richardson and others (1969), however, placed the triple point at 0.55 GPa and 622 °C, values that would increase the crustal depth where andalusite could occur. Warren Hamilton (written commun., 1983) suggested that andalusite alone equates to a crustal depth of formation of no more than 7 to 8 km, that andalusite and sillimanite coexist at depths of 8 to 13 km, and that sillimanite alone points to formation depths of 15 to 18 km.

PREHNITE

Prehnite occurs as lozenges in biotite crystals and in thin secondary veinlets in both granitic and metamorphic rocks (fig. 41). Most of the prehnite is in the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains and the related Cameron, Cinco, and Hoffman mafic terranes. Prehnite is also common in the quartz diorite to tonalite of Antimony Peak. The prehnite is definitely secondary but shows no evidence of local derivation (as, for example, from the alteration of plagioclase or mafic minerals).

Possibly significantly, prehnite appears to be almost exclusively localized in a zone within a few kilometers of the Garlock and Pastoria fault zones. It is tempting to speculate that the prehnite may have leaked up along these fault zones.

Ross (1976d) described much more abundant prehnite veins and lozenges in the western Santa Lucia Range (fig. 1) that appear to be concentrated near the San Gregorio-Hosgri fault zone. The prehnite there is also not a locally derived alteration product and may have leaked up along the fault zone. The source of the prehnite is unknown, but Ross (1976d) speculated that it may be a precipitate of metamorphic fluids (Barnes, 1970; White and others,

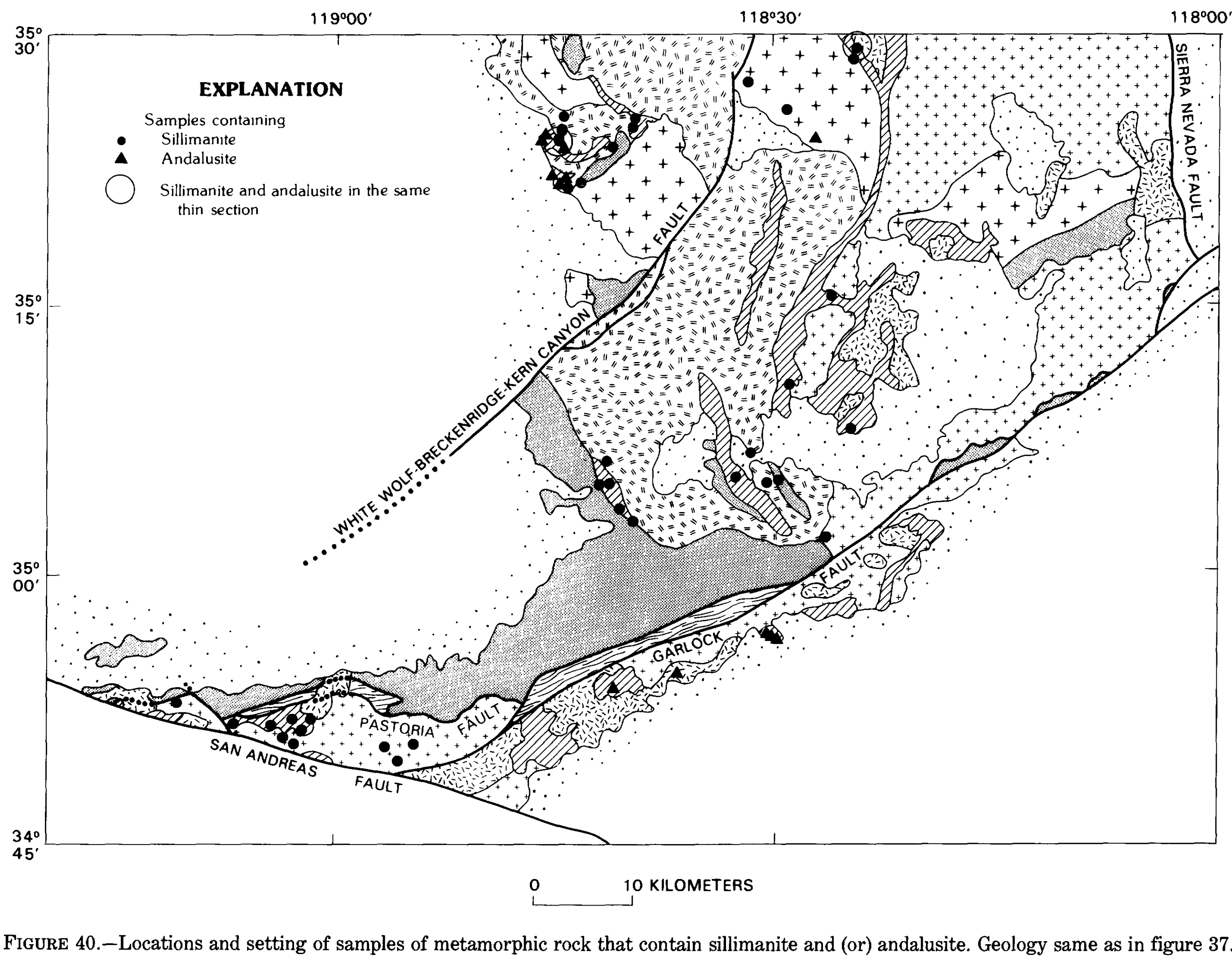


FIGURE 40.—Locations and setting of samples of metamorphic rock that contain sillimanite and (or) andalusite. Geology same as in figure 37.

1973) derived from the metamorphism of underlying graywacke.

Prehnite, hosted largely by granulite and related rocks and concentrated near major fault zones, both in the southernmost Sierra Nevada and the Santa Lucia Range, may be more than just coincidental. Relatively high grade metamorphic rocks and mafic plutonic and metamorphic rocks also host prehnite in Norway (Field and Rodwell, 1967), Germany (Maggetti, 1972), Sweden (Zeck, 1971), and Ireland (Hall, 1965). Could generation of secondary, late, but definitely introduced prehnite in mafic and high-grade rocks be a clue to some special pressure-temperature or chemical conditions that are crustal-depth indicators?

EPIDOTE AND ALLANITE

Discrete, but anhedral, coarse crystals of epidote that look primary are a sparse accessory mineral in a few granitic-rock samples in the southernmost Sierra Nevada. Euhedral allanite in rich-chocolate-brown crystals, as much as 2 mm long, is also widely scattered and commonly rimmed with epidote. Deer and others (1965) observed that this association is common in granitic rocks. Allanite, which has a distinctive red tint, is particularly abundant in the granodiorite of Lebec, where it favors samples that also contain reddish-brown biotite.

Zen and Hammarstrom (1984) noted that magmatic epidote has been found in several localities in the western Cordillera. They suggested that these magmatic, epidote-bearing granitic rocks, which appear to occur on the inboard margins of accreted terranes from northern California to southeastern Alaska, may be high-pressure facies of typical calc-alkaline granitic rocks. Thus, magmatic epidote may be an indicator of relatively deep crustal levels.

TOURMALINE

Accessory tourmaline was found in one granitic rock and in several metamorphic specimens. Excluded from the following discussion is the black tourmaline that was

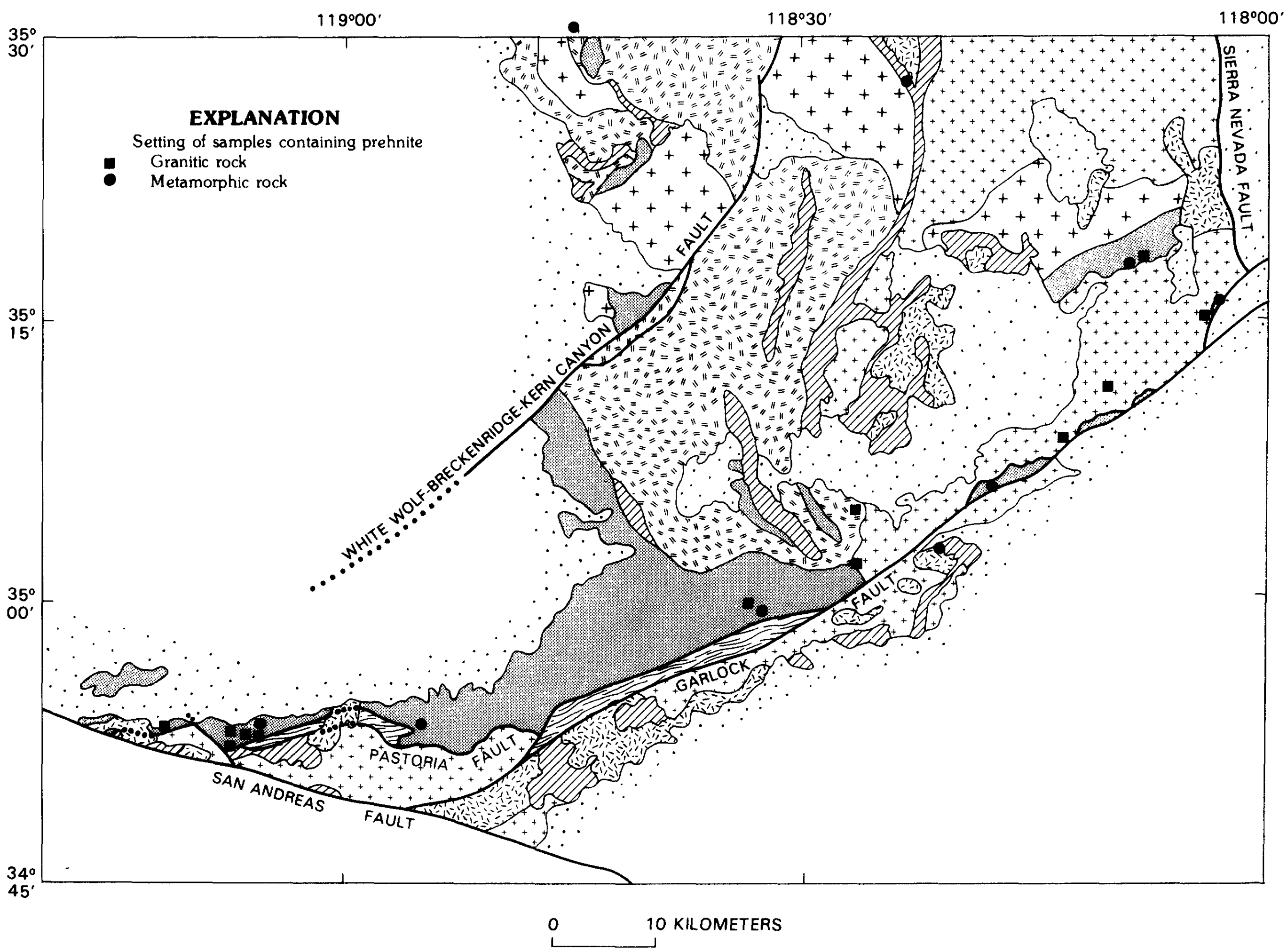


FIGURE 41.—Locations and setting of samples containing prehnite. Geology same as in figure 37.

found locally in pegmatite dikes and was also observed as core material in polka-dot dikes along Rancheria Road. Tourmaline is a rare but widely scattered accessory mineral throughout most of the metamorphic-rock units in the study area.

Traces of blue tourmaline were found in one specimen of average-appearing tonalite of Hoffman Canyon that did, however, contain 4 percent clinopyroxene.

Tourmaline is most common in the dark-gray, micaceous, andalusite-bearing schist and hornfels of the Pampa Schist. Most of the tourmaline is in discrete, dark-yellowish-orange to light-brown crystals. Within the outcrop area of the metavolcanic rocks of French Gulch, dark-gray, strongly pleochroic tourmaline is widely scattered.

Tourmaline occurs sparsely in the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains; distinctive reddish-brown, strongly pleochroic tourmaline was also found in one specimn of the felsic gneiss. The Rand(?) Schist sliver in the San Emigdio Mountains contains light-olive tourmaline in some specimens.

Locally, the impure quartzite and schist layers in the Salt Creek, Keene, and Bean Canyon metamorphic-rock units contain scattered light-olive to light-brown tourmaline. Possibly the most unusual occurrence of tourmaline is in a light-colored metasiltstone layer in the Bean Canyon Formation in Gamble Spring Canyon. This rock is liberally sprinkled with aggregates and poikilitic masses of tourmaline, as large as 2 to 3 mm across, that are zoned in light shades of brown and green.

POLKA-DOT DIKES

In the northwestern part of the study area, along Rancheria Road north of California Highway 178, conspicuous felsic dikes that intrude the tonalite of Bear Valley Springs contain abundant dark, round clots with white haloes (fig. 42). Similar dikes have been described in a broad region from the Salton Trough northward to the La Panza Range (Ehlig and Joseph, 1977). I examined the polka-dot rocks in the La Panza Range and found them to be strikingly similar in physical appearance to those along Rancheria Road. S.E. Joseph (oral commun., 1979) suggested, however, that because of differences in the Rb-Sr systematics, the dikes of Rancheria Road and the La Panza Range were derived from different source materials.

The Rancheria Road locality exposes several northeast-trending dikes, as much as 2 m thick. One dike is exposed for about 300 m, but the whole zone appears to have a strike length of at least 2 km. The polka-dot cores are almost invariably spherical, although some are elongate, and rarely they are angular. The largest cores have a diameter of 8.5 cm, but most are smaller and rarely exceed 5 cm in diameter. The cores are characterized by abundant muscovite and olive to light-brown biotite, as well as abundant quartz and sodic plagioclase in sugary mats. Scattered pink garnets are present in some cores. A few small cores are composed of poikilitic black tourmaline and quartz plus plagioclase. The white rims (haloes) around the cores are approximately 1 cm thick. They are distinguished by an almost complete absence of biotite and muscovite, and they commonly contain less K-feldspar than the matrix of the polka dots. The matrix is fine grained and is partly aplitic. Plagioclase in the matrix is in subhedral to embayed crystals that are in part weakly zoned in the range An_{25-30}. K-feldspar is abundant and weakly grid twinned. Quartz ranges from millimeter-size grains to small globules. Olive-brown biotite is rather common and is evenly scattered through these rocks. Muscovite is also sparsely present.

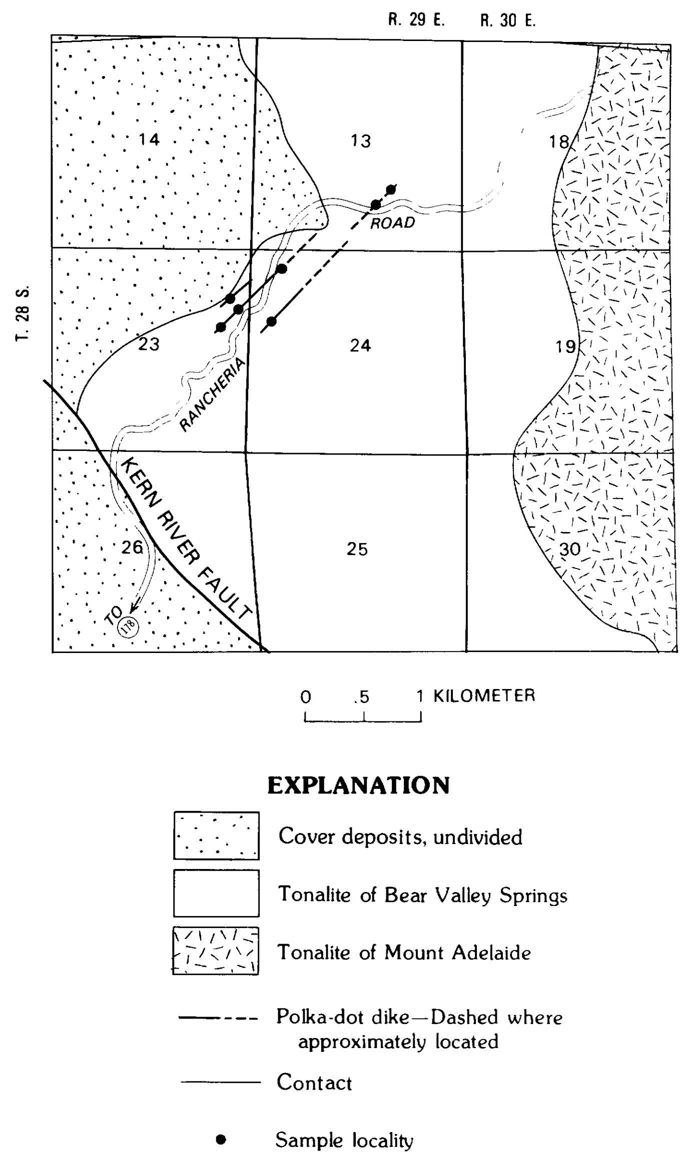


FIGURE 42.—Locations of polka-dot dikes. Base from Rio Bravo Ranch 1:24,000-scale quadrangle.

Just north of the northernmost exposure of the dike rocks (fig. 42), angular, blocky inclusions of juiced-up metadiabase(?) are exposed; these inclusions are composed of a mat of plagioclase laths, widespread interstitial quartz, and abundant brown biotite and pale-green hornblende crystals. The hornblende ranges from lathlike crystals to small equant clots, 1.5 mm across, that may have replaced single phenocrysts. The metadiabase inclusions are probably part of the Pampa Schist. No transition between the inclusion material and the polka dots of the felsic dikes has been seen, but a few isolated biotite-rich clots (unrimmed) nearby hint of some relation. Although exposures are too sparse to test whether the metavolcanic inclusions are a possible source material for the polka-dot cores, the mica-garnet polka-dot cores do, at least, suggest a metamorphic parentage.

A few scattered polka dots were also seen in a felsic dike in the Pampa Schist, just north of a thin tongue of tonalite of Bear Valley Springs in the SE¼ sec. 34, T. 29 S., R. 30 E. The cores are only about 1 cm in diameter, and they are composed of a concentration of biotite and muscovite and some pink garnet. The bleached rims are about 1 cm thick. The polka-dot matrix is aplite composed of plagioclase, K-feldspar, quartz, and very scattered biotite and muscovite. Pink garnets are scattered in both the dike and the bleached rims.

About 8 km south-southwest of Claraville, a float specimen from a fine-grained felsic dike contains scattered clots, as large as 2 cm across, of poikilitic black tourmaline, minor muscovite, quartz, and feldspar. These cores are rimmed by 1-cm-thick bleached haloes of quartz and feldspar. The dike in which the clots are set is a felsic aplitic body with scattered biotite crystals. The dike weathers red; that color accentuates the white rims and black cores of the polka dots. These polka dots are nearly identical to the sparse tourmaline polka dots at the Rancheria Road locality.

An unusual occurrence of dikes with small polka dots was found in a small canyon in the west-central part of sec. 27, T. 11 N., R. 14 W., west of Tehachapi-Willow Springs Road (lat 35°01′ N., long 118°21′ W.). There, fine-grained dikes that cut the granodiorite of Gato-Montes are peppered with individual magnetite crystals and aggregates that range in size from 2 to 5 mm across. A few polka dots are cored with biotite clots instead of magnetite crystals; these cores are rimmed by bleached haloes, 0.5 to 1 cm thick, in which dark minerals are virtually absent. The rims are much lighter in color than those in the aplite dikes, which are liberally sprinkled with tiny biotite flakes.

Polka-dot dikes that have some common characteristics (biotite and tourmaline cores, bleached haloes, felsic, fine-grained matrix dike rocks) intrude four different units that are widely separated in the study area. These data suggest that the polka-dot type may be more common, more widespread, and less diagnostic than earlier reports suggested.

POSSIBLE CORRELATION OF THE GRANODIORITES OF LEBEC AND GATO-MONTES

The granodiorites of Lebec and Gato-Montes are nearly identical in field appearance. Both are medium grained and peppered with small black biotite crystals and lesser hornblende. Scattered coarser biotite flakes are also common in both masses. Particularly distinctive in both units are red-stained cores of hornblende crystals, which are clues to skeletal ghosts of clinopyroxene. Small mafic inclusions are present, though sparse, in both masses. Accessory sphene, allanite, and muscovite are abundant in the Lebec mass and are also present in the Gato-Montes mass, where they (particularly muscovite) occur less commonly. Sugary, fine-grained dikes and masses are present in both granodiorites but are more abundant in the Lebec rock type.

The modal averages and fields of the two granodiorites are virtually identical; however, the Gato-Montes mass contains slightly more plagioclase and hornblende and slightly less K-feldspar, quartz, and biotite than the Lebec mass (figs. 16, 27).

Major-element chemistry of the two masses is consistent with the modal similarity (table 27). The averages of four analyses of the Gato-Montes mass and of five analyses of the Lebec mass are nearly equal. Al_2O_3 and FeO are 0.6 and 0.3 percent higher, and SiO_2, Fe_2O_3, and Na_2O are 0.4, 0.3, and 0.5 percent lower, respectively, in the Lebec average. Other oxides are within 0.2 percent. The chemical analysis of each rock type that is nearest the modal average of each rock type (sample DR–690 for the Lebec mass and sample DR–3534B for the Gato-Montes mass) shows a close similarity in major-element content. Trace-element concentrations are also virtually identical for the two masses. The presence of Gd, Sn, and Zn in the Gato-Montes samples determined in 1978, and the absence of these elements in the Lebec samples determined in 1968 (see table 31), reflect increased sensitivity of the analyses for these elements in the 1978 determinations (R.E. Mays, oral commun., 1979), not a difference in trace-element content between the two granodiorites. Similarities in the levels of all other trace elements suggest that with new analyses these elements would probably be found in the Lebec samples.

The coarse-grained granite of Brush Mountain intrudes the Lebec mass. Rocks of similar field appearance and modal composition (granite of Tejon Lookout) intrude the granodiorite of Gato-Montes. Although the Brush Mountain and Tejon Lookout masses are nearly identical (see table 31), modally (figs. 25, 30) and texturally, such low-melting-trough biotite granites are also very common rock

TABLE 27.—*Chemical comparison of the granodiorites of Gato-Montes and Lebec and associated granite bodies*

[All values in weight percent]

	Granodiorite of Gato-Montes					Granodiorite of Lebec						Granite of Tejon Lookout				Granite of Brush Mountain			
Sample	3356	3534B	3733A	3756A	Average	686	690	698	3088	3181	Average	3401	3466	3752	Average	726	3005	3221	Average
SiO_2	66.8	68.1	65.4	71.4	67.9	68.0	68.5	67.8	65.4	68.0	67.5	74.8	76.6	75.5	75.6	75.6	74.8	75.7	75.4
Al_2O_3	16.0	15.3	17.0	14.2	15.6	16.2	16.0	15.7	16.8	16.9	16.2	12.5	12.7	12.8	12.7	13.2	13.6	13.2	13.3
Fe_2O_3	.64	.94	1.3	.77	.9	.26	.45	.31	.9	.9	.56	1.4	1.2	1.1	1.2	.24	.71	.85	.60
FeO	2.7	2.5	2.6	1.4	2.3	2.4	2.5	2.9	2.7	2.4	2.6	.50	.48	.28	.42	.96	.96	.24	.72
MgO	1.1	.97	1.3	.51	1.0	1.0	1.0	1.6	1.1	1.1	1.2	.71	.23	.08	.34	.17	.08	.14	.13
CaO	3.6	3.4	4.3	1.9	3.3	3.2	3.2	3.3	4.1	3.4	3.4	1.4	.97	.60	.99	.87	.84	.43	.71
Na_2O	4.0	4.0	4.3	3.6	4.0	3.9	3.5	3.2	3.6	3.2	3.5	4.2	3.8	2.3	3.4	3.7	3.0	3.8	3.5
K_2O	3.2	3.1	2.1	4.3	3.2	3.4	3.4	3.2	2.8	2.9	3.1	4.8	4.2	6.9	5.3	4.4	4.4	4.3	4.4
H_2O^-	.13	.14	.16	.12	.14	.08	.06	.08	.2	.15	.11	.19	.25	.13	.19	.04	.06	.03	.04
H_2O^+	.71	.88	.92	.64	.79	.65	.55	.73	.68	.77	.68	.68	.57	.42	.56	.54	.49	.61	.55
TiO_2	.68	.60	.78	.35	.60	.54	.50	.54	.76	.60	.59	.13	.12	.08	.11	.11	.09	.08	.09
P_2O_5	.23	.19	.27	.13	.21	.16	.15	.15	.23	.18	.17	.07	.06	.05	.06	.04	.04	.04	.04
MnO	.05	.08	.05	.03	.05	.07	.07	.10	.03	.04	.06	.08	.04	---	.04	.07	.02	.02	.04
CO_2	.02	.06	---	---	.02	.19	.11	.30	.09	.05	.15	.01	.01	---	.01	<.05	.03	.02	.02
Total	99.9	100.3	100.5	99.4	100.0	100.1	100.0	99.9	99.4	99.8	99.8	101.5	100.8	100.2	100.9	99.7	99.1	99.5	99.5

types and are present in many granitic suites. The implications of these possible correlations are discussed below in the subsection entitled "Garlock Fault Displacement."

SELECTED MODAL-MINERAL PLOTS

Modal averages for various minerals and specific gravity from each plutonic unit were plotted against each other to test the consistency of the modal data (fig. 43). Mainly, I wanted to see whether the modes of the more mafic units are compatible with those of the rest of the plutonic rocks. Plots of the amounts of quartz and total dark minerals against specific gravity, as well as the plot of quartz against total dark minerals, are strikingly linear and are consistent for all units. The plot of quartz against plagioclase shows somewhat more scatter but still has a visible trend. The plot of hornblende against biotite shows all the units to be along a rather consistent trend except for the Bear Valley Springs mass and its suspected relatives, the Tehachapi Mountains, Antimony Peak, and hypersthene-bearing tonalite units, all of which are strongly enriched in hornblende.

The conspicuousness of hornblende in these darker units is well illustrated by figure 44. In most of the plutonic-rock units, the amount of hornblende is generally 5 to 10 percent less than that of biotite; units containing less than 10 percent biotite contain little or no hornblende. Moore (1963) used a similar plot for several granitic bodies in the east-central Sierra Nevada. He noted that in all the granitic bodies, biotite is more abundant than hornblende by about 4 percent. The plot for the southernmost Sierra Nevada shows a larger modal gap between amounts of hornblende and biotite for most plutons, in comparison with Moore's area, and hornblende-dominant plutonic-rock units characterize the more plagioclase rich end of the plot.

RADIOMETRIC-AGE DATA ON THE PLUTONIC ROCKS

Although radiometric dating has not been done systematically for the plutonic rocks of the southernmost Sierra Nevada, several K-Ar and Rb-Sr determinations have been made. The available radiometric data are summarized in table 28. Sample locations are plotted on figure 45, which also identifies the map units of plate 1 from which they were collected. Most of the radiometric work was done before the present reconnaissance study. Therefore, the following notes record my observations for each radiometric sample locality of the characteristics of each sample, and my estimates of how typical the sample is of the basement-rock units for which each radiometric age was determinated. Sample SR-8-73, from the northeastern part of the study area, is a typical representative of the peppery granodiorite of Claraville. Sample SR-9-73

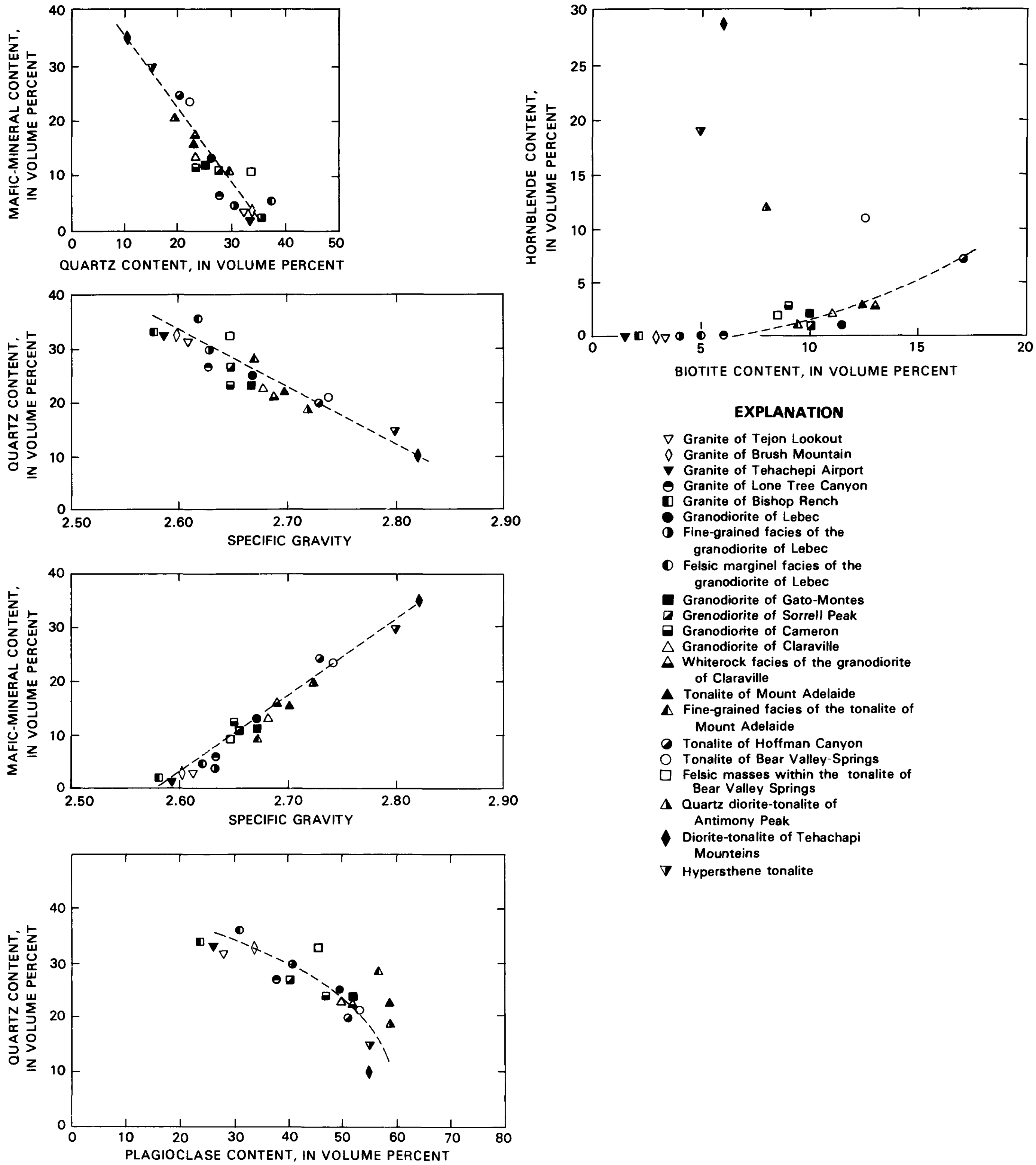


FIGURE 43.—Some ratios between modal averages of the plutonic-rock units. Trendlines were visually estimated.

is somewhat finer grained than sample SR–8–73 and is also porphyritic, but it falls well within the range of textures for samples of the granodiorite of Claraville. Sample SR–10–73 was collected within the tonalite of Hoffman Canyon, which varies considerably in composition. Most of the outcrop near the sample locality contains abundant hornblende and abundant mafic inclusions. There is some controversy as to whether the Hoffman Canyon mass is a facies of the Claraville unit or is a separate tonalite intrusive body. The initial Sr-isotopic ratio favors a facies relation, but more data are needed. Sample SR–12–73 (141) (the only sample from south of the Garlock fault) is from a small dark body of hornblende-biotite granodiorite whose relation to the surrounding granite of Tejon Lookout is uncertain. From the field relations, I had postulated that this body appeared to be a contaminated inclusion of the Gato-Montes mass, but the initial Sr-isotopic ratio argues against this interpretation. Sample SR–15–73 (140) is from the felsic facies of the tonalite of Bear Valley Springs, and sample SR–14–73 is from average tonalite of Bear Valley Springs. The sample analyzed by J.L. Morton is from a hypersthene-bearing rock in the Bear Valley Springs mass. Such rocks resemble the rest of the Bear Valley Springs mass but contain a few percent of hypersthene. I am uncertain what rock was sampled along Grapevine Canyon (samples 142, 153, Turner) because a wide variety of felsic to mafic gneiss and much felsic granitic rock are present there. Sample DR–698 (from near the San Andreas fault) is a representative sample of the biotite granodiorite of Lebec. Samples DR–671B and DR–1182C are representative of the anorthositic gabbro and hornblende-quartz gabbro that occur as basement islands surrounded by Cenozoic deposits near the west margin of the study area. Sample DR–1169A is a sample of quartz gabbro from within the main anorthositic gabbro outcrop, and sample DR–1189C is from a coarse-grained, hornblende-rich pegmatite in the anorthositic gabbro.

Sharry (1982) reported an Rb-Sr age of 86.1 ± 13.2 m.y. on two-pyroxene granulites from the Tehachapi Mountains. This age is very close to the K-Ar age determined by J.L. Morton (table 28) for a specimen of hypersthene-bearing tonalite of Bear Valley Springs south of Cummings Valley (biotite, 85.9 ± 2.6 m.y.). Sharry (1981) also noted an initial Sr-isotopic ratio on the granulite of 0.70513 ± 0.00008, which is comparable to the initial Sr-isotopic ratios of 0.7050 and 0.7055 determined on two specimens of the tonalite of Bear Valley Springs by Kistler and Peterman (1978).

Sharry (1981) suggested that the age of 86 m.y. on the granulite reflects the time of the end of metamorphism brought about by the underthrusting of cold Pelona Schist beneath the granulite terrane. I originally thought that the age of 86 to 88 m.y. dated emplacement of the tonalite of Bear Valley Springs, but concordant Pb-U zircon ages indicate an intrusive age of about 100 m.y. for the tonalite of Bear Valley Springs (Sams and others, 1983). The 86- to 88-m.y. age may reflect a time of rapid uplift and cooling that set both the K-Ar and Rb-Sr clocks.

Rb-Sr whole-rock data were recently (1981) obtained from samples of the granodiorites of Gato-Montes (samples DR–3356, DR–3534B, DR–3733A, DR–3756A, table 28), Claraville (samples DR–4093, DR–4472), and Lebec (sample DR–698) collected during the present

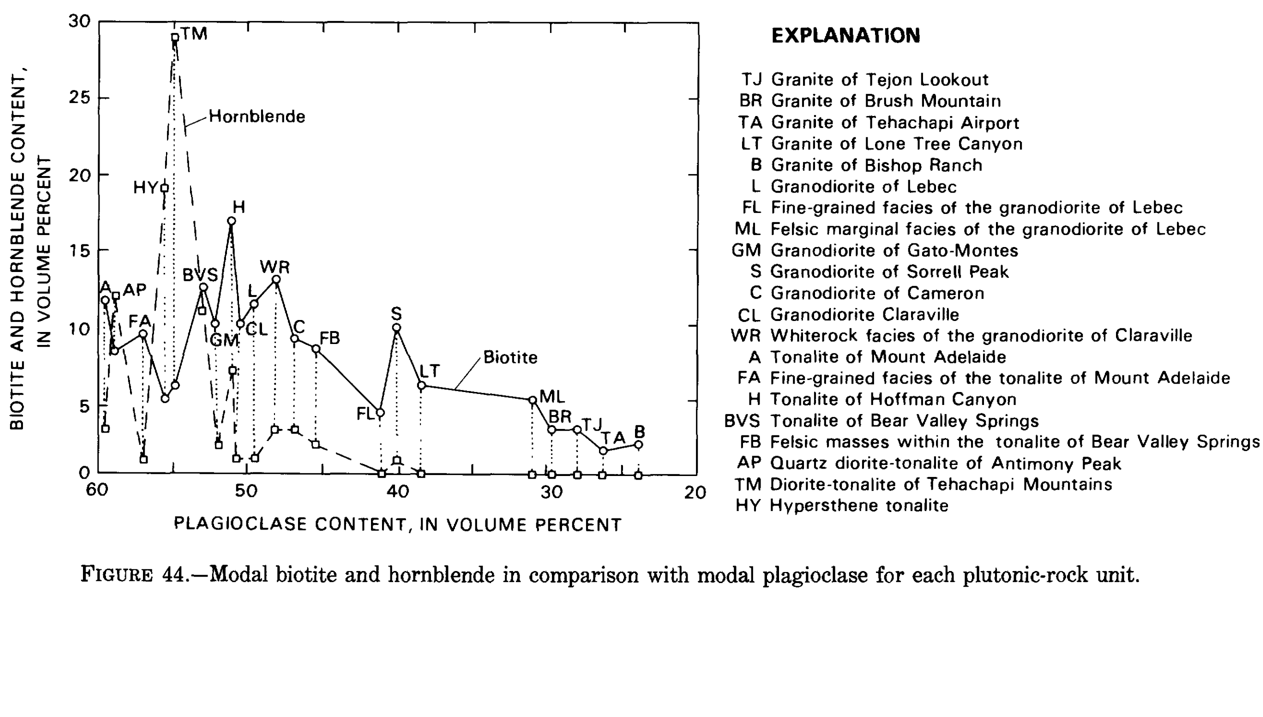


FIGURE 44.—Modal biotite and hornblende in comparison with modal plagioclase for each plutonic-rock unit.

TABLE 28.—*Summary of available K-Ar and Rb-Sr radiometric data for plutonic rocks of the southernmost Sierra Nevada*

[Whole-rock determinations except where mineral is listed. n.a., not available]

Sample (fig. 45)	Mineral	K_2O (wt pct)	Na_2O (wt pct)	Radiogenic ^{40}Ar (pct)	K-Ar age (m.y.)	Rb (ppm)	Sr (ppm)	Rb/Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$	r_i	Rb-Sr age (m.y.)
[1]140	Biotite-------	8.76	---	81	81.2							
[1]141	do----------	8.68	---	87	80.4							
	Hornblende----	1.13	---	83	81.4							
[1]142	Biotite-------	8.86	---	79	86.1							
[1]153	Hornblende----	.39	---	57	77.4							
Morton[2] (DR-3596)	Biotite-------	9.16	---	86	85.9±2.6							
	Hornblende----	.937	---	68	88.1±2.6							
Turner[3]	Biotite-------	n.a.	---	n.a.	87.3							
[4]SR-8-73	do----------	9.36	3.82	94	78.7±2.0	98	737				.7075	90
[4]SR-9-73	do----------	9.16	3.91	70	74.8±1.9	81.8	695				.7080	90
[4]SR-10-73		2.84	3.70			89.6	634				.7070	90
[4]SR-12-73 (141)		2.45	4.56			80.1	612				.7058	81
[5]SR-14-73		2.00	3.91			55.0	347				.7050	80
[5]SR-15-73 (140)		1.70	3.68			53.8	341				.7055	81
[5]DR-671B						<2	147				.7031	200
	Plagioclase					4.2	293				.7033	200
[5]DR-698		3.2	3.2			127	376	.338	.978	.70930±0.00011	.7078	85
[6]DR-1169A	Hornblende----	.204	---	55	134±4							
[5]DR-1182C						8.8	228				.7038	200
	Hornblende----	.190	---	26	165±6							
[6]DR-1189C	do----------	.107	---	41	207±10							
[7]DR-3023						<5	603			.70337±0.00010		
[7]DR-3097						2.4	261	.009	.026	.70364±0.00012		
[7]DR-3098A						16.6	289	.057	.166	.70446±0.00003		
[7]DR-3356						80.4	632	.127	.368	.70833±0.00004		
[7]DR-3534B						129	416	.310	.897	.70895±0.00003		
[7]DR-3733A						82.7	618	.134	.387	.70840±0.00005		
[7]DR-3756A						132	370	.357	1.032	.70991±0.00005		
[7]DR-4093						81.3	652	.125	.361	.70864±0.00003		
[7]DR-4472						93.1	665	.140	.405	.70904±0.00007		

[1]Evernden and Kistler (1970).
[2]Hypersthene-bearing tonalite west of Tehachapi (J.L. Morton, written commun., 1979).
[3]Rock type unknown, from Grapevine Canyon (D.L. Turner, written commun., 1975).
[4]Kistler and Peterman (1978).
[5]Kistler and others (1973).
[6]Ross and others (1973).
[7]These data indicate that the samples of the quartz diorite to tonalite of Antimony Peak (DR-3023) and of the diorite to tonalite of the Tehachapi Mountains (DR-3097, DR-3098A) are from a mafic source. The granitic samples (DR-698, DR-3356 through DR-4472) all have an apparent Rb-Sr age of about 85 m.y., with initial $^{87}Sr/^{86}Sr$ values ranging from 0.7079 to 0.7085 (R.W. Kistler, written commun., 1981).

study. These data suggest that all three granodiorites have an apparent age of about 85 m.y., based on the isochron range shown in figure 46.

Although a good start has been made in radiometric studies, clearly more needs to be done, now that a geologic base is available. All the radiometric ages so far determined are Late to mid-Cretaceous except those for the gabbroic rocks of the Eagle Rest Peak area, which appear to be Late Jurassic, at least in part. Is the entire width of the Sierra Nevada at this latitude part of the latest intrusive epoch of Evernden and Kistler (1970), or has the southernmost Sierra Nevada been influenced by the vast area of massive reheating that affected the adjacent Mojave block (Miller and Morton, 1980)? Answers to this and other fundamental questions require more radiometric data in the southernmost Sierra Nevada.

METAMORPHIC GRADE OF THE HORNBLENDE-RICH TERRANES

The hornblende-rich mafic terranes of the southernmost Sierra Nevada, on the basis of their mineralogy, have aspects of both upper amphibolite- and lower granulite-grade metamorphism. Unquestioned hypersthene granulite-facies rocks are present in the southernmost Sierra Nevada and are particularly well developed in the Tunis Creek area. Granofels that may or may not reflect granulite-facies conditions is more widespread. The absence of sillimanite throughout the San Emigdio-Tehachapi mafic terrane may reflect metamorphic conditions higher than amphibolite grade, in which sillimanite is no longer stable, for that entire mafic terrane.

Widespread hypersthene, brown hornblende, and coarse red garnets in the mafic terranes suggest relatively deep and high-grade (granulite) metamorphic conditions. Nonetheless, brown biotite, muscovite, and K-feldspar are rather common in the San Emigdio-Tehachapi terrane away from the Tunis Creek hotspot, green hornblende in discrete primary crystals is also widespread, and many of the mafic rocks do not contain pyroxene—all features pointing to amphibolite-grade conditions.

Problems of grade assignment in high-grade metamorphic terranes are widespread. Turner and Verhoogen (1951) noted that in many classic granulite terranes where the paragenesis of the granulite facies is clearly recognizable, the paragenesis is to some extent obscured by the presence of considerable hornblende or biotite. They cited, as an example, the famous Lewisian area of Scotland, where regions of normal amphibolite facies are associated with minor rocks with the granulite pair diopside-hypersthene. Buddington (1939) observed "characters of both amphibolite and granulite facies" in high-grade rocks of the Grenville Series of the Adirondack Complex. He interpreted the association of supposedly unstable biotite and hornblende in the granulite facies to indicate (1) disequilibrium (retrograde metamorphism), (2) transitional conditions between granulite and amphibolite facies, or (3) insufficient water for all the iron or magnesium to be accommodated by the hydrous minerals biotite and hornblende. This concept of localized dryspots is essentially what Compton (1960) postulated to explain the hypersthene-bearing rocks of the western Santa Lucia Range.

In the southernmost Sierra Nevada, retrograde metamorphism of the mafic metamorphic rocks, most recognizable in the pale-green actinolitic or cummingtonitic amphibole, is widespread and locally impressive. Nevertheless, much of the amphibolite is surprisingly fresh and does not resemble retrograde pyroxene rock. The concept of local dryspots is somewhat more difficult to evaluate, from my data, but it is certainly a valid possibility to explain local granulite in a largely amphibolitic terrane. From my own observations and petrographic study of the rocks of the southernmost Sierra Nevada, I favor concept 2 of Buddington (1939): "transitional conditions between granulite and amphibolitic facies." A not unreasonable

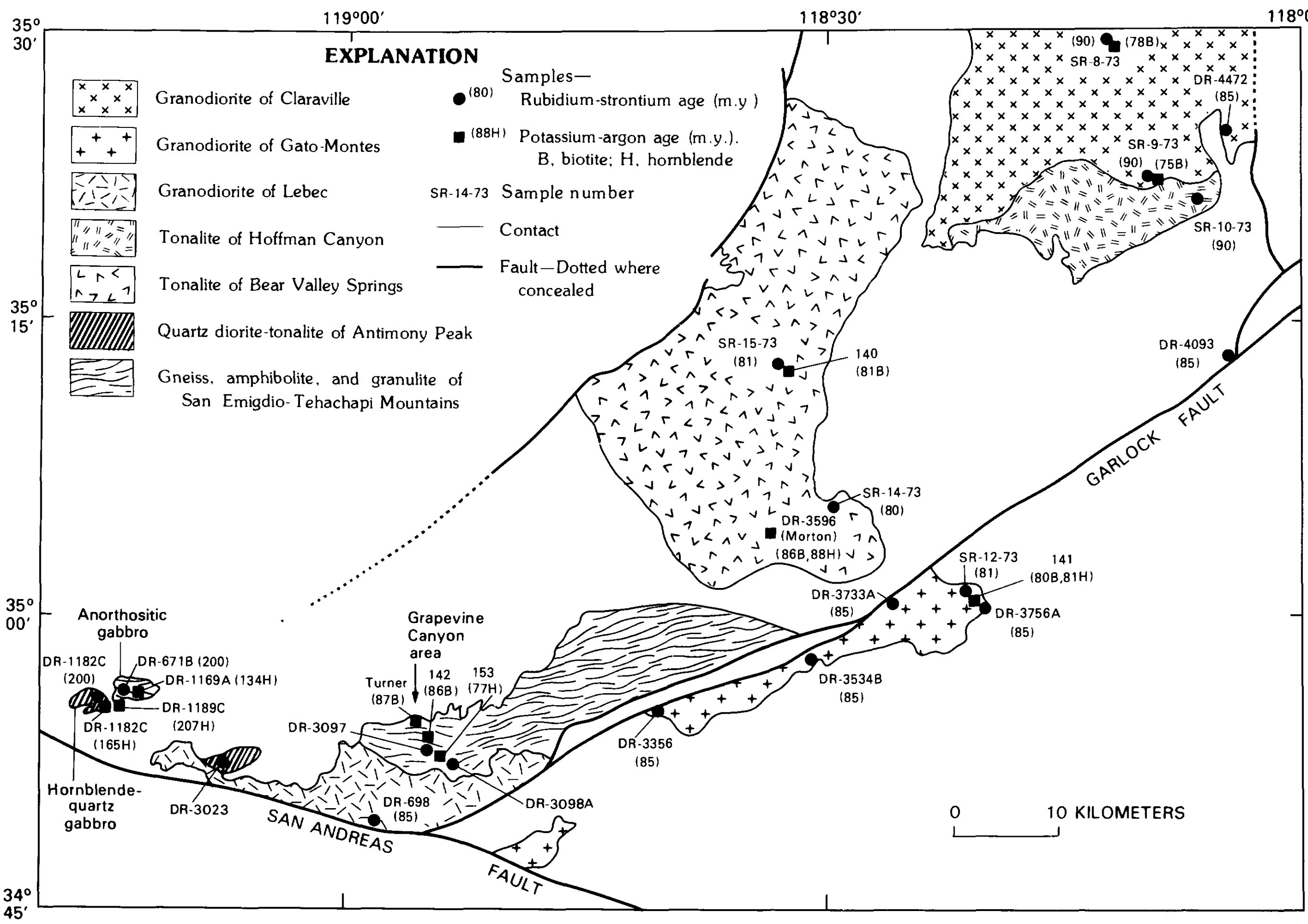


FIGURE 45.—Locations of samples of plutonic rock on which K-Ar and (or) Rb-Sr radiometric ages have been determined.

extrapolation from the distribution of granulite-grade indicators (hypersthene, brown hornblende, red biotite, and coarse red garnets), discussed earlier in this report, would suggest a vast terrane of granulite-facies rocks in the southernmost Sierra Nevada. However, I cannot ignore the presence of the amphibolite-grade indicators, green hornblende and brown biotite, and the absence of pyroxene and red garnets in large areas.

TECTONIC FRAMEWORK

The major emphasis of this report is on the characteristics and distribution of the basement-rock units. Geologic structure is discussed only for those gross features that help define the tectonic framework and for special features that might be useful in correlating the southern Sierra Nevada basement with tectonically separated fragments of this batholithic block. Detailed structural features of the basement-rock units that bear on such problems as the presence or absence, age, and style of multiple deformations are left for study by other workers. I do hope that this report can serve as a detailed reconnaissance base for such future studies. The following sections, which deal with various structural-framework features, are more or less self-contained topics that I feel show special characteristics of the basement. No attempt has been made to combine these topics into a broadly integrated tectonic model—that, too, I leave for future study.

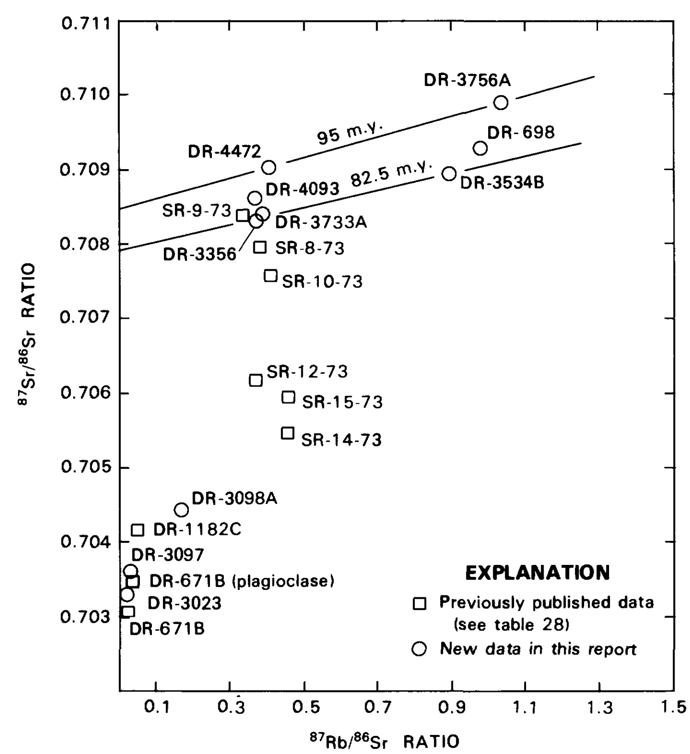


FIGURE 46.—$^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$ ratio for selected radiometrically dated samples (see table 28). Apparent ages are indicated for samples of the granodiorites of Gato-Montes, Claraville, and Lebec.

FAULTS THAT OFFSET THE BASEMENT SIGNIFICANTLY

The southernmost Sierra Nevada is cut off on the south by the San Andreas and Garlock faults, both of which have undergone major strike-slip dislocations. At this latitude, the Sierra Nevada block is bounded on the east by the Sierra Nevada fault, a presumed basin-and-range normal fault. This fault zone is well exposed at the head of Poleline Canyon near the Los Angeles Aqueduct road in the NE¼ sec. 7, T. 30 S., R. 37 E (lat 35°20′ N., long 118°02′ W.). The combined trend of the Garlock and Sierra Nevada faults essentially parallels the combined trend of the Kern Canyon, Breckenridge, and White Wolf faults (pl. 1). The significance, if any, of this parallelism is unknown at this time. Moore and du Bray (1978) suggested right-lateral strike-slip movements of as much as 13 km on the northern section of the Kern Canyon fault zone (north of the study area). I noted similar offsets along the Breckenridge fault zone and the southern section of the Kern Canyon fault zone. Focal-mechanism determination for aftershocks from the Kern County earthquakes of 1952 variously indicate left-lateral, right-lateral, and dip-slip movements on the White Wolf fault, with left-lateral strike slip predominating (Cisternas, 1963). I am presently trying to resolve this movement dilemma along the Kern Canyon, Breckenridge, and White Wolf faults, which together mark a significant basement break in the Sierra Nevada block.

The west margin of the Sierra Nevada basement block at lat 35° to 35°30′ N. is largely an irregular erosional surface overlapped by deposits of the sediment-filled San Joaquin Valley, and so its structure is uncertain. However, two distinct east-west- to northwest-trending fault breaks occur north of the White Wolf fault zone at this latitude. (1) The Edison fault appears to be a steeply dipping normal fault downthrown to the north, judging by the abrupt termination of basement outcrop and subsurface relations in the area of the Edison oil field (Beach, 1948; Smith, 1964). The relations of the east end of this fault with the White Wolf-Breckenridge-Kern Canyon fault zone is a puzzle that awaits resolution. (2) The northwest-trending Kern River fault (not to be confused with the Kern Canyon fault) is also a normal fault but is downthrown on the southwest side. It is marked by a conspicuous, faceted faultline scarp that dominates the scene where California Highway 178 enters the Kern River canyon. The Kern River fault causes a notable step westward of the basement terrane of the Sierra Nevada block (pl. 1). The interface between the Sierra Nevada basement and the overlapping, younger sedimentary deposits suggests similar westward steps north of the Kern River fault that may be fault controlled. Although faults have not been mapped along all these steps (Smith, 1965), their pattern suggests an analogy with that of the Kern River fault.

All the above-mentioned faults have been mapped, named, and known for many years. In contrast, a probable northwest-trending fault with postulated right-lateral strike-slip offset in the western San Emigdio Mountains appears to have been largely overlooked. Crowell and others (1964, pl. 1) did show a dashed fault in about the correct position for part of the trend of this fault, but no sense of movement was suggested. I tentatively name this fault the "San Emigdio fault"; its delineation was based on the abrupt termination of several basement-rock units along a common northwest-trending line. The fault is overlapped by Tertiary sedimentary rocks; mapping shows no northwest-trending faults in that area (T.W. Dibblee, Jr., and T.H. Nilsen, written commun., 1973). The proposed San Emigdio fault thus appears to be an old, pre-Tertiary basement structure. The northwestward continuation of the San Emigdio fault essentially coincides with the east outcrop limit of mafic ophiolitic(?) rocks of the Eagle Rest Peak area.

The Pastoria fault zone, which trends east-west across the San Emigdio Mountains, also appears to be an important structural break in the basement. The magnitude, age of displacement, and sense of movement of this fault zone are still open to speculation. This fault zone deserves a more detailed investigation. The fault zone is discussed at somewhat greater length below in the section entitled "Pastoria Thrust Controversy."

The two faults that most strongly affect the Sierra Nevada block are the San Andreas and Garlock faults. The San Andreas fault, with its well-documented movements of hundreds of kilometers, serves as a bounding fault for the Sierra Nevada block on the south. The basement terrane across this fault is totally unrelated to the Sierra Nevada block. The complex problems of San Andreas fault offset are not treated further in this report. The Garlock fault, however, has undergone less movement, and basement-rock units on both sides of this fault in the study area bear directly on the problems of its displacement.

GARLOCK FAULT DISPLACEMENT

The Garlock fault marks a structural break between the Sierra Nevada block and the Mojave block. This fault has long been a feature of controversy and discussion, and much has been written about the fault and the problem that its large offset poses. Davis and Burchfiel (1973, p. 1407) summarized much of these data on offset and regional relations and proposed the following interpretation: "* * * the Garlock fault is an intracontinental transform structure which separates a northern crustal block distended by late Cenozoic basin and range faulting from a southern, Mojave block much less affected by dilational tectonics." This model implies left-lateral slip on the Garlock fault, in which the north side of the fault (Sierra block) is more active and the southern side (Mojave block) more passive. Matching of several basement-rock units and structures, such as dike swarms, metasedimentary rocks, and a thrust fault, across the Garlock fault east of the study area, led to an estimate of left-lateral offset of about 60 km (Smith, 1962; Smith and Ketner, 1970; Davis and Burchfiel, 1973). Within the southernmost Sierra Nevada, Michael (1966) suggested that north of the Garlock fault and west of the Walker Basin area (pl. 1) are abundant northwest-trending faults, in contrast with far fewer faults to the northeast. He correlated this supposed change in fault pattern with a similar change in fault pattern that Hewett (1954) noted on the south side of the Garlock fault in the Mojave Desert. These two lines separating areas with different densities of northwest-trending faults appear to be offset by about 65 to 70 km in a left-lateral sense, but the validity of the "line" on the Sierran side is open to question.

Within the area of plate 1, offset of certain basement-rock units appears to provide some evidence regarding the amount of offset of the Garlock fault here near its west end. These basement-rock units are (1) the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains, (2) the Rand Schist in fault horses; and (3) the granodiorites of Lebec and Gato-Montes. The metamorphic rocks of the San Emigdio-Tehachapi Mountains are a distinctive hornblende-rich unit that lies on the north side of the Garlock fault throughout much of the Tehachapi Mountains. The most striking feature of these dark gneissic rocks is the presence of large, in part euhedral, red garnets, as large as 10 cm in diameter, that commonly display bleached haloes against their dark host. These widespread garnets are an index to the dark rocks. Fault slivers of the rock, which is identical with the main mass, are strung along the north side of the Garlock fault from Tehachapi Pass to the area of Cinco near the east edge of the study area. I interpret these slivers to be fragments of the San Emigdio-Tehachapi unit that were left behind as the block north of the Garlock fault moved westward against the relatively stable Mojave block. The distribution of these slivers suggests a minimum of about 50 km of left-lateral slip on the Garlock fault at its west end. Plate 1 suggests that the gneissic slivers east of Tehachapi Pass are on the same side of the Garlock as the main body of largely mafic rocks to the west. These slivers are, in fact, on the same side, if only the line of most recent movement (the Garlock fault) is considered. The Garlock fault zone there, however, is a wide, intertwined zone of metamorphic and granitic fault slivers whose details and north limit are not at present clearly defined. The major movements of the fault zone appear to have been distributed across this broad zone, much as Clark (1973) noted for the large horse composed of the Rand Schist in the Tehachapi Mountains, a horse that is bounded by two distinct fault branches.

Recent reexamination of the mafic basement rocks of the southernmost Sierra Nevada (Ross, 1983b, c) has suggested an alternative explanation for the gneissic slivers from Tehachapi Pass to Cinco. These slivers may be uplifted parts of an extensive mafic basement terrane, rather than dropped-off fragments of the more areally restricted, mafic San Emigdio-Tehachapi terrane. At present, there is no obvious choice between these two alternatives, but if the slivers are uplifted fragments of a widespread mafic basement, this fact would invalidate the fault-offset arguments put forth in the previous paragraph.

An elongate horse of dark schist, 32 km long and as much as 2.5 km thick, between the north and south branches of the Garlock fault is present in the Tehachapi Mountains. At about the longitude of Mojave is a much smaller horse (5 km long and no more than 0.5 km thick) of similar dark schist in the fault zone. Both of these horses are composed chiefly of schist containing various proportions of quartz, muscovite, biotite, and sodic plagioclase. Thin-banded metachert is a common associate, as is quartz, ranging from thin wisps and lenses to thick bull-quartz veins. Also widespread, and particularly abundant in the smaller eastern sliver, is spotted knobby schist. The spots are albite crystals around which actinolitic amphibole, sphene, and epidote are molded. This spotted schist, associated massive amphibolite, and local serpentinite indicate that the source rocks for the two horses had a volcanic component. The rocks of the horses in no way resemble the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains, but they strongly resemble the Rand Schist (Hulin, 1925), a unit that crops out extensively in the Rand Mountains to the east, on the south side of the Garlock fault. The separation from the westernmost exposures of the Rand Schist in the Rand Mountains and to the west end of the largest horse in the Garlock fault zone totals 100 km. The original distribution of the Rand Schist is unknown: For example, it could extend some distance westward under the alluvium of Fremont Valley; if so, the present separation is not necessarily a measure of true fault displacement. The Rand Schist, however, could extend no farther westward than the easternmost schist horse in the Garlock fault zone, because other basement rocks occupy the south side of the Garlock fault to the west of the easternmost horse. Even assuming this maximum possible westward extent of the Rand Schist on the south side of the Garlock fault, a left-lateral separation of the Rand Schist across the fault of about 55 km still remains.

This correlation of the Rand Schist across the Garlock fault does not take into account the newly discovered sliver of Rand(?) Schist in the San Emigdio Mountains. The map (pl. 1) suggests that the Garlock fault cuts the Pastoria fault; however, other interpretations of the relative ages of these two faults have been proposed (see subsection below entitled "Pastoria Thrust Controversy"). If the Garlock fault is, indeed, the younger of the two faults, then the San Emigdio sliver can be excluded from consideration, and the above correlation is still valid. However, if the Garlock fault is older than the Pastoria fault and if the San Emigdio sliver correlates with the schist in the Rand Mountains, then a few more tens of kilometers of offset of the Rand Schist must have occurred on the Garlock-Pastoria fault zone.

Haxel and Dillon (1978) suggested that their informally named Pelona-Orocopia Schist (equivalent to the Pelona, Orocopia, and Rand Schists) may underlie vast areas of the western Mojave region and may be exposed only along major fault zones. If so, then the Rand Mountains cannot be invoked as a unique source for the slivers of the Rand Schist, and correlations based on possible offsets of these slivers from the Rand Mountains are not necessarily valid. For the present, however, outcrops provide the only data we have, and they suggest that the Rand Mountains may have contributed the Rand Schist slivers to the southernmost Sierra Nevada. The resolution of this problem clearly requires subsurface data.

The similarity between the granodiorites of Lebec and Gato-Montes on the north and south sides of the Garlock fault, respectively, is discussed above in the section entitled "Plutonic Rocks." In summary, these two granodiorites are virtually identical in field appearance and in modal and chemical composition. Similar granite masses intrude both terranes. The easternmost outcrops of both of these granodiorite bodies show an apparent left-lateral separation of about 45 km across the Garlock fault.

In conclusion, the best estimates of displacement on the Garlock fault, based on basement-rock relations within the southernmost Sierra Nevada area, indicate left-lateral separation of about 50 km for the western section of the fault. Three distinct basement-rock units (dark amphibolitic gneiss, the Rand Schist, and granodiorite) show comparable offsets, somewhat less than those noted in the central and eastern parts of the fault (Smith, 1962; Smith and Ketner, 1970; Davis and Burchfiel, 1973). To assume that these offsets on the western section of the Garlock fault are valid implies that the Garlock fault merges with the San Andreas fault and, in turn, suggests that westward spreading of the Sierran block is actually aided by drag effects from relative northward movement of the active coastside block southwest of the San Andreas fault. Curvature of the Sierran tail could be accentuated if the Sierra block were being distended by motion on the San Andreas fault. A small component of left-lateral offset just west of the Garlock-San Andreas junction might be expected in this case, but the overall, larger right-lateral movements of the San Andreas fault would cancel it out. Westward distension of the Sierran block could also accentuate the westward bulge in the San Andreas fault at this latitude (as suggested by Davis and Burchfiel, 1973, fig. 5).

The foregoing discussion also has implications for the interpretation of the Big Pine fault, which is conventionally considered to be the westward continuation of the Garlock fault, somewhat offset by the San Andreas fault (Hill and Dibblee, 1953; Davis and Burchfiel, 1973). I suggest that if the basement-rock correlations which I have proposed are valid, the Big Pine fault is not necessarily related to the Garlock fault. The Garlock fault may merge with, and end at, the San Andreas fault, but for a short distance west of their junction the stress systems of the two faults may oppose one another, until the westward dying out of basin-and-range distension returns full dominance to the right-lateral San Andreas system.

The distribution of slivers of the dark, garnet-bearing amphibolitic gneiss unit between Tehachapi Pass and Cinco (pl. 1) suggests that these slivers are not simple vertical-fault slivers, as originally mapped by Dibblee (1959) and Samsell (1962). They not only dip back into the Sierran block, but they are also to some extent interlayered with granitic rocks of the Sierran block. The slivers of Rand Schist, however, seem to be normal slivers along a transcurrent fault; they are elongate along or parallel to the fault, and they have nearly vertical contacts. However, at the west end of the largest horse composed of Rand Schist is a west-facing bulge on the otherwise symmetrically elongate horse of schist.

J.C. Crowell (written commun., 1978) noted a north-dipping fault marked by cataclastic rocks near the west end of the largest horse between the Rand Schist on the south and the mafic metamorphic-rock unit to the north, which he interpreted to be a part of the Vincent-Rand thrust system. Alternatively, I suggest that this fault may be a local feature, resulting from piling up of the mobile schist as it impinged on the resistant gneissic and granitic units here at the west end of the largest horse of the Rand Schist. Thus, local shearing would be expected. A similar west-facing bulge is present between two branches of the Garlock fault near Cinco, but alluvial deposits mask the relations. Clearly, the western section of the Garlock fault still poses many unresolved problems; it is a complex zone between the Sierran and Mojave blocks, as noted by Dibblee (1959) and Clark (1973), that deserves more detailed study, now that basement-rock constraints are fairly well established.

PASTORIA THRUST CONTROVERSY

The Pastoria thrust was first mapped in the Lebec area (Crowell, 1952) and was later extended westward from the Garlock fault to the vicinity of U.S. Interstate Highway 5, a distance of some 13 km (Crowell, 1964). On the east side of the highway, the Pastoria fault zone is marked by several steeply dipping cataclastic zones that lie, at least in part, between the granodiorite of Lebec and the dark amphibolitic gneiss unit to the north. Near the east end of the Pastoria thrust, Crowell (1964) showed a klippe of granitic rocks resting on what he called "diorite" (fig. 47). Dibblee (1973) eliminated the klippe from his map by incorporating it into the main granitic body, and showed his Pelona Schist (equivalent to the Rand Schist of this report) as terminating north of and separated from the Garlock fault. In my rapid reconnaissance traverse across the reported klippe, I found outcrops and float of Rand Schist and dark amphibolitic rocks, but no granitic rocks. The rock-type distribution suggested to me that the Rand Schist extends southwestward to a brecciated zone along the Garlock fault which contains abundant blocks and slivers of the granodiorite of Lebec, marble, and other metasedimentary rocks. It seems geologically, as well as geometrically, simpler that the Rand Schist should pinch out to the west between the converging branches of the Garlock fault, as it evidently does at the east end of the large Rand Schist sliver. I interpreted a small, dark amphibolite outcrop within the Rand Schist sliver to be a fault lozenge in one of the Garlock strands. About 2 km to the west, the contact between mafic metamorphic rocks and the granodiorite of Lebec is marked by a brown sheared zone similar to those along U.S. Interstate Highway 5.

After a considerable period of doubt as to the existence of the Pastoria thrust, I became convinced that it, indeed, represents a fundamental break in the basement rocks of the Sierran tail. Westward from the Garlock fault on a somewhat crinkled but generally east-westward trend to the end of the basement outcrop at the west end of the San Emigdio Mountains, continental metasedimentary rocks of Salt Creek and granodiorite abruptly contact hornblende-rich metamorphic rocks of probable oceanic origin. Sr-isotopic data support this inference because $^{87}Sr/^{86}Sr$ ratios of 0.70337, 0.70364, and 0.70446 have been determined on hornblende-rich rocks north of the contact, and a ratio of 0.70930 has been determined on the granodiorite of Lebec to the south. The granite of Brush Mountain is the only basement-rock unit that appears to interrupt this contact. I have seen no apophyses of the granodiorite of Lebec crossing the above contact. Plate 1 shows the location of this abrupt (structural?) contact, the eastern part of which coincides with the Pastoria thrust of Crowell (1964).

As the Pacific plate moves northward around the big westward bend of the San Andreas fault, relative to the North American plate, it must be exerting tremendous pressure on the Sierra Nevada tail. This pressure has apparently been partly released by northward bending of the western part of the mountain block, but also partly by imbrication within the basement and overlying sedimentary cover. The Pleito thrust (W.R. Cotton, N.T. Hall, and E.A. Hay, written commun., 1978) and thrusts in the subsurface revealed by oil-well drilling (T.H. Nilsen, oral commun., 1976) probably reflect this imbrication. The Pastoria thrust may be another example of this shearing and imbrication, but following a zone of weakness along an older structure. If a structural break occurs between the granodiorite of Lebec and the gneiss, amphibolite, and granulite of San Emigdio-Tehachapi Mountains, it appears to predate both the granite of Brush Mountain and the present San Andreas fault. Across this postulated structural break, the distinct and sharp contrast of basement terranes persists for a strike length of more than 60 km. The break may be a strike-slip fault that has been crinkled by subsequent deformation and, much more recently, reactivated in places by thrust movements. Ovoid masses of metasedimentary rocks, east of Grapevine Canyon along the contact between the granodiorite of Lebec and the dark belt to the north, suggest fault slivers. Such

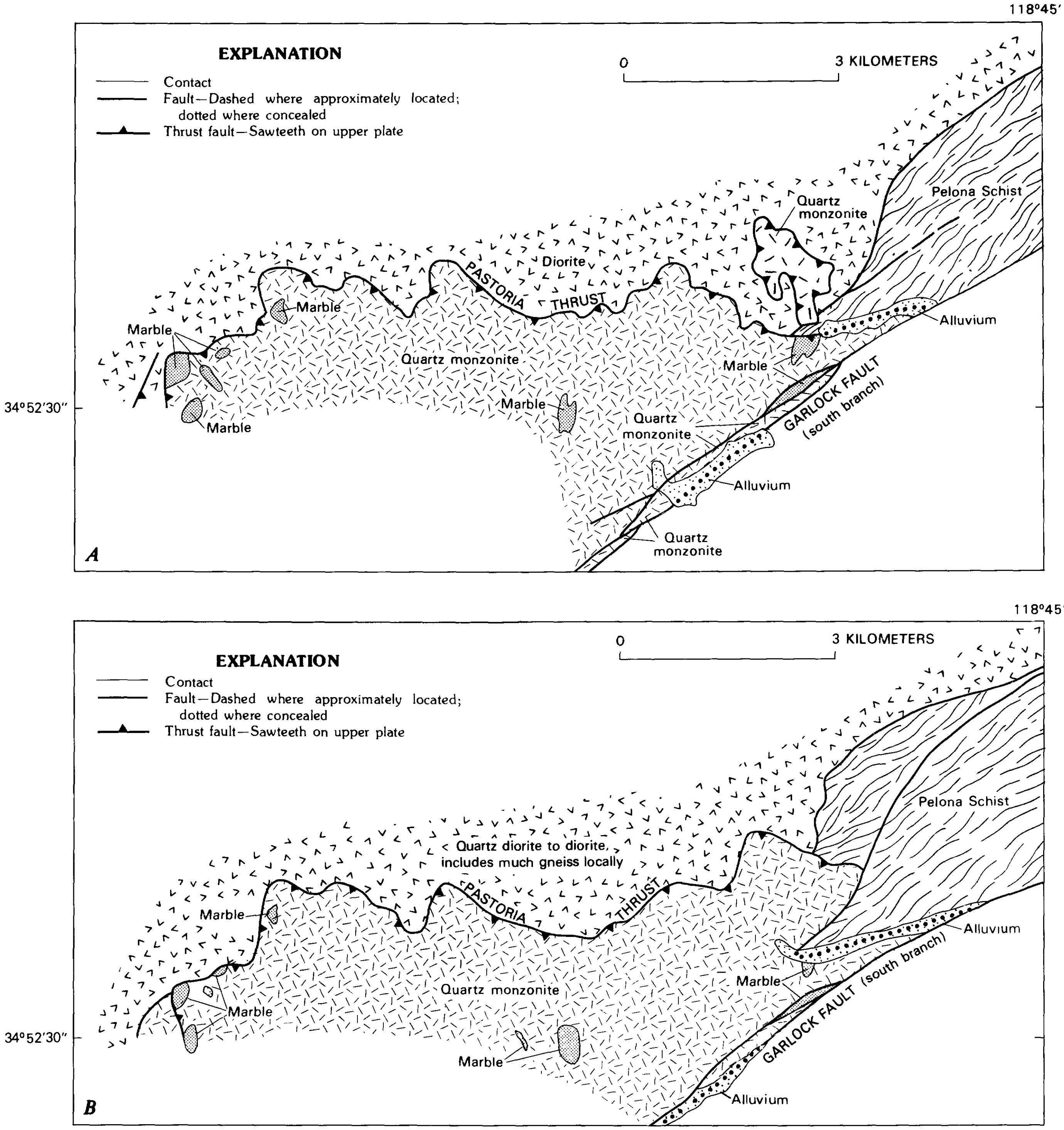


FIGURE 47.—Two earlier interpretations of the geology near the western convergence of the northern and southern branches of the Garlock fault. *A*, From Crowell (1964). *B*, From Dibblee (1973).

elongate slivers, along Crowell's Pastoria thrust, are also what would be expected along a strike-slip fault.

The sliver of Rand(?) Schist in the San Emigdio Mountains, which is lithologically distinct from the associated granitic and gneissic rocks and is of lower metamorphic grade, lends support to the thesis that the Pastoria fault is a transcurrent fault with significant displacement. The schist sliver is abruptly cut off on the south by the Pastoria fault, and the north contact also appears to be a structural contact at the few places I have seen it. West of Grapevine Creek, the contact of the schist with the dark amphibolitic gneiss unit to the north is abrupt and marked by a narrow shear zone that dips steeply north. The schist seems to nose out to the east, a feature suggesting a thrust contact that plunges under the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. On the ridge east of Pleito Creek, both the north and south contacts of the schist sliver are abrupt and marked by zones of sheared rocks, as much as a few tens of meters thick. Most certainly, both of these contacts are faults; the dips are uncertain, but they almost surely are steep.

CATACLASTIC ROCKS[5]

My investigations concentrated on the distribution and characteristics of the metamorphic and plutonic rocks of the study area; study of faults and deformation was, by design, secondary. Nevertheless, as the study progressed, it became difficult to ignore the widespread cataclastic deformation that ranges from granulation and slivering of quartz and a general mortar structure, to strong fluxion structure (mylonite and ultramylonite) (Higgins, 1971, p. 3). Figure 48 shows the locations of samples in which cataclasis was noted in thin section. The absence of notations of cataclasis along the San Andreas and Garlock faults reflects the fact that I generally avoided these two zones because of the difficulty of obtaining fresh basement-rock samples.

Although the coverage of figure 48 is neither systematic nor exhaustive, it probably provides a reasonable sample

[5]I use this term in the sense of Higgins (1971) and include the entire range of deformed rocks, from fault gouge to ultramylonite.

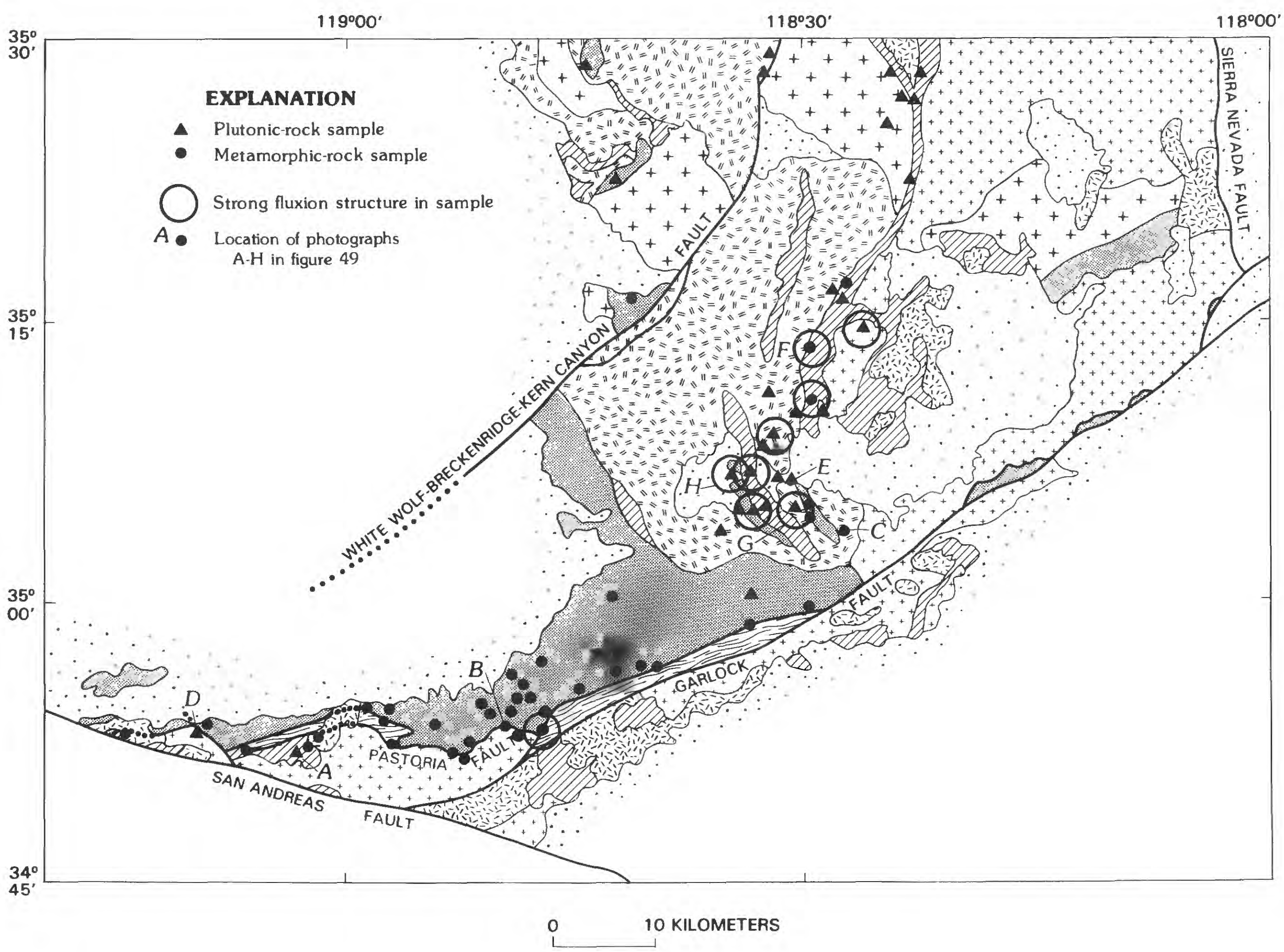


FIGURE 48.—Locations of selected samples that show cataclastic deformation. See figure 37 for explanation of geology.

of the distribution of cataclastic deformation away from the major faults. Of possible interest is the apparent concentration of cataclastic deformation in the east-west-trending tail of the Sierra Nevada, particularly in the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. Another apparent concentration of cataclastic rocks is along a generally north-southward trend close to the quartz diorite line of Moore (1959). Some of the more intensely deformed samples, from the marginal zones of the tonalites of Bear Valley Springs and Mount Adelaide, may reflect protoclastic deformation that accompanied emplacement of these masses. Figure 49 illustrates eight examples of cataclastic deformation that show the range from incipient granulation to intensely deformed ultramylonite.

SEISMICITY

From 1932, when relatively comprehensive instrumentation to record seismic events was begun in southern California (Hileman and others, 1973), to 1979, a total of 21 earthquakes of $M \geqslant 5$ were located within the area of plate 1 (fig. 50). Of these earthquakes, the largest was the Arvin-Tehachapi earthquake of $M = 7.7$, which occurred in July 1952. This earthquake seems to have activated the area seismically because only one earthquake of $M \geqslant 5$ occurred between 1932 and 1952, whereas 17 aftershocks of $M \geqslant 5$ occurred within a month after the 1952 Arvin-Tehachapi earthquake. Two more earthquakes followed early in 1954 that also were related, at least spatially, to the Arvin-Tehachapi earthquake pulse. A plot

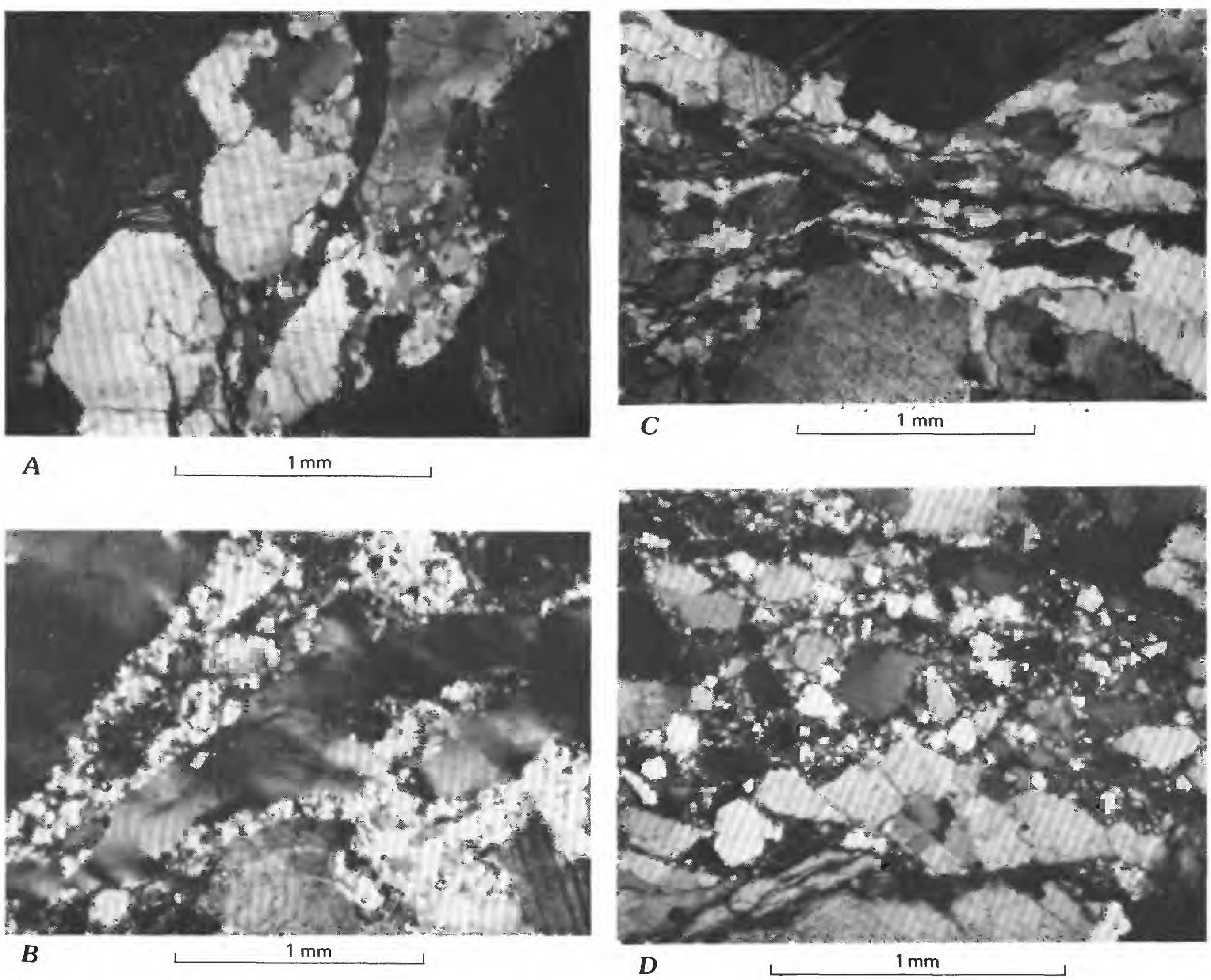
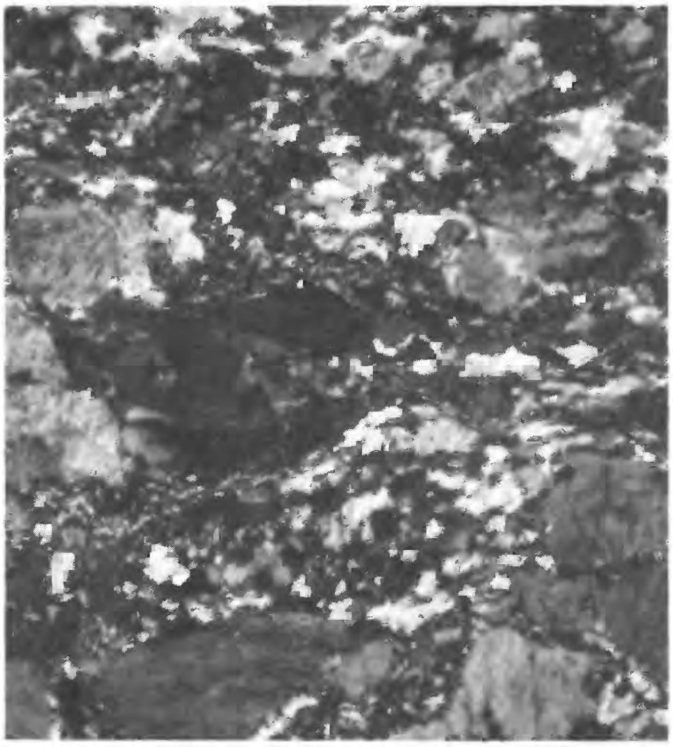


FIGURE 49.—Cataclasis in the southernmost Sierra Nevada—a range of examples shown in photomicrographs (see fig. 48 for locations of samples). *A*, Local microbrecciation of quartz in a felsic granitic rock. *B*, Microbrecciated (partly recrystallized) zones in felsic gneiss. *C*, Slivered quartz squeezed between two "buttress" plagioclase crystals in mildly cataclasized granitic rock. *D*, Extensive microbrecciation, but no fluxion structure,

of earthquakes per year (seismic flux) from 1932 to 1978 (fig. 51) shows the sudden increase in seismic activity in 1952 and the generally higher seismic flux since then. Included on this plot is the Walker Pass swarm of 1946, which comprised six shocks of $M \geqslant 5$. Although the epicenters of the Walker Pass earthquakes lay a short distance north of the study area, this seismic activity may be related to events within this area. Acknowledging that a seismic record of some 50 years is not a valid base from which to discuss recurrence rates and periodicity, I note that the available record indicates either a marked increase in seismic activity in 1952 or an increase in instrumentation and attention to this area after the 1952 earthquake.

The alignment of earthquake epicenters along major faults is not a feature of the seismic record for the past 50 years in the southernmost Sierra Nevada. The nonalignment of epicenters along the San Andreas, Garlock, Sierra Nevada, White Wolf, and Breckenridge faults is clearly evident in figure 50. Instead, the seismicity is relatively concentrated in two rather diffuse areas near



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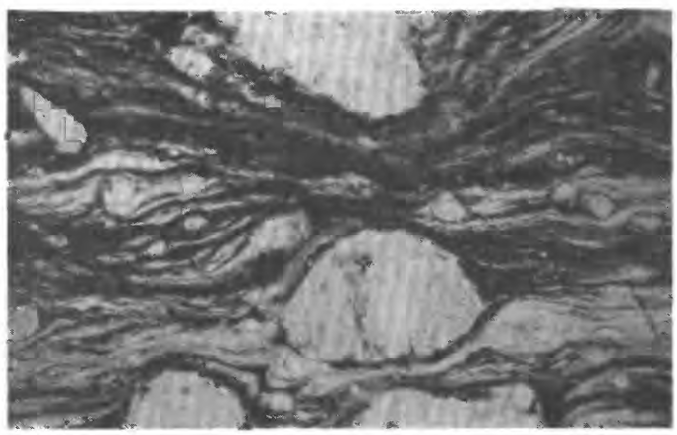

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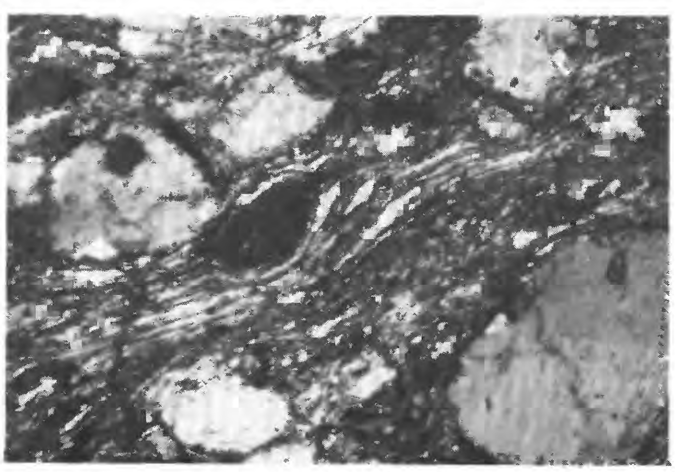

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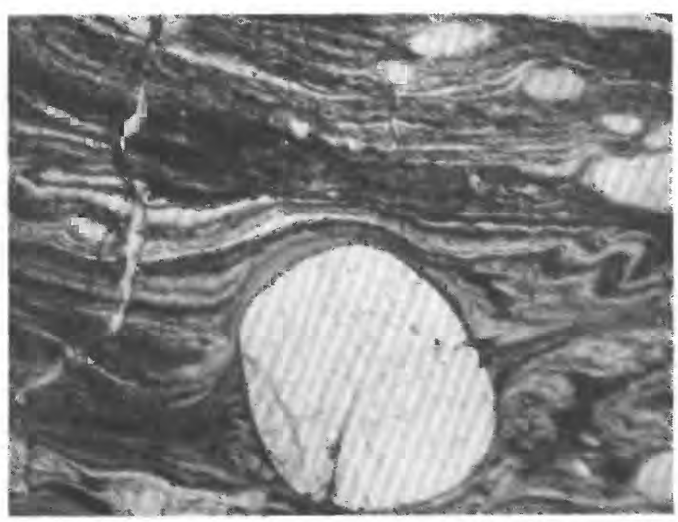

H 1 mm

in a biotite tonalite. *E*, Pervasive microbrecciation, with incipient fluxion structure, in a foliated felsic dike. *F*, Protomylonite to mylonite derived from granitic rock. *G*, Mylonite, probably derived from the tonalite of Bear Valley Springs. *H*, Ultramylonite—black, smeared rock probably derived from tonalite (protolith completely unrecognizable).

the ends of the White Wolf fault. Particularly noteworthy is the virtual absence of any epicenters along almost the entire length of the White Wolf fault. M.G. Bonilla (oral commun., 1980) suggested that the pattern of seismicity in figure 50 may indicate that the initial $M=7.7$ shock relieved nearly all the elastic strain along the White Wolf fault and that the extensive aftershock activity, which has been concentrated in two nodes at the ends of the fault zone, shows that strain is still being released at those points. This same pattern has been noted in the Oroville area of northern California (Lahr and others, 1976, fig. 10).

About 350 epicenters are plotted in figure 50. The relative concentrations of these epicenters were contoured (fig. 52) by means of a technique used in structural petrology (Knopf and Ingerson, 1938, p. 245–251). If the epicenters were randomly distributed across the map, one would expect to find 3.5 epicenters in each 1 percent of the map area. The contours show where the actual concentration is 1.5, 3, and 6 times greater than such a random distribution. Figure 52 confirms quantitatively what was evident from figure 50: The seismicity is not arranged in linear belts that correlate with mapped major faults.

From the area of relatively concentrated epicenters at the southwest terminus of the White Wolf fault, an arc of epicenters can be tentatively projected northeastward on the southeast side of the White Wolf fault to meet the elongate area of concentrated epicenters at the northeast terminus of the White Wolf fault (or near the south end of the Breckenridge fault). This arcuate belt might reflect aftershock activity along a gently dipping thrust plate whose surface expression is the White Wolf fault. Buwalda and St. Amand (1955) noted that the White Wolf fault "* * * appears to be a steep reverse fault or a thrust." Dibblee (1955), however, noted that the "plane of movement * * * dips into the mountains throughout

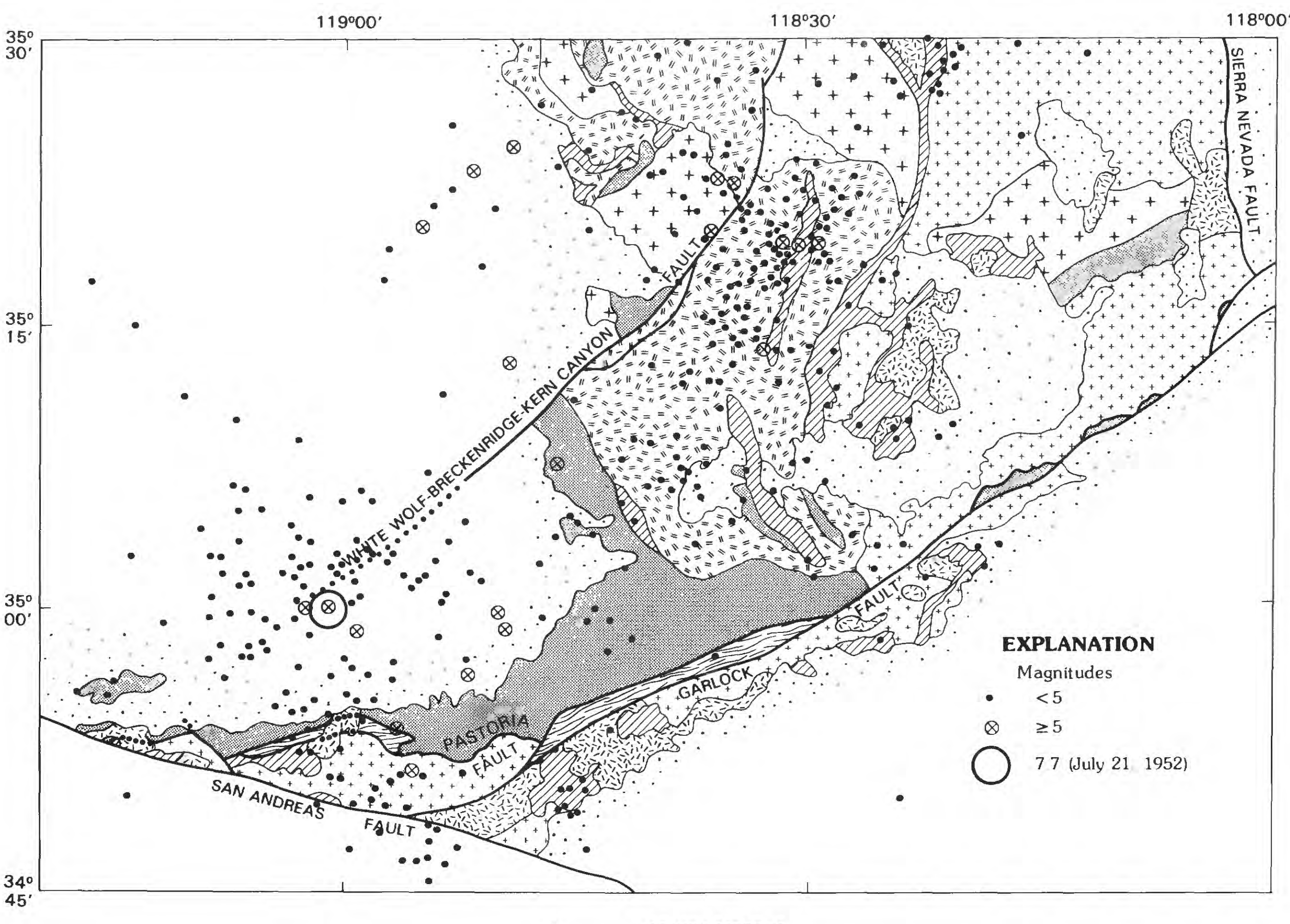

FIGURE 50.—Locations and magnitudes of earthquake epicenters in the southernmost Sierra Nevada, 1932 through 1979. See figure 37 for explanation of geology.

its course at an average angle of about 15°." Cisternas (1963) presented a composite vertical cross section (fig. 53) based on aftershocks of the 1952 event that shows the locations and depths of some epicenters and suggests that the White Wolf fault is dipping about 50° SE. His plot of epicenters, which also includes additional fault planes northwest of the White Wolf fault proper, suggests that the White Wolf fault zone is several kilometers wide.

Shocks that occurred from 1976 to 1978 were located much more precisely, and their focal depths more closely determined, than those of aftershocks of the 1952 Arvin-Tehachapi earthquake. The focal depths of some of these later shocks, plotted on a vertical cross section across the White Wolf fault, suggest a fault plane dipping about 60° SE. (M.G. Bonilla, oral commun., 1980). Thus, the pattern of strain release suggests a steeply dipping reverse fault, rather than a gently dipping thrust surface.

An interesting northeast alignment of epicenters (including three events of $M \geqslant 5$) occurs just east of Bakersfield (fig. 50). Movement on this zone caused the Bakersfield earthquake a month after the Arvin-Tehachapi event. Although no surface geologic feature corresponds to this epicenter trend, its general parallelism with the White Wolf-Breckenridge trend is noteworthy. Cisternas (1963) noted that focal-mechanism determinations for these aftershocks (just east of Bakersfield) indicate right-lateral strike-slip movement, in contrast to the left-lateral strike-slip or dip-slip movement indicated for most of the other White Wolf aftershocks. Figure 53 shows that Cisternas (1963), on the basis of a plot of hypocenters of the aftershocks, considered this zone to dip about 70° NW.—a dip that puts it markedly at odds with the other proposed faults on his cross section.

On a plot of epicenters for earthquakes of $M \geqslant 4$ in southern California for the period 1932 through 1972 by Hileman and others (1973), the two concentrations of epicenters at the ends of the White Wolf trend are readily apparent (fig. 54). In addition, a somewhat smaller concentration is evident on strike(?) to the northeast, which reflects the Walker Pass earthquake swarm of 1946. An

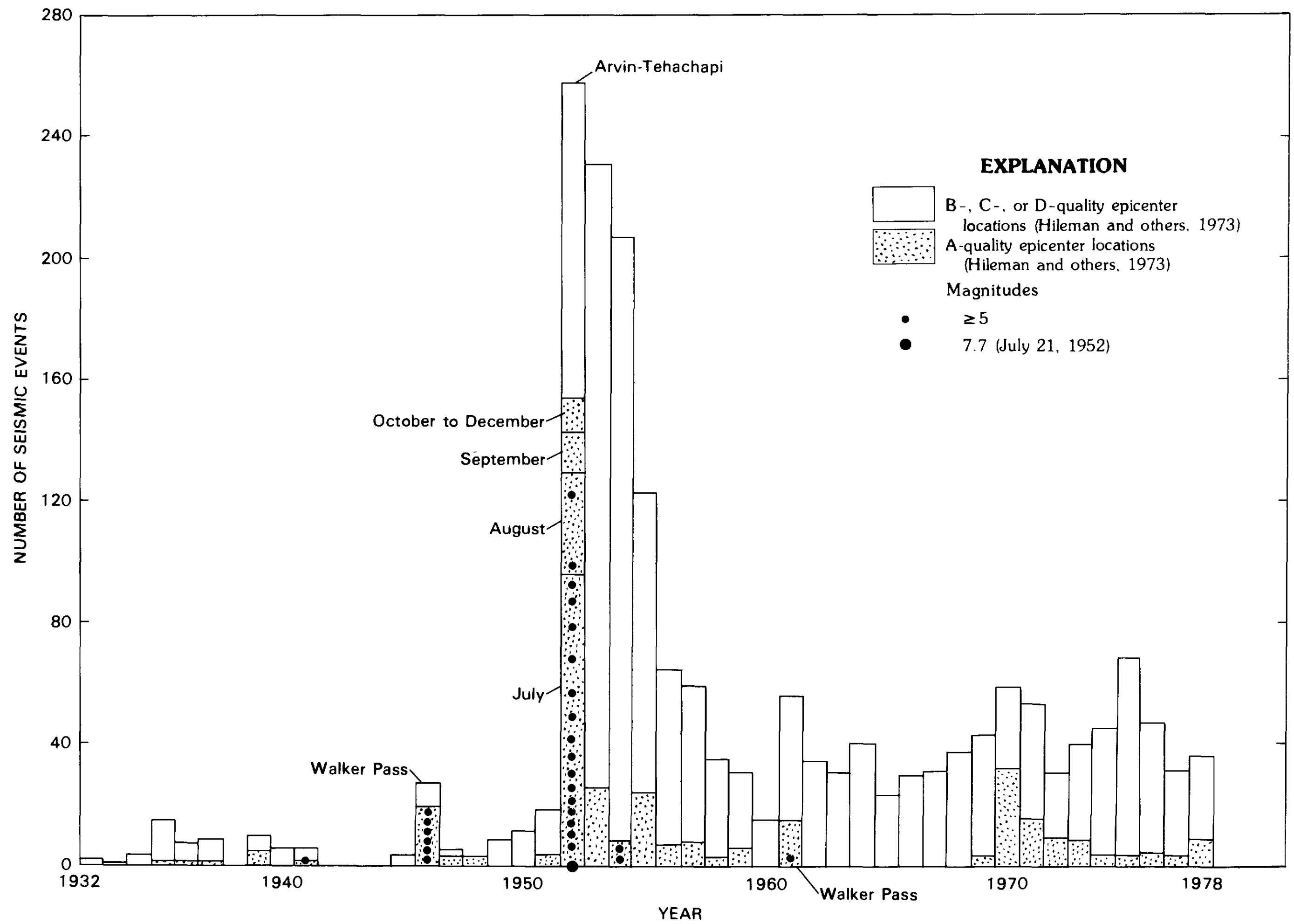

FIGURE 51.—Number of earthquakes per year in the southernmost Sierra Nevada, 1932 through 1978, showing magnitudes of selected earthquakes.

earthquake of $M \geqslant 5$ occurred in the Walker Pass area in 1961. The apparent alignment of these three concentrations may reflect a northeast-trending crustal feature that parallels the Garlock fault zone. The speculated alignment of these three concentrations of seismicity may be more apparent than real, however, because the Walker Pass swarm (approx 40 km northeast of Kelso Valley) may well be connected with the generally north-south-trending Sierra Nevada fault system and, if so, is not related to the White Wolf trend.

Walter and Weaver (1980) noted a possible northeast alignment of epicenters in their comparison of historical seismicity between 1932 and 1977 with the seismicity recorded on a 16-station seismographic network installed in the Coso Range (40 km northeast of Walker Pass) in 1975. They noted a possible left-lateral strike-slip zone projecting westward from the Coso Range through the area of the 1946 Walker Pass earthquakes and along the South Fork of the Kern River to the Kern Canyon fault near Isabella Lake. They further suggested that this hypothetical zone, rather than the parallel Garlock fault, may be the focus for current strain release across the southern Sierra Nevada. In an earlier study of the Walker Pass earthquake swarm, Gardner (1965) considered that it had occurred on a northwest-trending fault in the interior of the Sierra Nevada block, and saw no evidence of a continuous deep structure across the southern Sierra Nevada. He suggested that any relation between the White Wolf and Walker Pass earthquakes must be explained by means of mutual transfer of strain between different fault systems. At present, no known geologic structures trend northeast from the White Wolf fault to the Walker Pass area.

A plot of epicenters from April through September 1979 reported by Allen and Whitcomb (1980) shows a suggestion of a north-south-trending belt of activity from the

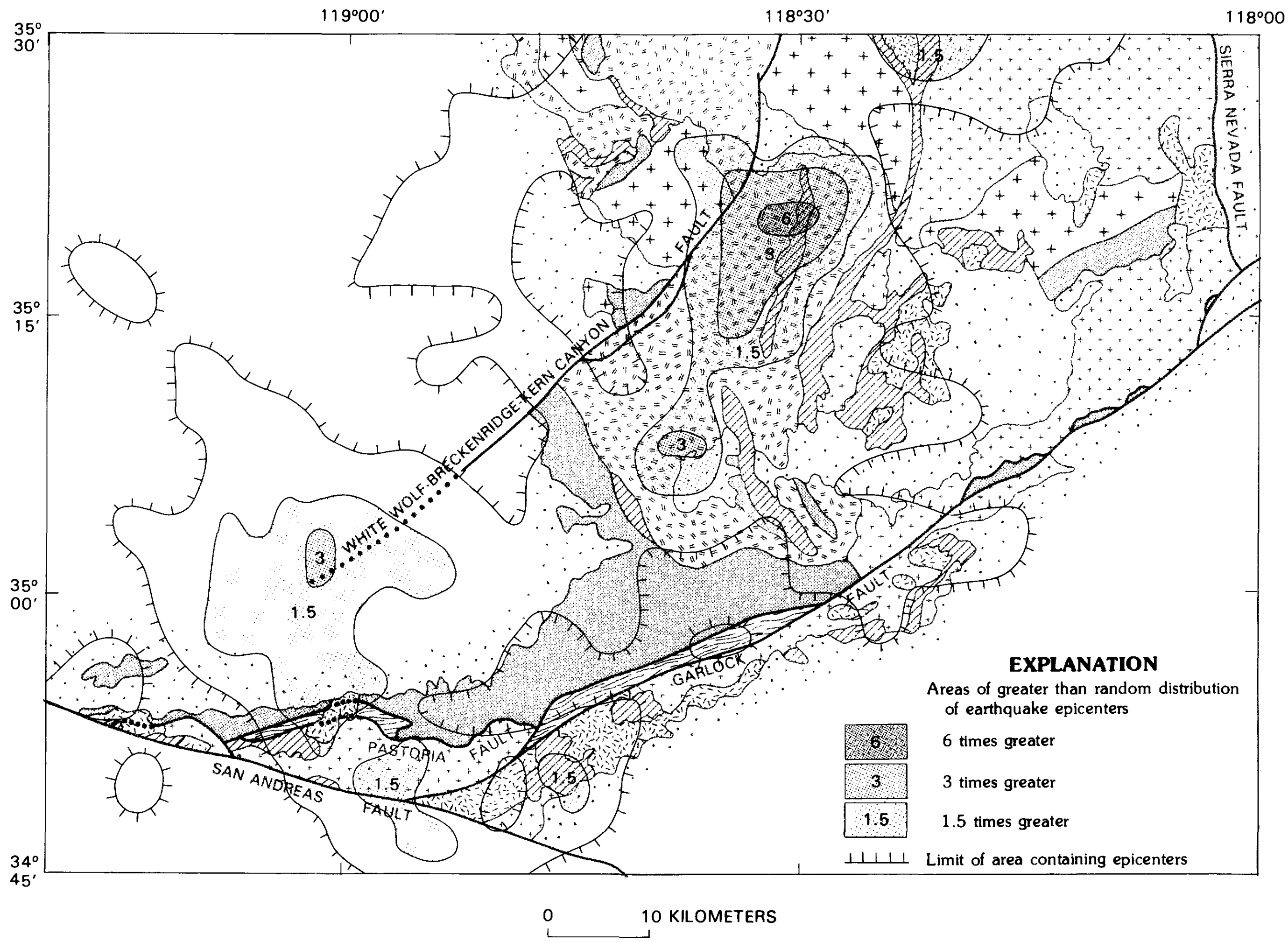

FIGURE 52.—Concentration of seismicity, relative to random distribution, in the southernmost Sierra Nevada, 1932 through 1979. See figure 37 for explanation of geology.

area of the White Wolf fault to Isabella Lake (approx 15 km north of the study area). The noteworthy feature of this plot is that the trend of this belt of activity can be visualized as grossly continuous from the White Wolf fault to the Breckenridge-Kern Canyon trend (fig. 55). Also on this short-term plot, the Walker Pass activity (fig. 54) appears to be separated from the north-south-trending belt and to be spatially related to activity in the Coso area (fig. 55), to the east across the Sierra Nevada fault.

STRUCTURAL IMPLICATIONS OF THE SIERRA NEVADA TAIL

The east-west-trending southern tail of the Sierra Nevada has long been an intriguing structural puzzle. For almost its entire length, the Sierra Nevada block is dominated by a north-southward to northwestward structural grain. Between Cummings Valley and the Garlock fault is an unexplained abrupt change to a dominant east-westerly structural grain. Field data indicate that this change cannot be the result of a simple oroclinal flexure; however, an oroclinal flexure with a strongly appressed and somewhat contorted axial region could satisfy the field relations. The nose of such a flexure is absent, possibly removed by movements along the Garlock fault. However, the proposed flexure raises serious space problems, particularly with regard to the basement rocks of the Mojave block.

On the basis of Sr-isotopic data, Kistler and Peterman (1978, p. 6) suggested that the east-west-trending tail reflects an original alignment of the continental margin. However, some of the basement-rock relations in the study area would present a simpler regional pattern if this tail were unbent.

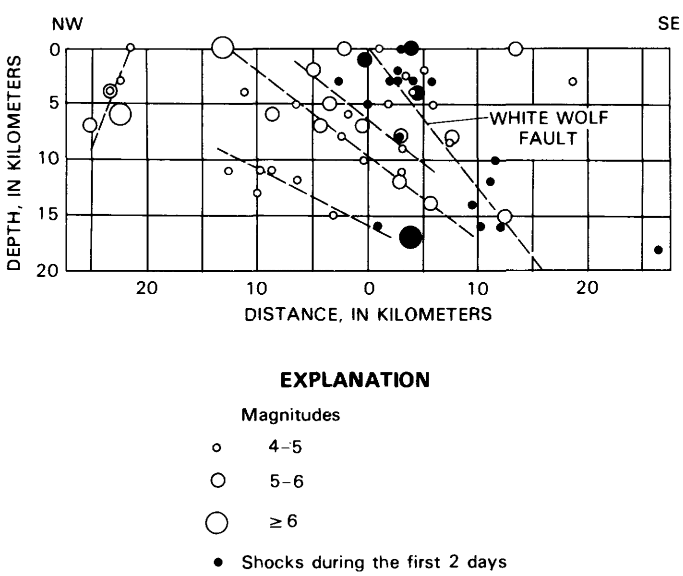

FIGURE 53.—Vertical cross section of the White Wolf fault, showing distribution of selected aftershocks of the 1952 Arvin-Tehachapi earthquake. Dashed lines indicate possible breaks north of the main fault (from Cisternas, 1963, p. 1081).

The quartz diorite line of Moore (1959), which within the study area separates tonalites on the west from felsic granodiorite and granite on the east, trends north-south and is cut off by the Garlock fault (fig. 56). In the tail portion of the Sierran block, the presumed quartz diorite line trends east-west, coincident with a major structural break, the Pastoria fault zone.

If the tail were unbent, not only would the similar granodiorites of Lebec and Claraville east of the quartz diorite line be lined up, but also the metasedimentary rocks of the two granodiorites would then be on strike alignment. The Salt Creek and Keene metasedimentary areas resemble each other; both are rich in calcareous and siliceous rocks.

The mafic metamorphic terrane of the San Emigdio-Tehachapi Mountains was intruded by the tonalite of Bear Valley Springs. The contact zone is a wide zone marked by mixing and interfingering of dark metamorphic rock and tonalite. Throughout the Bear Valley Springs mass are suggestions that the tonalite intruded a terrane occupied by these rocks of oceanic affinity and was much contaminated by them. North-south realignment of the dark amphibolitic rocks by unbending of the Sierran tail would simplify the paleocontinental-margin trend in the southernmost Sierra Nevada. Such realignment would also simplify the map pattern of the mafic metamorphic terrane and the tonalite of Bear Valley Springs—the zone of mixing and interfingering would be on strike with the general northerly direction of elongation of the tonalite mass.

All these basement-rock trends can also be accommodated if the bend in the Sierran tail is considered to be an original feature—in other words, if the present rock distribution reflects an ancient continental margin that originally had a bend. However, because north-southerly regional trends dominate throughout the major part of the Sierra Nevada block and, indeed, throughout the Cordilleran batholithic belt up and down the coast of both Americas, it seems reasonable to postulate that the southernmost Sierra Nevada has been rotated from a more north-southerly original trend.

Kanter and McWilliams (1982) noted a clockwise rotation of the Sierra Nevada of about 45°, on the basis of paleomagnetic measurements from eight sites in the Cretaceous tonalite of Bear Valley Springs about 10 km northwest of Tehachapi. The tonalite localities are in an area where the structural grain, as shown by foliation in the tonalite and elongation of metamorphic bodies, trends essentially north-south. Paleomagnetic measurements on plutonic rocks of about the same age in the central Sierra Nevada by Grommé and Merrill (1965) showed essentially no rotation. The change in the trend of structural grain

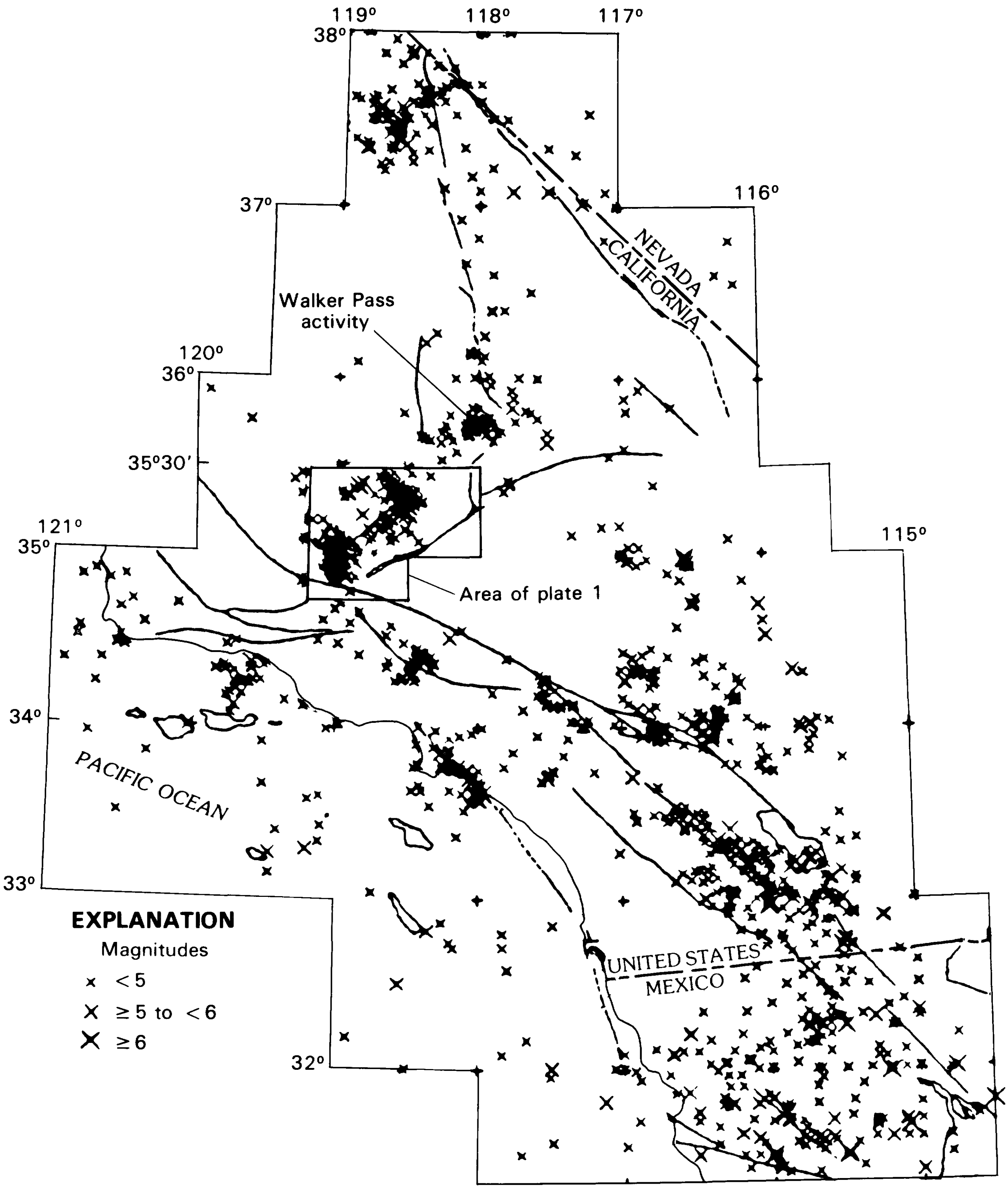

FIGURE 54.—Locations of epicenters of $M \geqslant 4$ earthquakes in southern California, 1932 through 1972. From Hileman and others (1973, p. 64). Heavy lines, faults—dashed where approximately located.

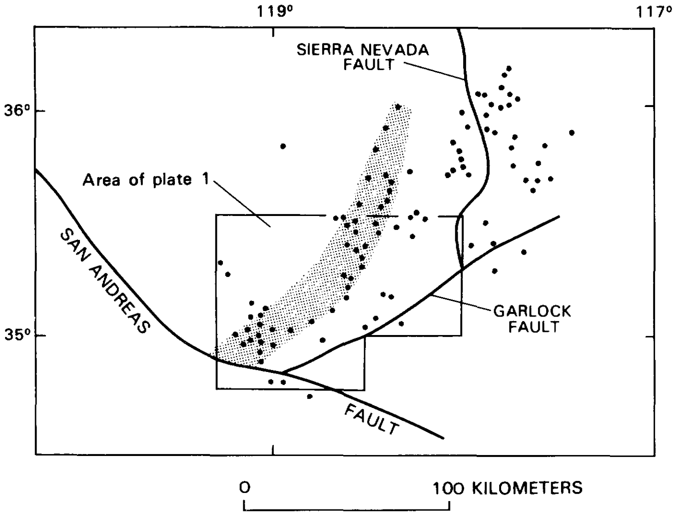

FIGURE 55.—Locations of epicenters of earthquakes from April through September 1979 in the southernmost Sierra Nevada (modified from Allen and Whitcomb, 1980, p. 418). Shading added for emphasis.

from northwestward in the central Sierra Nevada to north-southward in the Tehachapi area thus appears to reflect an actual bending of the batholith. Paleomagmatic measurements by Kanter and McWilliams (1982) on Miocene volcanic and sedimentary rocks near Tehachapi showed essentially no rotation. Therefore, the batholithic bending occurred sometime after emplacement of the Cretaceous plutons and before deposition and emplacement of the Miocene units. However, the absence of paleomagnetic data from the east-west-trending Sierran tail permits no more than speculation that the tail represents further bending.

South of the Garlock fault, all the granitic rocks have compositions that belong east of the quartz diorite line of Moore (1959). If the left-lateral movements on the Garlock fault zone reported by Smith (1962) and Smith and Ketner (1970) are removed, the granitic rocks south of the Garlock fault would nestle against the granodiorite of Lebec. Furthermore, east-west-trending metamorphic rocks of the Bean Canyon Formation would adjoin the

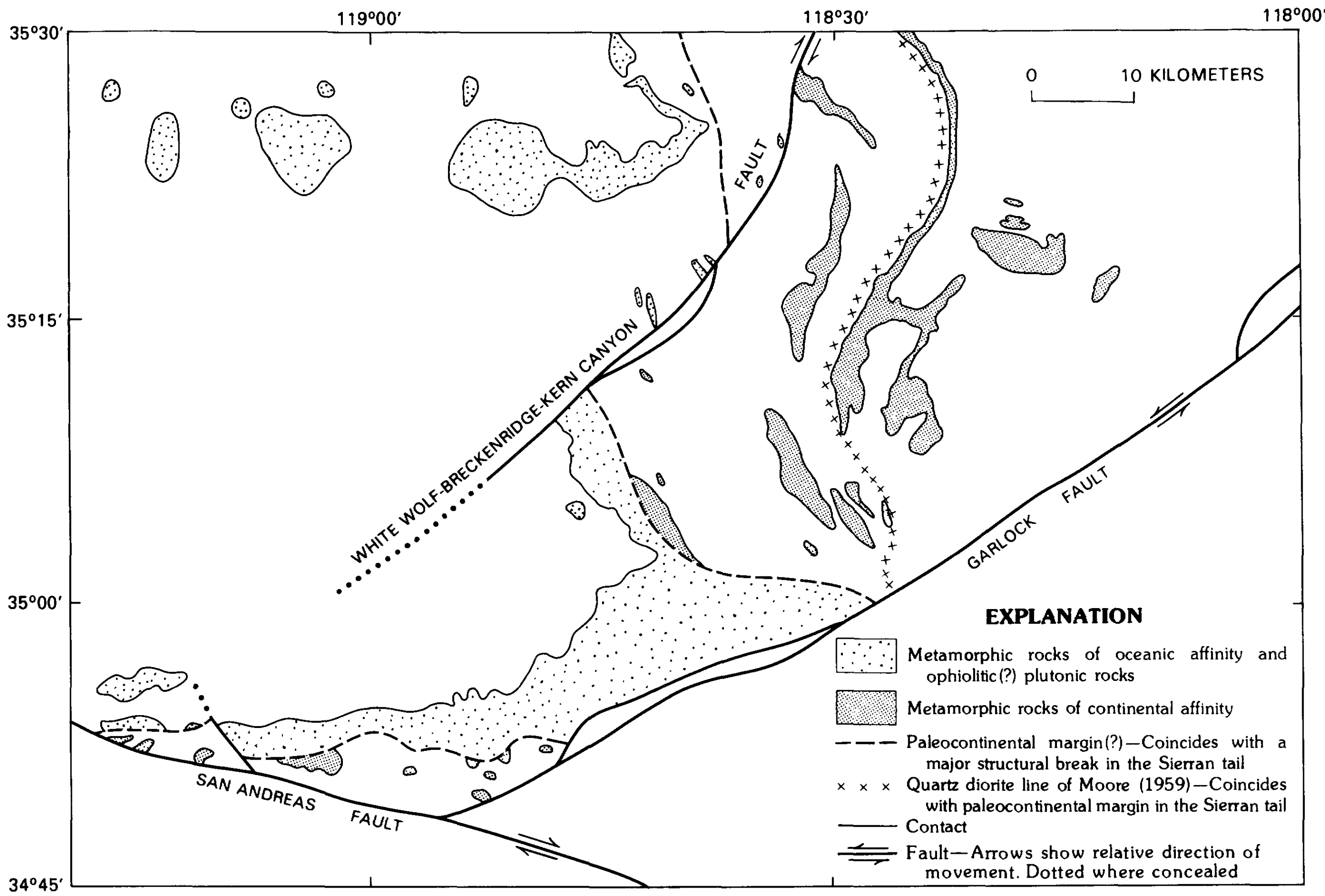

FIGURE 56.—Metamorphic framework rocks of the southernmost Sierra Nevada.

east-west-trending Salt Creek metamorphic rocks. If this reconstructed terrane were then unbent to a northwestward trend, the quartz diorite line would also be straightened out, and the similar Salt Creek and Keene metamorphic rocks would be on strike. The Bean Canyon Formation, however, would then strike north-south toward the eastern Sierra Nevada, where such rocks are not known. Even though unbending would greatly simplify the granitic-rock-distribution pattern, it would create an additional problem with metamorphic rocks of the Bean Canyon Formation. Thus, restoration of the lithologic and structural trends of the Sierran tail to a north-southerly direction raises some problems but notably simplifies the regional basement-rock-distribution pattern and aligns the southernmost Sierra Nevada with the rest of the Sierra Nevada block.

STRUCTURAL ELEMENTS OF THE SOUTHERNMOST SIERRA NEVADA

The southernmost Sierra Nevada (northward to lat 35°30′ N.) can be divided into structural elements based on differences in (1) metamorphic framework rocks, (2) type of crust invaded by intruding plutonic rocks, and (3) structural isolation of blocks of basement by major faults.

The following notes describe the characteristics of each of these structural elements, keyed to the numbers on figure 57.

ELEMENT 1

A major structural distinction in the southernmost Sierra Nevada is between basements of oceanic and continental composition. Thus, element 1 consists of basement that is almost certainly oceanic. The White Wolf fault possibly separates somewhat different oceanic basement elements. Therefore, in the following discussion, the basement is divided into subelement 1A (southeast of the White Wolf fault) and subelement 1B (northwest of the White Wolf fault).

SUBELEMENT 1A

The framework rocks of subelement 1A consist of the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains and the mafic and ultramafic rocks of the Eagle Rest Peak area. Essentially, this subelement comprises the largely mafic metamorphic basement between the White Wolf and Pastoria fault zones and southwest of the tonalite of Bear Valley Springs. Dark, somewhat gneissic diorite to tonalite intrudes the metamorphic framework rocks. Three initial Sr-isotopic ratios of 0.7031, 0.7033, and 0.7038 from gabbroic rocks of the Eagle Rest area and $^{87}Sr/^{86}Sr$ ratios of 0.70337, 0.70364, and 0.70446 from hornblende-rich plutonic rocks of the San Emigdio-Tehachapi terrane point to a mafic oceanic source terrane for the rocks of this structural subelement.

Sparse data from the gabbroic rocks of the Eagle Rest Peak area show a radiometric-age range of 135 to 210 m.y. from K-Ar determinations on hornblende, and an isochron Rb-Sr age of about 200 m.y. from several whole-rock determinations. A Pb-U age on zircon of about 160 m.y. was reported by Thomas Davis (oral commun., 1981) from the Eagle Rest Peak area, but the location and rock type were unavailable. Certainly, more data are needed, but as of now, the age of element 1A framework rocks most probably ranges from 160 to 210 m.y. I am inclined to favor an age of about 160 m.y. from the present data, largely on the basis of the reported zircon age and on the 165-m.y. hornblende age of sample DR–1182C (table 28). The hornblende age of about 210 m.y. is from an unusual hornblende pegmatite and thus may be suspect. Furthermore, the 200-m.y. isochron age probably is not closely constrained because of the limited number of samples. Therefore, the Eagle Rest Peak gabbroic area and, by inference, all of structural subelement 1A, if it is about 160 m.y. old, has a probable age that is comparable to that of several ophiolite occurrences in the Coast Ranges but is markedly younger than most of the Kings-Kaweah belt ophiolite occurrences in the foothills of the Sierra Nevada (Irwin, 1979).

The Eagle Rest Peak gabbroic area may be structurally separate from the rest of subelement 1A (see subsection above entitled "Eagle Rest Peak"). Several Pb-U radiometric-age determinations on zircons from mafic and felsic metamorphic rocks in the San Emigdio and Tehachapi Mountains have yielded ages ranging from 100 to 120 m.y. (Sams and others, 1983). The meaning of these preliminary determinations awaits further results and interpretation.

SUBELEMENT 1B

The framework rocks of subelement 1B consist of the Live Oak, Breckenridge, Pampa, Walker, and Caliente mafic terranes, the Pampa Schist, and a large area of the San Joaquin Valley northwest of the White Wolf fault where oil wells have penetrated greenschist, amphibolite, and gabbro. The framework rocks of subelement 1B have been extensively invaded by tonalite.

The metamorphic rocks of subelement 1B are of somewhat lower grade than those of subelement 1A, but there is also some overlap, and similar amphibolites occur in both areas. The framework rocks of both subelements 1A and 1B have strong oceanic affinities and contrast strongly with the marble and quartzite-bearing continental metamorphic rocks to the east.

The recognition that numerous geologic terranes along the Pacific continental margin of North America (many with oceanic affinities) are exotic and widely traveled naturally brings under suspicion the anomalous structural subelements 1A and 1B of the southernmost Sierra Nevada. Sr-isotopic data indicate that these two subelements are, indeed, underlain, at least in part, by oceanic crust. Present data, however, do not permit deciding whether subelements 1A and 1B are essentially in place and originated alongside the Sierra Nevada block proper, or are exotic blocks accreted to the Sierra Nevada.

Relations between the tonalite of Bear Valley Springs and the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains suggest that about 100 m.y. ago, subelements 1A and 1B were in their present positions relative to the Sierra Nevada block. Figure 57 shows the locations of modal samples of the tonalite of Bear Valley Springs and the hornblende-rich tonalite to diorite that is part of the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains. The considerable overlap of these two units suggests that the tonalite of Bear Valley Springs has invaded and mixed with the dark San Emigdio-Tehachapi rocks. This intrusion and intermixing probably took place about 100 m.y. ago, on the basis of Pb-U zircon ages (Sams and others, 1983) for the tonalite of Bear Valley Springs near Tehachapi. Paleomagnetic determinations on the tonalite of Bear Valley Springs in the same area indicate some rotation, but no translation (Kanter and McWilliams, 1982).

The original boundary between subelements 1A and 1B and the rest of the Sierra Nevada block is now obliterated by the tonalite of Bear Valley Springs. Whether this boundary was a passive continental margin, a collision zone between an accreting oceanic slab and North America, or something else is at present uncertain.

Saleeby (1982) noted a fundamental structural break (Foothills suture) in the central Sierra Nevada that separates continental-margin rocks consisting of Paleozoic and lower Mesozoic quartzite, marble, and volcanic rocks on the east from oceanic or ophiolitic basement on the west. Although the relations in the southernmost Sierra Nevada may be somewhat comparable (fig. 57), direct evidence of a continental margin or a major suture zone is largely erased by a flood of later magmatic rocks.

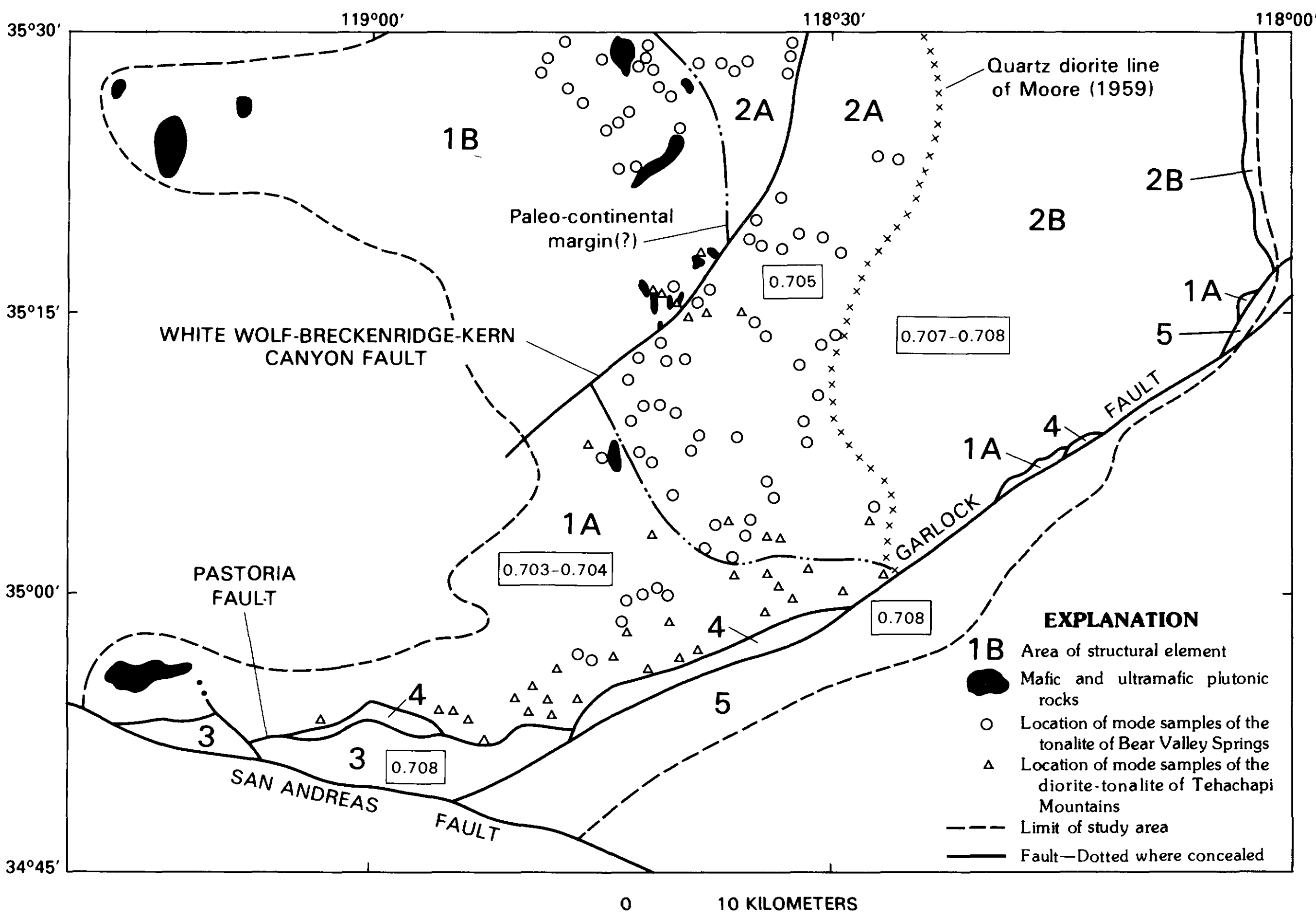

FIGURE 57.—Index to structural elements in the southernmost Sierra Nevada. Generalized initial Sr-isotopic ratios shown for some structural elements.

ELEMENT 2

The subdivision of element 2 into subelements 2A and 2B is to call attention to the quartz diorite line of Moore (1959) and the contrast in initial Sr-isotopic ratio across that line. The metamorphic framework rocks of both subelements are the metasedimentary rocks of the Keene area. This metamorphic package was most likely derived from a sedimentary pile of limestone, sandstone, and shale in various admixtures that was deposited on a near-shore platform or shelf (or in a miogeocline) from a cratonal, continental source. The age of these metasedimentary rocks is not known within the study area, but bivalves of Late Triassic to Early Jurassic age (Saleeby and others, 1978) were recovered from a calc-silicate layer in a section that is lithologically similar to the framework rocks of element 2, just southeast of Isabella Lake, about 15 km north of the study area.

The south end of a conspicuous belt of metavolcanic rocks, some 3 km wide, extends about 5 km into the study area from the north just east of the quartz diorite line. The relation of these rocks to the nearby metasedimentary rocks is not presently known. Although the metavolcanic rocks are generally dark colored, their mineral composition suggests that they were originally intermediate to felsic volcanic rocks. Broken, angular megacrysts in some layers suggest clasts in a tuff or tuffaceous sediment. Interlayers of marble and chert(?) also point to a water-laid tuff origin for at least part of the section. North of the study area in this same metavolcanic belt, near Isabella Lake, I mapped rhyolitic layers with striking fluxion textures that suggest ash-flow tuff.

Subelement 2A

The area between the quartz diorite line of Moore (1959) on the east and the paleocontinental margin on the west is marked by tonalite intruding continental-framework rocks. Two initial Sr-isotopic ratios 0.7055 and 0.7050 from specimens of the tonalite of Bear Valley Springs suggest transitional crust west of true continental crust, but on which largely continental sediment was deposited.

Subelement 2B

The area east of the quartz diorite line of Moore (1959) is truly continental crust of cratonal North America, on the basis of three initial Sr-isotopic ratios of 0.7070, 0.7075, and 0.7080. There, the continental-framework metamorphic rocks are intruded by granodiorite and granite plutons.

ELEMENT 3

South of the Pastoria fault zone in the Sierran tail, the metamorphic framework rocks are the metasedimentary rocks of Salt Creek. This metamorphic section is virtually identical to the framework rocks of element 2. Intrusive into element 3 is the granodiorite of Lebec, a peppery biotite granodiorite with very distinctive, scattered hornblende crystals with red-stained cores that contain skeletal clinopyroxene crystals. An Rb-Sr age of about 85 m.y. on a sample that has an initial Sr-isotopic ratio of 0.7078 shows that it has truly continental affinities. A similar and possibly correlative granodiorite (granodiorite of Gato-Montes—see element 5) is offset on the south side of the Garlock fault.

ELEMENT 4

Element 4 consists of three fault horses along the Garlock and Pastoria fault zones that are petrographically similar to the Rand Schist in the Rand Mountains of the western Mojave Desert. Although these three areas in aggregate are small relative to the other elements, they are tectonically significant because they are so obviously out of place and because they impinge on the Pelona-Orocopia Schist controversy. The overall lithology of the three fault horses suggests a protolith rich in volcanic material and chert and less common argillaceous material that may have been at least partly tuffaceous. Preliminary study suggests there may be a higher percentage of originally sedimentary material in the westernmost horse.

ELEMENT 5

South of the Garlock fault, the metamorphic framework rocks (Bean Canyon Formation) have considerable similarity to those of element 3 but, in addition, contain metavolcanic layers and dark hornfels that includes coarse-grained andalusite—both rock types unknown in element 3. Furthermore, a scrap of ultramafic rock in Bean Canyon may be part of a dismembered ophiolite (G.A. Davis, written commun., 1977). The rock types of the Bean Canyon Formation led Dunne and others (1975) to suggest that it is an arc-trench gap deposit, most likely of early Mesozoic age, although no fossil data are available. Rb/Sr dating of the dacitic metavolcanic rocks in Bean Canyon gave a tentative age of about 150 m.y. (R.A. Fleck, written commun., 1976), which probably reflects the age of the volcanism.

Intrusive into the metamorphic framework are the granodiorite of Gato-Montes and other, more felsic granitic rocks. $^{87}Sr/^{86}Sr$ ratios ranging from 0.7083 to 0.7099 have been determined on the granodiorite. These values, which equate to initial ratios of about 0.708, correspond to the initial ratio determined for the granodiorite of Lebec in element 3 and thus suggest similar continental (cratonal)-crustal environments for elements 3 and 5. However, a strontium anomaly exists in element 5. A probable inclusion of dark granodiorite in a felsic granitic rock near Bean Canyon has an initial Sr-isotopic ratio of

0.7058 (sample SR–12–73, table 28), which differs markedly from those of samples of the granodiorite of Gato-Montes and suggests transitional crust like that which presumably underlies subelement 2A. Although the granitic rocks of element 5 appear to be compatible with those of element 3, some differences exist in in the metamorphic framework and, possibly, also in the underlying crust. The area of element 5 shown in figure 57 is not meant to imply a boundary on the southeast, merely the limit of my study. The isolated and dominantly felsic granitic islands in the largely alluviated western Mojave Desert are probably part of this structural element. For example, an initial Sr-isotopic ratio of 0.708 was determined (Kistler and Peterman, 1978) from a granitic sample about 35 km east of sample SR–13–73.

CRUSTAL LEVEL OF THE BASEMENT ROCKS OF THE SIERRA NEVADA

In a recent study, Sharry (1981, 1982) reported that geobarometry and geothermometry of two-pyroxene granulite north of the Garlock fault suggest pressure-temperature conditions of 0.83 GPa and 740 °C, which equates to a crustal depth of about 30 km. Ghent and others (1977) suggested that migmatitic amphibolitic granulites in British Columbia may represent crustal depths of as much as 25 km; some of these rocks are comparable to the mafic terranes of the southernmost Sierra Nevada. Hamilton (1981), in describing migmatitic middle-crustal rocks in which metamorphic rocks dominate, noted that they represent a depth range from about 12 to 25 km. The shallower parts are in upper amphibolitic facies (and commonly include granitic rocks with primary muscovite), and the deeper parts are in lower granulite facies, characterized by orthopyroxene and brown hornblende. These two environments can be matched in the southernmost Sierra Nevada.

These scattered data suggest that maximum crustal depths of about 25 to 30 km may be represented by the mafic terranes of the southernmost Sierra Nevada. A.H. Lachenbruch (in Bateman and Eaton, 1967, p. 1416) suggested that at these depths melting temperatures (630 °C) could develop. At least local melting and mobilization of the mafic basement rocks in the southernmost Sierra Nevada is suggested by diking and local homogeneous texture, as well as by the presence, in rocks with metamorphic textures, of cleanly twinned and somewhat zoned plagioclase. The common quandary of deciding in the field, or even in thin section, whether a sample is metamorphic or magmatic also suggests at least near-melting conditions. Closely similar U-Pb radiometric ages of associated metamorphic and granitic rocks in the Sierran tail (Sams and others, 1983) also point to depths suitable for melting and equilibration of the metamorphic rocks.

In contrast, Bateman and Eaton (1967) suggested that present exposures in the central Sierra Nevada represent much shallower crustal depths of about 10 km. Other estimates of the crustal depth of present exposures elsewhere in the Sierra generally range from 5 to 15 km (Evernden and Kistler, 1960; Putnam and Alfors, 1965; Hietanen, 1973; Presnall and Bateman, 1973).

Hypersthene- and augite-bearing tonalite was described along the west margin of the central Sierra Nevada batholith by Durrell (1940) and Macdonald (1941). These localities suggest possible analogies with the relatively high grade and deep crustal conditions of the southernmost Sierra Nevada, as does the report of brown hornblende and hypersthene in hornblende gabbro by Durrell (1940). As already noted (see subsection above entitled "Hypersthene"), neither Durrell (1940) nor Macdonald (1941) reported granulite-grade metamorphic rocks in association with the pyroxene tonalites. Saleeby (1977), who examined the areas of Durrell (1940) in considerable detail, considered the pyroxene tonalite to be part of a suture-filling suite of rocks emplaced directly into a major suture zone between oceanic and continental crust. Saleeby (1977), however, made no comment about the crustal depth that these suture-filling rocks represent, but they may well be deeper than the more nearly normal granitic rocks of most of the Sierra Nevada. In any event, it might be worthwhile to examine the mafic inclusions in these pyroxene tonalites, as well as any associated amphibolites and hornblende gabbro, to determine whether granulite-grade metamorphic rocks are absent or may have been overlooked.

Near the east margin of the Sierra Nevada batholith in the Inyo Mountains, I noted occurrences of accessory orthopyroxene and abundant clinopyroxene in dark granitic rocks and coarse-grained prismatic sillimanite in associated metasedimentary rocks (Ross, 1969). Now, in the light of the southernmost Sierra Nevada data, these anomalous Inyo mineral occurrences take on possible significance as crustal-depth indicators.

Clinopyroxene is particularly abundant in these dark granitic rocks (as much as 20 percent, partly in subhedral to euhedral crystals), but coexisting orthopyroxene is relatively sparse. Coarse-grained, prismatic sillimanite from three localities along the east base of the Inyo Mountains was chemically analyzed by Dodge (1971). Coarse-grained, prismatic sillimanite is virtually unknown in the metamorphic terranes of the central and northern Sierra Nevada, or elsewhere in the Inyo-White Mountains.

The dark granitic rocks of the Inyo Mountains are Jurassic, in contrast to the largely mid-Cretaceous age of the southern Sierra Nevada granitic rocks. Dunne and others (1978) considered the dark Inyo rocks to be part of an alkalic suite that is chemically and petrographically distinct from the calc-alkaline granitic rocks which make up most of the Sierra Nevada batholith. The dark Inyo rocks are, indeed, anomalously high in K-feldspar (for such dark rocks) and low in quartz in comparison with typical Sierran granitic rocks. The abundance of clinopyroxene is also atypical of Sierran granitic rocks.

Even if the orthopyroxene-clinopyroxene-bearing granitic rocks of the Inyo Mountains are from a different magmatic suite (and they certainly have a different age) from the granitic rocks of the southernmost Sierra Nevada, they suggest a relatively deep crustal level, possibly comparable to that of much of the southernmost Sierra Nevada terrane.

The granodiorite of Lebec, the major granitic pluton south of the Pastoria fault in the Sierra Nevada tail, has the physical appearance of a shallow, upper-crustal, normal granitic rock similar to the extensive granite and granodiorite plutons to the north and east of the tail. The granodiorite of Lebec, however, contains abundant reddish-brown biotite (fig. 38) and widespread coarse-grained primary(?) muscovite (fig. 39), and coarse-grained sillimanite is more abundant in associated metamorphic rocks (fig. 40) than in metamorphic rocks associated with granodiorite and related granite plutons to the north and east. These mineral associations suggest that the Lebec mass represents a somewhat deeper crustal level than the other granitic plutons and that the whole Sierra tail, not just the mafic terrane north of the Pastoria fault zone, is relatively deep. The crustal-depth differential across the Pastoria fault may be less than that across the Garlock fault to the east, where Sharry (1982) estimated a 30-km crustal depth north of the fault but only a 10-km depth south of the fault. The presence of andalusite and the absence of sillimanite in the metamorphic framework rocks south of the Garlock fault (fig. 40) is compatible with a relatively shallow crustal depth south of that fault.

About 10 km west of Tehachapi, the tonalite of Bear Valley Springs includes a northwest-trending pendant, about 15 km long, that is composed of calcareous, siliceous, and pelitic metasedimentary rocks (pl. 1). The pendant rocks and the enclosing tonalite are grossly similar to other areas of exposure of these widespread rock types. The contact between the metamorphic and tonalitic rocks, however, is a strongly foliated and deformed envelope of tonalite, as thick as 2 km. In part, the tonalite is intensely deformed to mylonite and even ultramylonite. The belt of deformed tonalite also contains felsic granitic rocks, gabbro, diorite (amphibolite?), gneiss, and hybrid-looking rocks. Most contacts between the plutonic and metasedimentary rocks in the southernmost Sierra Nevada are relatively sharp and do not show such pronounced zones of shearing. Two areas of conspicuously sheared plutonic rock are present, however, farther to the north (see subsection above entitled "Felsic Gneiss Bodies Related(?) to the San Emigdio-Tehachapi Terrane"). About 30 km north of Tehachapi, strongly deformed tonalite forms a belt, as much as 1 km thick and several kilometers long, against metasedimentary rocks. About 10 km north-northwest of Tehachapi, a large ovoid mass of felsic granitic rock, included in metasedimentary rock, is strongly deformed to augen gneiss. These two localities and some minor nearby local sheared rocks can most simply be explained as protoclastic deformation at the edge of a pluton. However, this essentially north-south trending zone of deformation, close to the quartz diorite line of Moore (1959), could also mark a zone of considerable difference in crustal depth between the deeper tonalites on the west and the shallower, younger, more felsic granitic bodies on the east.

Just east of the sheared envelope (pl. 1) is the Loop inclusion swarm, containing hypersthene granulite and the Mountain Park gneiss body, and just to the west is the Cummings belt of hypersthene granulite and hypersthene-bearing tonalite. In discussing the Mountain Park body, I have already noted a possible transition between mafic rocks of deeper affinity with metasedimentary rocks of presumably less depth. Is the strongly deformed tonalite envelope evidence of protoclasis, but at considerable depth, on the basis of the presence of both ultramylonite and nearby hypersthene granulite?

I have observed the coexistence of hypersthene and clinopyroxene (in the same thin section) locally in the Pampa and Cummings mafic terranes but have not seen these two minerals together in the San Emigdio-Tehachapi terrane (on the basis of examination of about 150 thin sections). Sharry (1981), in a more detailed study of the eastern part of the San Emigdio-Tehachapi terrane, reported several occurrences of coexisting orthopyroxene and clinopyroxene. Sharry (1981, 1982) used this mineral pair in his geothermometric and geobarometric calculations that suggested crustal depths of formation of about 30 km. If these determinations are valid, such crustal depths are also indicated for the Pampa and Cummings areas (fig. 4), mafic terranes some distance to the north of the main San Emgidio-Tehachapi terrane.

APLITE-ALASKITE-PEGMATITE DIKES AS CRUSTAL-DEPTH INDICATORS

Coarse-grained pegmatite dikes and commonly associated alaskite and aplite dikes are widespread and common in the southernmost Sierra Nevada. Swarms of dikes were impressive enough to be mapped by Dibblee and Chesterman (1953) both east and west of the Caliente mafic terrane, as well as in and near the Pampa body. Conspicuous pegmatite dike swarms were also mapped by Dibblee and Warne (1970) in and near the Loop inclusion swarm and in a large area west of the Comanche mafic rocks.

In my field investigations, I found no additional dike swarms comparable to the five just mentioned. Felsic dike material is notably abundant south of the Garlock fault, locally conspicuous in some felsic gneiss of the San Emigdio-Tehachapi terrane, and widespread and locally impressive throughout the granitic rocks to the north.

In the swarms, pegmatite dikes are as much as 20 m thick, and some can be traced for more than a kilometer. Most, however, are no more than 1 to 2 m thick and are much shorter. Only rarely are the dikes composite. Feldspar and quartz alone make up most dikes, but locally coarse-grained muscovite and (or) biotite and distinctive black tourmaline are present. Some of the dikes in the felsic gneiss contain small amounts of pink garnet.

My impression is that pegmatite and other felsic dikes are much more abundant in the southernmost Sierra Nevada than in areas of the central Sierra Nevada and Inyo-White Mountains that I have studied. However, Bateman and others (1963), Moore (1963), and Bateman (1965) noted locally impressive dikes and swarms of pegmatite and related felsic rocks in the central Sierra Nevada. Hamilton (1981) noted that pegmatite sheets and veins are particularly abundant in middle-crustal granitic terranes (below depths recorded by exposed granitic rocks of the central Sierra Nevada batholith). Pegmatite abundance alone is no compelling argument for deeper exposed crustal levels in the southernmost Sierra Nevada but, in combination with the other indicators previously discussed, supports such a view.

THOUGHTS ON BATHOLITHIC ROOTS

The mafic, largely metamorphic terrane of the Sierran tail, representing possible crustal depths of as much as 30 km, may be an exposure of the root zone of the Sierra Nevada batholith and fragments of the substrate beneath that batholith. In the central Sierra Nevada, Bateman (1979) postulated a layer at a depth of 10 to 20 km in which settled mafic minerals and plagioclase are abundant. Below this level, he envisioned a thick layer of hornblende-bearing garnet pyroxene rocks down to the seismic Moho. The amphibolite- and granulite-grade rocks that are exposed in the southernmost Sierra Nevada have at least some compositional compatibility with Bateman's (1979) suggestions for lower-crustal rocks.

Barnes and others (1981) argued that the compositions of waters from several soda springs in the northern and central Sierra Nevada point to the presence of serpentinite (antigorite), metamorphosed marine sedimentary rock, and marble at depth in areas where the present outcrop is dominantly granitic. They furthermore suggested that the metamorphosed marine clastic rocks may have had a mafic volcanic source. Soda springs sampled in the southernmost Sierra Nevada showed only shallow circulation of meteoric water and provided no hints of the basement composition (Ivan Barnes, written commun., 1980). The "metamorphosed marine clastic rocks of possible mafic volcanic source" that Barnes postulated in the basement of the northern and central Sierra Nevada might be similar to the hornblende-rich mafic rocks of the southernmost Sierra Nevada. Preliminary oxygen-isotope studies on selected samples of gneiss, amphibolite, and granulite from the San Emigdio-Tehachapi terrane show a strong enrichment in ^{18}O (table 1) that points to a possible marine sedimentary protolith for these high-grade metamorphic rocks (Ivan Barnes, written commun., 1983).

The widespread tonalite of Bear Valley Springs has characteristics which suggest that it may also represent a relatively deep crustal level. This tonalite contains widespread and abundant ovoid mafic inclusions, partly of amphibolite and locally of hypersthene granulite. The tonalite locally contains hypersthene, which is fresh and in equilibrium with the rest of the rock. Abundant gneissic streaks and patches, dark schlieren, and some impressive mylonite zones are present in the tonalite. Hornblende and biotite are generally in anhedral, ragged crystals and clumpy intergrowths that give the rock a dirty, messy texture. The tonalite of Bear Valley Springs also appears to grade into, and be mixed with, rocks of the San Emigdio-Tehachapi terrane through a wide, indistinct contact zone. The tonalite has all the traits of a body that has been extensively granitized, ultrametamorphosed, and (or) contaminated by a gneissic terrane—in other words, just the sort of plutonic body that might be envisioned near the bottom of a batholith. In any event, the tonalite of Bear Valley Springs differs markedly from most plutons of the southernmost Sierra Nevada, which look unequivocally clean and intrusive and suggest shallower crustal levels.

Scattered Sr-isotopic data point to an oceanic-crustal environment for the mafic metamorphic terrane of the San Emigdio-Tehachapi Mountains and suggest its relation to the tonalite of Bear Valley Springs. The following initial Sr-isotopic ratios have been determined: mafic terrane, 0.703 to 0.704; tonalite of Bear Valley Springs, 0.7050 to 0.7055; and granite-granodiorite plutons to the northeast, 0.707 to 0.708. Kistler and Peterman (1978) indicated that the line of 0.706 initial Sr-isotopic ratios marks the west limit of Precambrian continental crust. Thus, both the mafic terrane and the tonalite of Bear Valley Springs have oceanic affinities.

Evidence from the southernmost Sierra Nevada suggests that some, possibly most, or even all the mafic ellipsoidal inclusions in the plutonic rocks of the Sierra Nevada batholith are remnants of a mafic root zone or metamorphic substrate. Typical-looking mafic inclusions, but with the mineral content of hypersthene granulite, are found in both the Loop and Comanche areas. A significant number of other mafic inclusions are of amphibolite, composed almost entirely of andesine-labradorite and olive-green to olive-brown hornblende. Other inclusions have, in addition, small amounts of brown to reddish-brown biotite and quartz, but hornblende and plagioclase invariably dominate. These inclusion lithologies closely match those of characteristic rocks of the San Emigdio-Tehachapi mafic terrane and the other widely scattered mafic scraps in the southernmost Sierra Nevada. Although I have seen no inclusion-calving grounds, it takes little imagination to extrapolate from the impressive Loop swarm to the mafic root or substrate of the Sierran tail.

Xenoliths of upper-mantle and lower-crustal affinities have been recovered from trachyandesite that intrudes granodiorite in the central Sierra Nevada (Domenick and others, 1983). Some of the lower-crustal xenoliths appear to be similar to rocks that are presently exposed in the southernmost Sierra Nevada. One xenolith is a garnet granulite, composed principally of clinopyroxene, garnet, plagioclase, orthopyroxene, and minor amounts of quartz, biotite, and hornblende. Domenick and others (1983) inferred that the granulite xenolith represents residua from a source that was partially melted to produce granodiorite. On a diagrammatic crustal cross section, they noted that the "point of origin" of the garnet granulite is at a crustal depth somewhat below 30 km. The garnet granulite, with dominant clinopyroxene, probably is an upper-granulite-grade rock that represents a crustal level somewhat deeper than the exposed lower-granulite-grade rocks of the southernmost Sierra Nevada. I did not observe the mineral pair clinopyroxene-red garnet in the same thin section in the southernmost Sierra Nevada. Other xenoliths described by Domenick and others (1983) are of amphibolite, minor biotite and hornblende hornfels composed principally of plagioclase and amphibole, and lesser orthopyroxene, biotite, and quartz. These xenoliths, shown diagrammatically to have come from depths of about 15 km, appear to be similar to the two most characteristic rock types of the mafic terranes of the southernmost Sierra Nevada. The hornblende hornfels, which has a mineral content much like that of hypersthene granulite of the San Emigdio-Tehachapi terrane, may represent a somewhat deeper crustal level than the postulated 15 km. A sillimanite gneiss xenolith is also described that consists of plagioclase, biotite, garnet, and minor sillimanite and amphibole. Domenick and others (1983) suggested that this xenolith came from a depth of about 10 km. The amphibolite-grade xenolith is compatible with metasedimentary rocks in the broad area north of the San Emigdio-Tehachapi mafic terrane, where sillimanite is the sole aluminosilicate (fig. 40). The absence of quartz in the xenolith is, however, atypical of most southernmost Sierra Nevada sillimanite-bearing rocks.

The fortuitous preservation of these xenoliths, coupled with the diligent observation required to spot them, provides a natural core through the crust and upper mantle, as Brooks and others (1980) noted. Particularly noteworthy is the similarity in mineral content between some xenoliths and the exposed basement of the southernmost Sierra Nevada. The upper part of the "subgranitic crust of metamorphosed sedimentary and igneous rocks" inferred from the xenoliths by Domenick and others (1983), may thus have an exposed counterpart at the south end of the batholith.

A possible analog of the suspected root rocks of the southernmost Sierra Nevada was described from the Coast Range crystalline belt near Prince Rupert in British Columbia, Canada, by Hollister (1975) and Lappin and Hollister (1980). They described granulite-facies mineral assemblages and estimated pressures of 0.5 to 0.9 GPa that suggest crustal depths of 15 to 25 km and temperatures of 750 to 850 °C, high enough for partial melting. They further noted that this is a "practically unique occurrence" of granulite-facies metamorphism of Cretaceous age (66–97 m.y., on the basis of concordant Pb-U ages on zircon). Judging from the general descriptions, the proposed depth-temperature conditions, and the age of metamorphism, these Canadian rocks have much in common with the suspected root rocks of the southernmost Sierra Nevada.

Another batholithic-root zone analog may be preserved in the Santa Lucia Range (see fig. 1). A general comparison of these rocks with the southernmost Sierra Nevada basement follows in the next section of this report. Ross and McCulloch (1979) noted an extensive migmatitic-gneissic mixed zone in the Santa Lucia Range that might represent a batholithic root zone. Although they presented no data on pressure-temperature conditions, the presence of both high-amphibolite- and low-granulite-grade rocks suggests depths comparable to those suggested for similar terranes by Ghent (1977), Hamilton (1981), and Sharry (1982)—depths of about 20 to 30 km. The Santa Lucia Range rocks contain more quartzofeldspathic gneiss and less amphibolitic material than is found in the mafic root terrane of the southernmost Sierra Nevada. Suggestively, the Santa Lucia basement is more continental, and the southernmost Sierra basement is more oceanic. Alternatively, the Santa Lucia basement may be a somewhat more reactive, higher level in a root zone, whereas the southernmost Sierra Nevada basement may represent a somewhat more refractory (and deeper?) part of a root zone.

COMPARISON OF THE HYPERSTHENE-BEARING AND RELATED ROCKS OF THE SOUTHERNMOST SIERRA NEVADA WITH SIMILAR ROCKS IN THE SANTA LUCIA RANGE

The hypersthene-bearing granitic and metamorphic rocks (Compton, 1960, 1966; Ross, 1976a, c, e) of the western Santa Lucia Range (across the San Andreas fault and about 300 km to the northwest near Monterey, Calif.) naturally invite a comparison with the hypersthene-bearing rocks of the southernmost Sierra Nevada. Granitic rocks, of about the same radiometric age as the metamorphic rocks they intrude, have the same, mid-Cretaceous age in both areas. Various zircon fractions from two samples of the hypersthene-bearing tonalite of the western Santa Lucia Range have U-Pb ages ranging from 97 to 100 m.y. (Mattinson, 1978). J.H. Chen (in Hamilton, 1981, p. 282) noted that granulitic metamorphic rocks from the same area gave U-Pb zircon ages of middle Cretaceous (about 100 m.y.). Sams and others (1983) reported a 99-m.y. concordant U-Pb zircon age on the tonalite of Bear Valley Springs (which has hypersthene-bearing parts) and concordant U-Pb zircon ages ranging from 100 to 115 m.y. on samples of the hornblende-rich gneiss terrane in the southernmost Sierra Nevada. These similarities between intrusive ages of the tonalites and possible equilibration ages of the metamorphic rocks for both areas argue in favor of a similar geologic environment for both the western Santa Lucia Range and the southernmost Sierra Nevada.

Hypersthene-bearing tonalite in the Santa Lucia Range contains somewhat less quartz than normal tonalite, markedly more hornblende than biotite, and only traces of K-feldspar (Ross, 1976c, e). Thus, in mineral content and general appearance, the hypersthene-bearing tonalite of the Santa Lucia Range is similar to some hypersthene-bearing rocks of the southernmost Sierra Nevada.

On the basis of fieldwork and petrography in both areas, I am even more impressed by the similarities of some of the high-grade metamorphic rocks of the two areas—particularly quartz-bearing granofels and impure quartzite. These rocks have a clean, fresh granoblastic texture and contain abundant reddish-brown biotite, conspicuous coarse shiny flakes of graphite, and brown hornblende. Red garnet is also characteristic of both terranes. Also noteworthy is the absence of sillimanite, both in the belt of rocks containing granofelsic rocks in the western Santa Lucia Range and in the southernmost Sierra Nevada.

Amphibolite is locally abundant in the Santa Lucia Range, commonly in relatively thin interbeds with marble and calc-hornfels. I interpret much of the amphibolite in the Santa Lucia Range as metasedimentary, in marked contrast to the extensive massive to gneissic amphibolite in the San Emigdio-Tehachapi Mountains, which I suggest is dominantly or exclusively metaigneous. The granofelsic belt in the Santa Lucia Range is strikingly rich in marble relative to the rest of the metamorphic terrane of the Range (Ross, 1976e). This is a complete reversal of the pattern in the southernmost Sierra Nevada, where the high-grade San Emigdio-Tehachapi terrane has only rare and local calcareous material, and the metasedimentary rocks to the north and east are rich in marble.

In earlier work in the Santa Lucia Range, I suggested (Ross, 1976e) that "Most of the metamorphic rocks appear to be high in the amphibolite facies and contain widely distributed sillimanite and red garnets. West of the Palo Colorado-Coast Ridge fault zone, sillimanite is absent, but hypersthene and coarse red garnets are found both in the gneiss and in associated 'charnockitic' plutonic rocks, suggesting that, at least locally, the granulite facies (Compton, 1960, 1966) was reached." On the basis of more recent work in the southernmost Sierra Nevada, I suspect that both Compton and I were conservative in assigning rocks to the granulite facies in the Santa Lucia Range. The almost mutually exclusive occurrence of sillimanite and orthopyroxene (Ross, 1976a, fig. 4) is so similar to the relation in the southernmost Sierra Nevada (fig. 36) that I now suspect that much of basement in the western Santa Lucia Range with limited hypersthene, but abundant granofels, is a granulite-grade terrane. The extensive sillimanite-garnet gneissic and schistose terrane to the east that encompasses the rest of the Santa Lucia Range and the Gabilan Range seems to be largely of amphibolite grade. Compton (1960) noted that granulite-facies rocks "* * * occur here and there in the central part" (of the Santa Lucia Range) within the area of dominantly sillimanite bearing, amphibolite-grade metamorphic rocks. Another notable enclave in the amphibolite-grade terrane is in the central Gabilan Range (Ross, 1976a). There, in a largely marble and mica schist pendant, is a layer of spotted amphibolite that is dominated by polygonal plagioclase (approx An_{45-50}) and light- to moderate-olive-brown hornblende. Present in the rock are conspicuous crystals of hypersthene, pleochroic in shades of pink and green, and less clinopyroxene in fresh crystals, both of which sharply contact adjacent plagioclase and hornblende. Abundant metallic opaque-mineral grains and minor quartz are also present. This rock is most probably a hypersthene-granulite.

Mattinson (1978) noted that initial emplacement of the plutons of the central part of the Salinian block (Santa Lucia Range) was evidently at a considerable depth in the crust (catazone?). Granulite-facies metamorphism and charnockitic tonalite suggest depths of 15 to 20 km and temperatures of 650 to 700 °C, on the basis of mineral assemblages of the metamorphic rocks (Wiebe, 1970). These mineral assemblages in the Santa Lucia Range are compatible with metamorphism above the second sillimanite isograd (Evans and Guidotti, 1966) and, depending on load and fluid pressures, probably developed above temperatures of 650 to 700 °C, at which partial melting could occur. Wiebe (1970) further noted that the widespread occurrence of red garnets and cordierite suggests intermediate load pressures of possibly 0.4 to 0.6 GPa, as estimated by Winkler (1965). This pressure range corresponds to crustal depths of 15 to 20 km. Although these geothermometric and geobarometric calculations indicate a somewhat shallower crustal level than is suggested by similar calculations for the southernmost Sierra Nevada, both terranes reflect significantly deeper crustal levels than the basement terranes of the central and northern Sierra Nevada.

In summary, both the western Santa Lucia Range and the Sierra Nevada tail contain strikingly similar hypersthene granulite, hypersthene-bearing tonalite, and distinctive granofels, and reflect relatively deep crustal emplacement. Both areas are bounded on the east by vast granitic terranes containing sillimanite-bearing, amphibolite-grade metamorphic rocks in which are enclaves of granulite-facies rocks. Thus, the metamorphic environments of the two areas are strikingly similar. Lithologic dissimilarities, such as the form and protolith of amphibolitic rocks and the distribution of calcareous rocks, leave the question of correlation tantalizing, but moot. Correlation would be more acceptable if the metamorphic-grade distribution in the Santa Lucia-Gabilan and southernmost Sierra Nevada terranes were "mirror images" of each other. The northeastward decrease in metamorphic grade in both terranes is a serious obstacle to their original juxtaposition.

The presence of the schist of Sierra de Salinas in the Santa Lucia Range also provides a possible analogy to the southernmost Sierra Nevada. This schist is a possible correlative of the Rand-Pelona-Orocopia Schist terrane (Ross, 1976b). Throughout southern California, these schists occur characteristically in structural contact with high-grade metamorphic rocks. The schist of Sierra de Salinas is largely isolated from the high-grade gneissic and granulitic(?) rocks that are exposed in the main part of the Santa Lucia Range. Migmatitic rocks containing red garnets, however, are in fault contact with the schist at the north end of the Sierra de Salinas, although the relations are unclear, owing to poor exposures and limited study. Granitic rocks intrude the schist of Sierra de Salinas at the south end of the Sierra de Salinas and in the Gabilan Range. Although this relation is atypical of the Rand-Pelona-Orocopia terrane, felsic granitic rocks presumably intrude the westernmost Rand Schist sliver in the southernmost Sierra Nevada.

The near-juxtaposition of a block of possible Rand-Pelona-Orocopia Schist (anomalous to the Salinian block) with hypersthene-bearing tonalitic and metamorphic rocks, both of which contain abundant red garnets, makes it tempting to propose the same structural setting for the Santa Lucia Range as for the other areas of the Rand-Pelona-Orocopia Schist (including the southernmost Sierra Nevada?).

CHEMICAL DATA

CHEMICAL CHARACTERISTICS OF THE PLUTONIC ROCKS

A total of 44 new chemical analyses and semiquantitative emission-spectrographic analyses were obtained for this study. For some samples, instrumental neutron-activation analyses for some elements, including rare-earth elements, were made. In addition, four previously published analyses by Ross (1972) and five by Kistler and Peterman (1978) of samples from the study area are included here. Individual chemical analyses, norms, and emission-spectroscopic and neutron-activation results are listed in tables 29 through 34. The locations of the chemically analyzed specimens are shown on plate 2. The modes of the chemically analyzed samples plotted on figure 58 correlate well with the field shown by the plot of modal averages (fig. 6); therefore, these chemically analyzed samples appear to represent the plutonic terrane fairly well.

Silica-variation diagrams (Harker diagrams; fig. 59) show generally consistent trends for all the oxides, but also a tendency for scatter in the rocks low in SiO_2. From the heterogeneous appearance of these low-SiO_2 rock types (Bear Valley Springs, Tehachapi Mountains, and Antimony Peak units) in hand specimen and thin section, they would be predicted to show chemical scatter. Two samples of the Antimony Peak mass (784B, 3023) are notably anomalous: They are considerably higher in Al_2O_3 and markedly lower in FeO, K_2O, and TiO_2 than the rest of the suite. The other analyzed sample from the Antimony Peak unit (3158) is anomalously high in K_2O and low in Na_2O. These erratic values reinforce the field data, which suggest that the Antimony Peak unit, though probably intrusive, is highly contaminated.

The highest K_2O value in figure 59 (6.9 weight percent) is from a sample (3752) of the granite of Tejon Lookout whose K-feldspar content (about 50 volume percent) is notably high relative to the other samples plotted on figure 59. Other modal samples of the granite of Tejon Lookout that are as high in K-feldspar as sample 3752 would probably have comparably high K_2O values.

Comparatively low CaO and high Na_2O result in a markedly low normative anorthite content for a sample (RWK-11) collected along the Willow Springs-Tehachapi Road, about 2 km south of the Oak Creek Road junction, by Kistler and Peterman (1978). This sample is from a small plug(?) or large inclusion(?) in the Tejon Lookout mass; it contains unusually abundant anhedral pale-green hornblende in clumps and clusters in a gray rock that is noticeably contaminated. Originally, on the basis of the texture and the hornblende aggregates, I had considered this rock type to be a somewhat contaminated inclusion of the Gato-Montes mass in the granite of Tejon Lookout. R.W. Kistler (oral commun., 1979), however, suggested that the initial Sr-isotopic ratio of this sample (0.7058) does not support the idea that it is a contaminated sample of the Gato-Montes mass. At present, the origin of this unusual granitic rock type on the south side of the Garlock fault is uncertain.

The alkali-lime index of all the analyzed plutonic rocks is presented in figure 60. Total alkalis (Na_2O+K_2O) and lime (CaO) are plotted against SiO_2 after the technique of Peacock (1931). The intersection of these two lines (the alkali-lime, or Peacock, index) occurs at about 62.5 weight percent SiO_2, which is within the calcic field (of Peacock). This Peacock index is comparable to that obtained for a suite of western Sierra Nevada rocks, is more calcic than the index for eastern Sierra Nevada rocks, and is less calcic than the index for the southern California batholith (Ross, 1972). Ross (1972) also noted that a plot of granitic rocks from the Salinian block gave a Peacock index of about 61, which lies at the boundary between the calcic and calc-alkalic fields. If the ultramafic and gabbroic rocks of the Salinian block are included with the granitic rocks, the Peacock index shifts to 62.5. The mafic rocks of the southernmost Sierra Nevada were almost certainly undersampled—representative sampling of them would tip the Peacock index to a more calcic value. However, mafic rocks from the southern California batholith with SiO_2 as low as 43 weight percent lie on the same linear trend for both Na_2O+K_2O and CaO as the more silicic granitic rocks (Ross, 1972), and so there is no assurance that more mafic samples would notably alter the Peacock index for the southernmost Sierra Nevada. I am unsure how much significance should be placed on the similarity of the Peacock indices of the plutonic rocks of the southernmost Sierra Nevada and the Salinian block. At the least, the similarity of these indices suggests a similar chemical environment for the two terranes. Also, the calcicity of the southernmost Sierra Nevada plutonic terrane suggests a stronger affinity with western than with eastern Sierra Nevada rocks.

Plots of the relations between FeO and MgO and between CaO and FeO + MgO are shown in figure 61; these plots suppress the overpowering influence of SiO_2 in the standard silica-variation (Harker) diagrams. The plot of total Fe as FeO ($0.9Fe_2O_3+FeO$) against MgO is remarkably linear. Two samples are relatively high in MgO in comparison with their total Fe content. These are the two samples of the Antimony Peak unit (784B, 3023) that were also anomalous in the silica-variation diagrams. In fact, it is their relative depletion in FeO, not their enrichment in MgO, that throws them off the main trend here. The plot of total Fe and MgO against CaO is also remarkably linear. Although one of the five numbered anomalous points on the plot (tonalite of Bear Valley Springs, sample 3837) is not particularly unusual modally, it does contain more than average hornblende and less than average plagioclase. It is richer in FeO, Fe_2O_3, MgO, and CaO than other samples of the Bear Valley Springs mass but is still relatively lime poor in comparison with the other points in figure 59. Sample 3158, from the Antimony Peak unit, is notably low in CaO; this sample is also anomalously low in Na_2O (fig. 59). A sample of the Mount Adelaide unit (3631) is enriched in CaO relative to its $FeO+Fe_2O_3$ and MgO contents, much more so than would be indicated by the silica-variation diagram for CaO (fig. 59). The two samples of the Antimony Peak unit (784B, 3023) that had anomalous Fe/Mg ratios are also anomalously high in CaO, but here, too, it is the low Fe content that pulls these samples off the main trend.

The Alk-F-M plot ($Alk=Na_2O+K_2O$; $F=FeO+2Fe_2O_3+MnO$; $M=MgO$) shows a pronounced trend (fig. 62) that is comparable to those of many other granitic suites in the Western United States (Ross, 1972, p. 83). The trendline for the plutonic rocks of the southernmost Sierra Nevada is bowed up more comparably to trends of calcic suites of the western Sierra Nevada and the southern California batholith than to the trend of the calc-alkalic rocks of the east central Sierra Nevada (Ross, 1972, p. 83).

Normative quartz (Q), orthoclase (Or), and albite + anorthite (Ab + An) were plotted on a triangular diagram (fig. 63) to provide a normative-mineral comparison with the modal quartz-K-feldspar-plagioclase diagram (fig. 58). The trend is linear overall, but somewhat diffuse. One sample (3752) that plots away from the main field is the K_2O-rich sample of the granite of Tejon Lookout. The three normative feldspar components orthoclase (Or), albite (Ab), and anorthite (An) are shown on a triangular diagram (fig. 64). The pronounced trend virtually connects the midpoints on the Or-Ab and Ab-An lines. Again, the K_2O-rich sample from the granite of Tejon Lookout (3752) is markedly off that trend.

A plot of K against Rb contents for selected plutonic rocks (fig. 65*A*) shows a good linear, though somewhat scattered, trend. Only a few samples are markedly off the main trend. The upper point (sample 3752) is the K-feldspar-rich sample of the granite of Tejon Lookout. Sample 3631 of the tonalite of Mount Adelaide is significantly low in both K and Rb contents (table 35) relative to the rest of the Mount Adelaide samples and, in fact, to all the plutonic-rock samples. The thin section of this sample shows a relatively fresh rock with discrete cataclastic zones; the biotite looks unusual and bleached. Some potassium and (or) rubidium could have been selectively removed from this sample. In addition, the samples from the Antimony Peak unit are markedly off the main trend.

The K-Rb relations of the granitic rocks of the central Sierra Nevada and from near the San Andreas fault in the Coast and Transverse Ranges were described by Dodge and others (1970). They noted that the samples from the Sierra Nevada had an overall K/Rb ratio near 230, which is the commonly accepted value for igneous rocks (Shaw, 1968). Their ratio for the rocks near the San Andreas fault was about 260, but only 31 samples were analyzed from this rather extensive terrane. Thus, the general average for the rocks near the San Andreas fault is close to the average of 266 for the rocks of the southernmost Sierra Nevada (table 35) and would be even closer if the highly anomalous values for Antimony Peak were excluded. Dodge and others (1970) noted an enrichment in rubidium in the more potassic (as well as more silicic) rocks of the central Sierra Nevada. Rubidium enrichment is not so pronounced in the southernmost Sierra Nevada, but figure 65B does indicate some enrichment because the trendline that best fits the average values ranges from 258 to 270. However, the field of individual plots (fig. 65*A*), shows a rubidium "hook" at higher potassium values, only if specimen 3752 is ignored. If the anomalous average values of the Antimony Peak and Brush Mountain units were used, the trend would be to even more rubidium enrichment at higher potassium values.

Emission-spectrographic analyses of selected trace elements were performed on all the chemically analyzed samples. In addition, instrumental neutron-activation analyses for rare-earth elements (REE's) and selected other trace elements were performed on some of the chemically analyzed samples. All the trace-element data are summarized in histograms (figs. 66, 67) that show the elemental concentration for each plutonic-rock unit. Figures 66 and 67 provide points of reference for comparing the trace-element abundances in the various plutonic units and should also enable a comparison between the plutonic rocks of the southernmost Sierra Nevada and other plutonic-rock suites.

RARE-EARTH ELEMENTS

Instrumental neutron-activation methods were used to determine the content in 42 of the granitic samples of the following REE's: lanthanum, cerium, neodymium, samarium, europium, gadolinium, terbium, holmium, thulium, ytterbium, and lutetium. The REE-distribution patterns are shown in figures 68 to 76, which plot the contents of these elements, relative to the REE content of average chondrite, in each granitic sample (Haskin and others, 1968). These standardized plots permit easy comparison of the REE data and provide some clue to the amount of fractionation or differentiation that the various plutonic-rock units have undergone. For example, the relatively flat patterns of most samples of the Bear Valley Springs unit, of the sample of one of the felsic masses within the Bear Valley Springs unit, and of the samples from the Tehachapi Mountains and Antimony Peak units (figs. 68, 76) indicate a more primitive, less differentiated rock type than do the steeper, more fractionated patterns of the Claraville and Gato-Montes units (figs. 69, 71).

The most striking REE characteristic of these southernmost Sierra Nevada plutonic rocks may be the widespread presence of a europium (Eu) anomaly. Such an anomaly was to be expected from the felsic granites of Tejon Lookout, Brush Mountain, and Tehachapi Airport, but not from the granodiorite and tonalite bodies, which also have distinct Eu anomalies. Figure 77 summarizes this feature with a plot of Eu/Eu* ratio against SiO_2. Only a few samples in this plot show positive Eu anomalies. One, the sample of the granite of Tejon Lookout (3752), is most difficult to understand because the other two samples of this unit have very strong, negative Eu anomalies, as would be expected in such a felsic body. Sample 3752 is separated from the main Tejon Lookout mass by about 30 km but is probably correlative with it. It is high in modal K-feldspar (50 percent) and consequently high in K_2O content (6.9 weight percent) relative to the other chemically analyzed specimens of the Tejon Lookout unit, but otherwise it appears to be chemically similar to the rest of the Tejon Lookout unit. Other modal samples in the main Tejon Lookout mass also contain as much K-feldspar as sample 3752, and so a high K-feldspar value is not unusual for that unit.

A strong, positive Eu anomaly is shown by one sample of the Mount Adelaide mass (3631). The other Mount Adelaide samples do not have pronounced Eu anomalies, although sample 3631 has a particularly erratic pattern (fig. 70). In general, sample 3631 is modally and chemically compatible with the rest of the Mount Adelaide samples, but modally it is somewhat high in plagioclase and low in quartz, and chemically it is low in K_2O, relative to the other samples.

Most samples of the granodiorites of Claraville and Gato-Montes have modest, but consistent, negative Eu anomalies, although each of those masses provided one sample that does not show a negative Eu anomaly (Claraville, 3737; Gato-Montes, 3356). The anomalous Claraville sample (3737) has no unusual modal or chemical characteristics, but it is separated from the other Claraville samples by deeply weathered, poorly exposed terrane, and it may have been somewhat contaminated by the nearby Bear Valley Springs unit. The anomalous Gato-Montes sample (3356) is typical, both modally and chemically, of the rest of the mass. Except for the Eu deviation, its REE pattern is similar to that of the other Gato-Montes samples.

Six of the Bear Valley Springs samples, as well as the samples of the felsic phase of the Bear Valley Springs unit, and all the samples from the Antimony Peak and Tehachapi Mountains units have relatively flat and primitive patterns that suggest a possible genetic relation between these units. However, three of the Bear Valley Springs samples (3427, 3575, 3670) show a marked bifurcation from the main trend, and these samples are relatively enriched in lanthanum and cerium. F.C.W. Dodge (oral commun., 1980) suggested that the three samples may have been contaminated relative to the other Bear Valley Springs samples. Although reexamination of the three samples revealed no obvious contamination effects and only one (3427) is from near the mafic metamorphic rocks of the San Emigdio-Tehachapi Mountains, the entire Bear Valley Springs mass is characteristically heterogeneous and contains abundant inclusions and schlieren of mafic gneissic material, and so various chemical anomalies would be expected in this unit.

During field investigations, I had considerable trouble in deciding whether the tonalite of Hoffman Canyon is related to the surrounding granodiorite of Claraville or is a separate tonalitic intrusive body related to the Bear Valley Springs mass. The REE pattern in the one sample of the Hoffman Canyon unit suggests a greater similarity to the Claraville unit (fig. 74). This gives some support to considering the Hoffman Canyon unit to be a facies of the Claraville mass.

Another feature of the REE data that is commonly plotted is the La/Yb ratio, which is an index of the flatness or primitiveness of the sampled unit. Figure 78 plots this ratio against SiO_2 content. The trend for most units is approximately linear, with a positive slope from the primitive Tehachapi Mountains and Bear Valley Springs units to the most highly fractionated Claraville, Gato-Montes, and Sorrell Peak units. Somewhat interestingly, the felsic phase of the Bear Valley Springs unit, as well as the Tehachapi Airport, Brush Mountain, and Tejon Lookout masses, plot well off this trend. The anomalous plot for the felsic phase of the Bear Valley Springs unit is not surprising because these rocks are much closer, chemically and modally, to the Bear Valley Springs unit than their higher SiO_2 content would suggest. Nonetheless, I would have predicted that the felsic Tejon Lookout, Brush Mountain, and Tehachapi Airport units would have shown very strongly fractionated REE trends.

Another test of differentiation or fractionation is to compare the level of both the light and heavy end members of the REE span with their abundances in chondrite. Table 36 lists averages and the ranges of values for the sampled plutonic units. The table shows that the felsic granites of Tejon Lookout, Brush Mountain, and Tehachapi Airport are not particularly high in light REE's relative to chondrite but are rich in heavy REE's relative to chondrite, with values of 13 to 20 times chondrite, relative to core granites of the Tuolumne Intrusive Series in the central Sierra Nevada that have values of heavy REE's of less than 3 times chondrite (Frey and others, 1978).

HEAT PRODUCTION AND HEAT-PRODUCING ELEMENTS

According to the calculations of Birch (1954), potassium, uranium, and thorium produce heat by radioactive decay of unstable isotopes in the following amounts: 1 weight percent K produces 0.27 microcalories per gram per year; 1 ppm Th produces 0.20 microcalories per gram per year; and 1 ppm U produces 0.73 microcalories per gram per year. Potassium elemental abundances are available for all the chemically analyzed plutonic samples, but uranium and thorium values are available for only 35 of the 53 samples. These values available are summarized in table 37.

Enough uranium and thorium determinations are available for some of the plutonic-rock units, such as those of Bear Valley Springs, Mount Adelaide, Claraville, Gato-Montes, and Tejon Lookout, to provide some idea of the heat production of the individual units. Other units have been sampled very sparsely, or no data for them are available. Table 37 suggests that heat production in the southernmost Sierra Nevada is comparable to that in other major batholithic terranes of the Western United States.

Tilling and Gottfried (1969), in a study of heat production and heat-producing elements in the Boulder batholith of Montana, also compiled comparative data for other batholithic terranes. They noted that composite heat-production values (in microcalories per gram per year) for the southern California, Sierra Nevada, and Idaho batholiths by rock type were as follows: gabbro and diorite, 0.7; tonalite, 2.7; granodiorite, 5.0, and granite, 7.6. These figures are close to the average values listed in table 37. The major tonalite bodies of the Bear Valley Springs and Mount Adelaide units have an average heat-production value of somewhat less than 3 microcalories per gram per year, the granodiorites of Claraville and Gato-Montes average 5 microcalories per gram per year, and the granite of Tejon Lookout averages 8 microcalories per gram per year.

Tilling and Gottfried (1969) also calculated an overall average value of heat production, weighted according to the areal abundance of rock types in the southern California, Sierra Nevada, and Idaho batholiths, of 4.7 microcalories per gram per year. This value compares closely with the unweighted average of 4.5 microcalories per gram per year listed in table 37 for the rocks of the southernmost Sierra Nevada. Considering the sample density of the plutonic-rock types in the study area, I estimate that a weighted average would be very close to this value (4.5) or possibly slightly lower. With regard to radiogenic heat production, therefore, the plutonic rocks of the southernmost Sierra Nevada appear to be compatible with other Western United States batholithic terranes.

METALLIC MINERALIZATION

In the present study, the mineral deposits were not reexamined, but in the course of the fieldwork it became apparent that metallic mineralization was widespread and included various metals. It therefore seemed worthwhile to compile a list of the productive mines and significant prospects of the area. Almost all of these data have been abstracted from the report by Troxel and Morton (1962). Table 38 lists only those properties that have known production; in addition, figure 79 shows the locations of prospects with metallic mineralization, but no known production. Although these data are incomplete, they include most, if not all, significant occurrences of mineralization. A crude compilation solely from published data indicates metallic-mineral production in the study area of more than $5 million, on the basis of prices at the time of production.

Gold was certainly the most sought after metal and the one that led all others in production in the study area; five properties are reported to have produced more than $600,000 each. Silver production, which was valued at more than $1 million, was limited to one district. Modest production of tungsten from scheelite in contact-metamorphic deposits and of mercury from mineralized Tertiary rhyolite has also been reported. Antimony is widespread, but production has been limited.

Figure 79 shows that most of the mineralization is concentrated between lat 35°00′ and 35°30′ N. The Sierra Nevada tail and the basement terrane south of the Garlock fault contain fewer metallic mines and prospects. Nevertheless, metallic mineralization is locally present, and a small production of antimony and gold has been recorded from the Sierran tail. Also noteworthy is a small amount of tin mineralization south of the Garlock fault—a rare, possibly unique occurrence in California.

Because some paleotectonic models have suggested that the Salinian block is a ripped-off tectonic sliver that was once attached to the present-day southernmost Sierra Nevada, it is worthwhile to summarize what is known of metallic-mineral deposits in the Salinian block and to compare it with that in the southernmost Sierra Nevada. Basement outcrops in the Salinian block consist of abundant intermediate to felsic granitic rocks in contact with calcite marble. In such a setting, contact-metamorphic, metasomatic, and replacement mineral deposits might be expected, particularly of tungsten, molybdenum, copper, lead-zinc, and iron. Nonetheless, evidence of such mineralization is almost entirely absent in the Salinian-block basement rocks. Tactite is rare, and copper staining is virtually nonexistent, as are iron-rich gossans. In addition, quartz veins, the common host for gold and silver mineralization, are also uncommon. Table 39 lists 17 locations where mineralized rocks have been reported; these locations are shown on figure 80. The only metallic-mineral production is placer gold from the east slopes of the La Panza Range (loc. 6), where about $100,000 was produced from 1840 to 1934, but the source of the placers was never located. The only mineral lode deposit, though not metallic, is in the Gabilan Range (loc. 11, fig. 80, near Fremont Peak), where about $31,000 worth of barite was produced from replacement pods during World War I. Even though figure 80 suggests mineralization, particularly in the Gabilan and Santa Lucia Ranges, most of the reports of mineralized rocks listed in table 39 are questionable at best. I am impressed by the rarity of prospect pits and dumps in the basement terrane of the Salinian block; in an area that has been populated since Spanish colonial days, this alone is good evidence of the absence of mineralization because, although exposures are poor, significant mineralization would not likely escape notice. A study by Pearson and others (1967) of the Ventana Cones Wilderness area, which includes a large part of the basement-outcrop area of the Santa Lucia Range, presented many semiquantitative spectrographic analyses of altered rocks and stream sediment; nothing of ore grade was found, and no fluorescent tungsten minerals were seen.

The Salinian block is, therefore, virtually barren of metallic mineralization in the basement rocks. Paleotectonic reconstructions which propose that the Salinian block was ripped away from the south end of the Sierra Nevada are faced with the problem of the profound difference between the extensive mineralization of the southernmost Sierra Nevada and the virtual barrenness of the Salinian block.

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TABLES 29–39; FIGURES 58–80

TABLE 29.—*Chemical data for the tonalite of Bear Valley Springs*

[Chemical analyses in weight percent. Cross, Iddings, Pirrson, and Washington (CIPW) norms in volume percent. Semiquantitative emission-spectrographic analyses and instrumental neutron-activation analyses in parts per million. See table 34 for analysts]

Sample-------	3427B	3575	3670	3791	3837	3858	4181	4184	4281	RWK-13
					Chemical analyses					
SiO_2---------	63.6	62.5	62.1	62.3	57.4	60.3	64.7	60.1	63.5	61.8
Al_2O_3--------	16.2	16.3	16.1	16.2	16.1	16.2	16.8	17.9	17.1	16.8
Fe_2O_3--------	1.6	1.7	1.6	1.6	2.3	1.7	1.5	1.7	1.3	1.2
FeO----------	3.8	4.0	4.2	4.2	5.8	4.6	3.2	4.4	3.5	4.7
MgO----------	2.2	3.1	3.0	2.7	4.5	3.2	2.2	3.3	2.0	2.2
CaO----------	5.4	6.1	6.2	6.0	6.9	6.5	4.8	6.2	4.8	4.6
Na_2O---------	3.8	3.3	3.3	3.8	3.0	3.2	3.7	3.2	3.4	4.1
K_2O----------	1.9	1.9	1.8	1.5	2.4	1.9	1.7	1.5	2.2	1.7
H_2O^+---------	.97	.87	.87	.81	.98	1.0	.53	.67	.68	.79
H_2O^----------	.15	.15	.11	.10	.21	.16	.12	.17	.11	.31
TiO_2---------	.72	.79	.74	.74	.74	.77	.60	.69	.54	.91
P_2O_5---------	.16	.17	.16	.18	.15	.16	.10	.11	.77	.22
MnO----------	.08	.08	.09	.08	.15	.10	.08	.11	.09	.09
CO_2----------	---	---	.01	---	.01	---	.03	.03	.02	.08
Total---	100.58	100.96	100.38	100.21	100.64	99.79	100.06	100.08	100.01	99.50
					CIPW norms					
Q------------	18.74	17.76	17.70	17.05	8.33	17.73	21.94	15.51	22.38	16.54
or-----------	11.29	11.23	10.71	8.93	14.26	11.74	10.11	8.93	13.10	10.21
ab-----------	32.33	27.94	28.12	32.38	25.53	28.32	31.27	27.06	28.85	35.26
an-----------	21.65	24.07	23.97	22.88	23.51	16.78	23.30	30.27	18.93	21.22
wo-----------	1.77	2.13	2.46	2.47	4.12	6.34	---	---	---	---
en-----------	5.51	7.73	7.52	6.77	11.27	8.33	5.51	8.28	5.02	5.57
fs-----------	4.64	4.79	5.37	5.36	7.85	6.04	3.82	5.78	4.67	6.41
mt-----------	2.33	2.47	2.34	2.34	3.35	2.58	2.19	2.48	1.90	1.77
il-----------	1.38	1.50	1.42	1.42	1.41	1.53	1.15	1.32	1.03	1.76
ap-----------	.38	.40	.36	.43	.36	.40	.24	.26	1.84	.53
C------------	---	---	---	---	---	---	.43	.05	2.29	.57
cc-----------	---	---	.02	---	.02	.24	---	---	---	.19
Total---	100.02	100.02	100.01	100.03	100.01	100.03	99.96	99.94	100.01	100.03

TABLE 29.—*Chemical data for the tonalite of Bear Valley Springs*—Continued

Sample-------	3427B	3575	3670	3791	3837	3858	4181	4184	4281	RWK-13
Emission-spectrographic analyses										
B-----------	<4.6	<4.6	<4.6	<4.6	<4.6	8.3	<4.6	27	10	---
Ba----------	790	580	650	550	660	490	540	490	800	---
Be----------	1.9	1.8	1.5	1.3	1.8	1.5	<1.0	<1.0	1.2	---
Co----------	14	13	14	13	18	18	16	26	16	---
Cr----------	17	53	33	33	82	43	40	61	24	---
Cu----------	5.8	19	11	14	27	14	5.1	33	9.2	---
Ga----------	27	25	26	24	25	23	19	21	16	---
Gd----------	10	11	8.4	9.1	13	12	<15	<15	<15	---
La----------	39	32	31	23	22	30	<10	16	16	---
Mn----------	1,400	1,400	1,500	1,400	2,500	1,600	550	1,300	640	---
Nb----------	7.5	5.1	3.8	4.7	5.5	5.4	7.4	5.2	8.5	---
Ni----------	7.0	15	11	17	14	10	15	19	6.4	---
Pb----------	11	12	13	9.1	13	9.5	12	20	20	---
Sc----------	18	15	14	13	28	25	11	26	18	---
Sr----------	450	580	520	460	410	460	460	450	400	---
V-----------	120	120	120	95	150	150	78	130	89	---
Y-----------	32	25	16	16	30	31	7.6	13	14	---
Yb----------	4.6	4.4	3.6	3.2	3.7	4.2	.97	1.7	1.7	---
Zn----------	81	91	82	82	110	96	63	86	64	---
Zr----------	110	39	40	21	100	42	260	110	190	---
Instrumental neutron-activation analyses										
Ba----------	742	581	667	636	645	480	640	434	770	---
Ce----------	56	52	59	29	27	29	21	24	30	---
Co----------	12.6	13.9	15.4	14.3	20.3	15.3	11.0	16.6	11.0	---
Cr----------	15.3	45.0	30.9	28.4	71.2	36.8	34.0	42.6	21.5	---
Cs----------	1.4	4.5	1.9	1.3	4.0	4.3	2.5	3.5	4.3	---
Eu----------	1.03	1.1	.93	1.04	.92	.88	.66	.76	.75	---
Gd----------	4.8	5.5	4.3	4.5	4.9	3.5	2.7	3.9	3.8	---
Hf----------	5.0	3.3	3.2	3.2	3.4	2.2	3.2	3.5	3.9	---
Ho----------	.8	.6	.6	1.0	.9	<.8	<.6	.6	.6	---
La----------	28	25	32	12	12	13	10	11	16	---
Lu----------	.41	.38	.30	.35	.43	.31	.16	.28	.29	---
Nd----------	26	26	24	19	18	16	10	14	14	---
Rb----------	56	87	65	49	87	67	60	60	92	---
Sc----------	14.59	15.96	16.42	16.43	26.29	18.51	9.0	18.0	12.7	---
Sm----------	4.7	5.5	4.3	4.3	4.6	3.8	2.5	3.4	3.3	---
Ta----------	.62	.93	.38	.45	.43	.59	.26	.34	.53	---
Tb----------	.77	.95	.67	.81	<1.38	<1.23	.44	.58	.45	---
Th----------	5.4	8.4	13.0	.8	4.0	6.0	8.0	8.4	6.7	---
Tm----------	.38	.45	.30	.42	.39	.38	.17	.28	.29	---
U-----------	<1.1	1.5	<1.0	<.9	2.1	1.5	1.5	1.7	1.6	---
Yb----------	2.8	2.8	2.0	2.4	2.8	2.0	1.1	2.0	2.0	---
Zn----------	91	88	92	105	134	98	64	74	60	---
Zr----------	198	<310	<319	<314	<422	<377	135	150	150	---

TABLE 30.—*Chemical data for the granodiorite of Claraville*

[Chemical analyses in weight percent. Cross, Iddings, Pirrson, and Washington (CIPW) norms in volume percent. Semiquantitative emission-spectrographic analyses and instrumental neutron-activation analyses in parts per million. See table 34 for analysts]

Sample	3737	4093	4102A	4314	4341	4354	4402	4472	4528	RWK-7	RWK-8
Chemical analyses											
SiO_2	63.7	67.5	68.0	64.3	65.5	67.3	68.2	66.9	70.9	65.8	63.8
Al_2O_3	16.1	16.3	16.0	16.6	16.5	16.1	16.0	16.2	15.4	16.2	17.1
Fe_2O_3	1.4	1.6	.97	1.3	.93	.42	1.1	1.4	.91	1.5	1.3
FeO	3.0	2.0	2.3	2.8	3.0	3.2	1.6	2.2	1.5	2.6	3.5
MgO	1.4	1.2	.99	1.4	1.0	1.1	.87	1.1	.79	1.2	1.3
CaO	4.2	3.5	3.2	4.0	3.8	3.7	3.3	3.5	2.1	3.5	3.9
Na_2O	4.1	3.6	3.7	3.3	4.0	3.8	3.7	4.1	3.6	4.0	4.1
K_2O	3.7	2.7	2.7	2.8	2.9	2.7	3.2	2.8	3.5	2.9	2.5
H_2O^+	2.1	.61	.86	.85	.78	.69	.76	.73	.71	.52	.81
H_2O^-	.37	.18	.14	.15	.04	.14	.13	.20	.22	.38	.29
TiO_2	.91	.63	.63	.75	.70	.66	.64	.70	.37	.77	.82
P_2O_5	.26	.15	.11	.22	.17	.15	.17	.16	.07	.23	.25
MnO	.12	.06	.05	.07	.06	.06	.05	.08	.05	.04	.04
CO_2	.04	.04	.06	.02	.22	.11	.15	.14	.10	.03	.02
Total	101.40	100.07	99.71	98.56	99.60	100.13	99.87	100.21	100.22	99.67	99.73
CIPW norms											
Q	15.14	27.11	27.66	23.68	21.68	24.55	27.10	23.68	31.14	22.09	19.26
or	22.10	16.07	16.16	16.96	17.35	16.08	19.11	16.67	20.83	17.35	14.98
ab	35.07	30.39	31.27	28.47	32.62	31.59	30.51	33.90	29.94	34.27	35.18
an	14.76	16.50	15.36	18.87	17.96	17.52	15.42	16.44	10.03	15.87	17.83
wo	1.81	---	---	---	---	---	---	---	---	---	---
en	3.52	3.01	2.50	3.57	2.52	2.51	2.19	2.76	1.98	3.03	3.28
fs	3.11	1.43	2.51	3.03	3.74	4.59	1.08	1.89	1.50	2.37	4.13
mt	2.05	2.34	1.43	1.93	1.37	.61	1.61	2.05	1.33	2.20	1.91
il	1.75	1.21	1.21	1.46	1.35	1.26	1.23	1.34	.71	1.48	1.58
ap	.62	.36	.26	.53	.41	.36	.41	.38	.17	.55	.60
C	---	1.52	1.54	1.46	.60	.72	1.08	.65	2.20	.75	1.22
cc	.09	---	---	---	---	---	---	---	---	.07	.05
Total	100.02	99.94	99.90	99.96	99.60	99.79	99.74	99.76	99.83	100.03	100.02

TABLE 30.—*Chemical data for the granodiorite of Claraville*—Continued

Sample-------	3737	4093	4102A	4314	4341	4354	4402	4472	4528	RWK-7	RWK-8
					Emission-spectrographic analyses						
B-----------	<4.6	5.5	7.9	19	<4.6	5.8	<4.6	4.9	<4.6	---	---
Ba----------	1,500	1,000	1,100	1,100	1,800	1,300	1,200	1,500	910	---	---
Be----------	2.1	1.9	1.8	2.1	1.6	2.3	1.6	2.1	2.3	---	---
Co----------	6.6	7.9	7.2	10	7.3	7.7	3.7	6.4	3.7	---	---
Cr----------	3.0	3.3	2.6	1.2	2.0	3.0	1.9	1.9	1.9	---	---
Cu----------	1.8	6.2	2.3	3.8	2.2	2.3	2.4	3.8	2.7	---	---
Ga----------	25	21	25	24	28	25	22	25	21	---	---
La----------	32	35	<10	34	40	37	22	37	24	---	---
Mn----------	850	470	420	680	480	470	340	690	330	---	---
Nb----------	11	10	11	11	20	8.7	12	23	13	---	---
Ni----------	3.8	3.9	3.6	5.2	3.4	2.7	2.3	3.3	3.4	---	---
Pb----------	16	17	24	21	22	25	20	21	27	---	---
Sc----------	7.0	6.2	2.8	7.7	6.1	5.7	1.6	4.2	3.6	---	---
Sr----------	1,200	840	670	710	820	710	700	840	610	---	---
V-----------	45	58	40	57	33	31	30	43	34	---	---
Y-----------	14	8.2	3.2	15	7.3	8.3	5.9	10	5.8	---	---
Yb----------	1.0	.48	.50	1.8	.57	.52	1.4	1.4	.86	---	---
Zn----------	86	75	84	97	91	71	37	89	43	---	---
Zr----------	130	110	94	69	360	120	76	310	85	---	---
					Instrumental neutron-activation analyses						
Ba----------	1,519	1,130	1,100	1,150	1,755	1,170	1,285	1,240	915	---	---
Ce----------	52	70	49	75	92	76	81	96	51	---	---
Co----------	5.7	5.9	5.1	7.1	5.0	5.0	3.4	4.7	3.1	---	---
Cr----------	4.3	3.8	3.7	12.2	3.3	4.7	3.2	1.9	3.0	---	---
Cs----------	2.1	1.2	3.1	3.3	2.7	2.5	1.5	1.0	1.7	---	---
Eu----------	1.47	1.07	.94	1.31	1.23	1.06	.78	1.72	.71	---	---
Gd----------	3.7	4.3	3.7	7.1	5.2	5.0	3.4	7.4	2.7	---	---
Hf----------	5.8	4.9	4.1	5.1	6.4	5.3	5.0	5.8	3.5	---	---
Ho----------	<.8	<.7	<.9	1.0	<.9	<1.0	.3	<.9	<.7	---	---
La----------	21	39	25	37	50	42	40	50	29	---	---
Lu----------	.13	.12	.11	.27	.12	.13	.22	.22	.10	---	---
Nd----------	27	28	20	37	35	32	29	42	20	---	---
Rb----------	89	76	104	100	84	104	92	95	112	---	---
Sc----------	4.71	4.15	2.96	5.78	3.95	3.55	2.0	3.39	2.85	---	---
Sm----------	5.1	4.7	3.9	7.8	5.7	5.6	4.0	9.0	3.3	---	---
Ta----------	1.13	.9	.87	1.68	.75	.9	2.70	1.88	.75	---	---
Tb----------	.39	.51	.41	.81	.50	.55	.30	.79	.26	---	---
Th----------	3.4	8.5	6.3	11.5	10.4	12.0	16.8	11.0	8.6	---	---
Tm----------	<.14	.17	.14	.35	.22	.20	<.20	.20	.11	---	---
U-----------	1.9	1.8	2.7	3.4	2.5	3.3	5.0	3.2	2.3	---	---
Yb----------	.9	.9	.8	2.2	.8	1.0	1.5	1.5	.7	---	---
Zn----------	93	71	78	80	91	77	48	74	51	---	---
Zr----------	215	200	162	200	245	210	215	239	153	---	---

TABLE 31.—*Chemical data for the granodiorites of Lebec and Gato-Montes*

[Chemical analyses in weight percent. Cross, Iddings, Pirrson, and Washington (CIPW) norms in volume percent. Semiquantitative emission-spectrographic analyses and instrumental neutron-activation analyses in parts per million. See table 34 for analysts]

	Granodiorite of Lebec					Granodiorite of Gato-Montes			
Sample-------	686	690	698	3088	3181	3356	3534B	3733A	3756
				Chemical analyses					
SiO_2---------	68.0	68.5	67.8	65.4	68.0	66.8	68.1	65.4	71.4
Al_2O_3--------	16.2	16.0	15.7	16.8	16.1	16.0	15.3	17.0	14.2
Fe_2O_3--------	.26	.45	.31	.90	.90	.64	.94	1.3	.77
FeO----------	2.4	2.5	2.9	2.7	2.4	2.7	2.5	2.6	1.4
MgO----------	1.0	1.0	1.6	1.1	1.1	1.1	.97	1.3	.51
CaO----------	3.2	3.2	3.3	4.1	3.4	3.6	3.4	4.3	1.9
Na_2O---------	3.9	3.5	3.2	3.6	3.2	4.0	4.0	4.3	3.6
K_2O----------	3.4	3.4	3.2	2.8	2.9	3.2	3.1	2.1	4.3
H_2O^+---------	.65	.55	.73	.68	.77	.71	.88	.92	.64
H_2O^----------	.08	.06	.08	.20	.15	.13	.14	.16	.12
TiO_2---------	.54	.50	.54	.76	.60	.68	.60	.78	.35
P_2O_5---------	.16	.15	.15	.23	.18	.23	.19	.27	.13
MnO----------	.07	.07	.10	.03	.04	.05	.08	.05	.03
CO_2----------	.19	.11	.30	.09	.05	.02	.06	---	---
Total---	100.05	99.99	99.91	99.39	99.79	99.86	100.26	100.48	99.35
				CIPW norms					
Q-----------	23.52	26.06	27.00	23.19	29.09	21.30	23.90	24.64	29.31
or----------	20.23	20.22	19.08	16.80	17.33	19.10	18.50	7.20	25.77
ab----------	33.23	29.80	27.32	30.92	27.39	34.18	34.11	36.94	30.90
an----------	13.72	14.30	13.62	18.55	15.51	16.39	14.75	19.90	8.70
wo----------	---	---	---	---	---	---	.26	---	---
en----------	2.51	2.51	4.02	2.78	2.77	2.77	2.43	3.29	1.29
fs----------	3.50	3.55	4.40	3.06	2.78	3.43	3.00	2.54	1.43
mt----------	.40	.66	.50	1.33	1.32	.94	1.37	1.91	1.13
il----------	1.03	.96	1.04	1.47	1.15	1.30	1.15	1.50	.67
ap----------	.40	.36	.36	.55	.43	.55	.45	.65	.31
C-----------	1.12	1.37	2.05	1.17	2.09	.01	---	1.48	.49
cc----------	.44	.25	.69	.21	.12	.05	.14	---	---
Total---	100.10	100.04	100.08	100.03	99.98	100.02	100.06	100.05	100.00

TABLE 31.—*Chemical data for the granodiorites of Lebec and Gato-Montes*—Continued

	Granodiorite of Lebec					Granodiorite of Gato-Montes			
Sample-------	686	690	698	3088	3181	3356	3534B	3733A	3756
				Emission-spectrographic analyses					
B------------	1.5	10	10	9.1	4.4	8.0	5.6	8.3	<4.6
Ba-----------	1,000	1,000	500	900	750	1,200	1,100	640	980
Be-----------	2	3	3	2.3	1.8	2.5	3.7	3.1	2.6
Ce-----------	100	---	70	110	75	91	250	140	83
Co-----------	5	5	7	5.6	5.8	5.4	5.2	6.8	1.7
Cr-----------	5	5	10	4.4	8.9	2.6	2.6	3.0	<1.0
Cu-----------	7	5	7	2.5	2.4	<1.5	<1.5	<1.5	2.2
Ga-----------	20	30	20	22	19	26	31	33	23
Gd-----------	---	---	---	<6.8	<6.8	10	11	9.1	10
La-----------	50	30	50	47	32	43	150	69	45
Mn-----------	---	---	---	400	370	730	1,200	750	480
Nb-----------	15	15	15	8.7	4.7	9.3	19	12	9.3
Nd-----------	---	---	---	50	---	---	88	---	---
Ni-----------	---	---	3	5.2	5.4	4.1	3.5	4.0	2.0
Pb-----------	20	30	20	14	19	16	23	14	26
Sc-----------	3	7	7	5.7	6.1	4.6	5.6	5.9	3.1
Sr-----------	500	500	500	570	420	1,000	720	910	570
V------------	50	30	50	32	38	33	27	39	16
Y------------	15	20	20	14	7.2	12	26	12	12
Yb-----------	.7	1.5	1.5	1.8	.8	1.2	4.3	1.5	1.5
Zn-----------	---	---	---	73	51	71	85	120	28
Zr-----------	150	200	100	82	87	130	160	150	180
				Instrumental neutron-activation analyses					
Ba-----------	---	---	---	1,435	1,090	1,320	970	650	1,180
Ce-----------	---	---	---	70	63	74	82	88	62
Co-----------	---	---	---	5.7	6.4	5.3	4.5	6.5	2.3
Cr-----------	---	---	---	5.6	9.8	4.7	<8.8	3.1	2.3
Cs-----------	---	---	---	2.8	2.7	2.5	3.8	2.9	1.5
Eu-----------	---	---	---	1.28	1.02	1.20	1.16	1.23	.91
Gd-----------	---	---	---	5.5	3.9	3.3	5.3	4.1	3.8
Hf-----------	---	---	---	5.9	4.2	4.8	5.9	6.3	5.1
Ho-----------	---	---	---	---	---	<.6	<.8	<.7	<.6
La-----------	---	---	---	36	---	40	45	45	33
Lu-----------	---	---	---	.20	.12	.11	.33	.14	.15
Nd-----------	---	---	---	35	26	27	35	34	27
Rb-----------	---	---	---	96	95	83	129	88	135
Sc-----------	---	---	---	4.53	4.43	3.4	4.06	4.86	3.3
Sm-----------	---	---	---	6.9	---	4.1	6.0	5.1	4.7
Ta-----------	---	---	---	1.90	.76	.88	2.74	1.2	1.1
Tb-----------	---	---	---	.64	.34	.47	.80	.59	.5
Th-----------	---	---	---	9.7	9.9	8.4	18.0	13.9	10.1
Tm-----------	---	---	---	.22	.26	.21	.46	.24	.21
U------------	---	---	---	4.2	---	2.0	4.3	2.1	1.4
Yb-----------	---	---	---	1.5	.7	.7	2.2	.9	1.0
Zn-----------	---	---	---	98	74	84	100	100	64
Zr-----------	---	---	---	212	188	222	269	305	227

TABLE 32.—*Chemical data for the tonalites of Mount Adelaide and Hoffman Canyon, the granodiorite of Sorrell Peak, and sample RWK-11*

[Chemical analyses in weight percent. Cross, Iddings, Pirrson, and Washington (CIPW) norms in volume percent. Semiquantitative emission-spectrographic analyses and instrumental neutron-activation analyses in parts per million. See table 34 for analysts]

	Tonalite of Mount Adelaide				Tonalite of Hoffman Canyon		Granodiorite of Sorrell Peak	
Sample-------	3631	4189	4273	4294	3842A	RWK-9	4366	RWK-11
				Chemical analyses				
SiO_2---------	65.2	67.3	65.2	61.5	62.8	63.8	71.0	59.0
Al_2O_3--------	17.8	16.8	16.7	17.9	16.8	16.3	15.5	16.8
Fe_2O_3--------	1.3	.77	1.8	1.2	1.8	1.7	.85	1.4
FeO----------	2.2	2.4	1.9	3.9	3.0	3.4	2.0	4.5
MgO----------	1.6	1.4	1.2	2.2	1.8	2.1	.68	2.6
CaO----------	5.2	4.1	4.3	5.5	4.8	4.5	2.3	5.2
Na_2O---------	4.6	3.9	3.9	3.7	4.0	3.7	3.1	4.7
K_2O----------	.84	1.4	2.4	1.8	2.3	2.8	3.8	2.6
H_2O^+---------	1.1	.74	.71	.85	1.0	.45	.59	.75
H_2O^----------	.36	.18	.29	.15	.16	.28	.14	.35
TiO_2---------	.55	.47	.68	.86	.92	.87	.40	1.2
P_2O_5---------	.19	.11	.15	.18	.29	.23	.07	.38
MnO----------	.05	.06	.06	.08	.07	.06	.04	.10
CO_2----------	.01	.03	.02	.03	.26	.04	.03	.08
Total---	101.00	99.66	99.31	99.85	100.00	100.23	100.50	99.66
				CIPW norms				
Q------------	21.38	27.78	22.79	16.65	18.31	18.25	31.82	6.91
or-----------	4.99	8.38	14.43	10.76	13.75	16.63	22.51	15.59
ab-----------	39.10	33.20	33.42	31.45	34.24	31.47	26.07	40.35
an-----------	24.60	19.87	20.70	26.41	20.51	19.70	10.98	17.31
wo-----------	---	---	---	---	---	.41	---	2.43
en-----------	4.00	3.53	3.04	5.54	4.54	5.26	1.70	6.57
fs-----------	2.16	3.15	1.01	4.96	2.66	3.53	2.39	5.39
mt-----------	1.89	1.13	2.66	1.76	2.64	2.48	1.24	2.06
il-----------	1.05	.90	1.31	1.65	1.77	1.66	.76	2.31
ap-----------	.45	.26	.36	.43	.70	.55	.17	.91
C------------	.35	1.74	.26	.34	.30	---	2.32	---
cc-----------	.02	---	---	---	.60	.09	---	.19
Total---	99.99	99.94	99.98	99.95	100.02	100.03	99.96	100.02

TABLE 32.—*Chemical data for the tonalites of Mount Adelaide and Hoffman Canyon, the granodiorite of Sorrell Peak, and sample RWK-11*—Continued

	Tonalite of Mount Adelaide				Tonalite of Hoffman Canyon		Granodiorite of Sorrell Peak	
Sample	3631	4189	4273	4294	3842A	RWK-9	4366	RWK-11
	Emission-spectrographic analyses							
B	5.4	6.9	7.7	6.5	<4.6	---	6.7	---
Ba	370	710	730	700	710	---	1,500	---
Be	1.5	1.2	1.5	1.3	2.6	---	1.2	---
Ce	68	<63	<63	<63	130	---	---	---
Co	7.3	9.9	8.2	15	8.7	---	4.6	---
Cr	12	15	1.9	10	7.0	---	3.0	---
Cu	<1.5	2.6	3.1	6.2	3.2	---	1.9	---
Ga	24	20	24	24	27	---	23	---
Gd	<6.8	<15	<15	<15	9.6	---	---	---
La	28	<10	19	26	58	---	19	---
Mn	850	540	590	680	1,300	---	310	---
Nb	6.7	<3.2	10	8.4	13	---	8.9	---
Ni	6.6	5.6	3.6	5.8	6.1	---	3.2	---
Pb	8.3	17	25	21	12	---	22	---
Sc	8.4	8.3	4.6	10	9.3	---	2.9	---
Sn	4.3	7.1	7.4	---	1.6	---	4.1	---
Sr	990	570	830	700	880	---	610	---
V	49	48	51	86	88	---	20	---
Y	5.9	5.3	5.8	11	22	---	5.0	---
Yb	.54	.32	.61	.81	2.3	---	.61	---
Zn	79	73	83	94	94	---	45	---
Zr	140	86	65	98	270	---	76	---
	Instrumental neutron-activation analyses							
Ba	363	603	619	687	818	---	1,265	---
Ce	32	15	35	53	64	---	68	---
Co	7.2	6.8	5.6	10.2	8.5	---	3.5	---
Cr	11.2	13.5	2.4	9.8	7.5	---	3.7	---
Cs	<1	1.8	3.1	2.6	1.3	---	3.8	---
Eu	.8	.63	.87	1.1	1.38	---	.94	---
Gd	1.5	2.0	2.8	4.1	4.1	---	4.0	---
Hf	2.9	2.2	3.8	3.2	5.6	---	4.5	---
Ho	<.6	.2	<.7	.5	<.9	---	<.7	---
La	16	8	18	27	32	---	37	---
Lu	.06	.09	.1	.16	.18	---	.12	---
Nd	12	8	17	26	31	---	25	---
Rb	15	37	69	60	71	---	112	---
Sc	6.28	5.85	3.73	6.70	7.25	---	3.03	---
Sm	1.7	1.8	3.3	4.6	4.9	---	4.5	---
Ta	<.57	.26	.63	.71	1.06	---	.73	---
Tb	<.7	.23	.45	.55	.66	---	.45	---
Th	3.3	1.4	10.1	7.0	8.2	---	13.6	---
Tm	<.11	.12	<.10	.23	.21	---	.17	---
U	<.8	.5	3.0	1.4	1.4	---	3.2	---
Yb	.3	.7	.7	1.1	1.2	---	.9	---
Zn	71	60	75	86	96	---	64	---
Zr	<227	88	138	120	333	---	182	---

TABLE 33.—*Chemical data for the granites of Tejon Lookout, Brush Mountain, and Tehachapi Airport*

[Chemical analyses in weight percent. Cross, Iddings, Pirrson, and Washington (CIPW) norms in volume percent. Semiquantitative emission-spectrographic analyses and instrumental neutron-activation analyses in parts per million. See table 34 for analysts]

	Granite of Tejon Lookout			Granite of Brush Mountain			Granite of Tehachapi Airport
Sample	3401	3466	3752	726	3005	3221	4100A
Chemical analyses							
SiO_2	74.8	76.6	75.5	75.6	74.8	75.7	75.6
Al_2O_3	12.5	12.7	12.8	13.2	13.6	13.2	14.1
Fe_2O_3	1.4	1.2	1.1	.24	.71	.85	.57
FeO	.50	.48	.28	.96	.96	.24	.36
MgO	.71	.23	.08	.17	.08	.14	.16
CaO	1.4	.97	.60	.87	.84	.43	.67
Na_2O	4.2	3.8	2.3	3.7	3.0	3.8	3.3
K_2O	4.8	4.2	6.9	4.4	4.4	4.3	4.9
H_2O^+	.68	.57	.42	.04	.49	.61	.48
H_2O^-	.19	.25	.13	.54	.06	.03	.16
TiO_2	.13	.12	.08	.11	.09	.08	.07
P_2O_5	.07	.06	.05	.04	.04	.04	.01
MnO	.08	.04	---	.07	.02	.02	.03
CO_2	.01	.01	---	<.05	.03	.02	.02
Total	101.47	100.83	100.24	99.94	99.12	99.46	100.43
CIPW norms							
Q	29.15	36.05	34.56	34.77	38.86	36.61	36.13
or	28.20	24.72	40.91	26.17	26.38	25.71	29.02
ab	35.33	32.02	19.52	31.51	25.75	32.54	27.83
an	1.07	4.34	2.66	4.08	3.77	1.77	3.27
wo	2.22	---	---	---	---	---	---
en	1.76	.57	.20	.43	.20	.35	.40
fs	---	---	---	1.52	1.08	---	.13
mt	1.49	1.32	.67	.35	1.04	.61	.83
il	.25	.23	.15	.21	.17	.15	.13
ap	.17	.14	.12	.10	.10	.10	.02
C	---	.31	.58	.87	2.58	1.68	2.21
cc	.02	.02	---	---	.07	.05	---
hm	.37	.28	.64	---	---	.44	---
Total	100.03	100.00	100.01	100.01	100.00	100.01	99.97

TABLE 33.—*Chemical data for the granites of Tejon Lookout, Brush Mountain, and Tehachapi Airport*—Continued

	Granite of Tejon Lookout			Granite of Brush Mountain			Granite of Tehachapi Airport
Sample-------	3401	3466	3752	726	3005	3221	4100A
			Emission-spectrographic analyses				
B------------	<4.6	<4.6	7.8	10	<3.2	5.0	<4.6
Ba-----------	600	650	1,900	1,000	610	250	1,100
Be-----------	4.9	5.8	1.4	3	2.1	3.2	<1.0
Ce-----------	62	48	<43	---	41	84	<43
Co-----------	<1.0	1.0	1.0	---	1.6	1.3	<1.0
Cr-----------	---	---	---	1	---	---	1.3
Cu-----------	---	---	---	3	---	---	2.1
Ga-----------	21	22	17	15	15	17	14
La-----------	37	17	20	---	21	36	22
Mn-----------	440	470	150	---	250	270	240
Nb-----------	13	13	<2.2	10	7.8	13	5.7
Nd-----------	47	---	---	---	---	---	---
Ni-----------	<1.5	2.1	<1.5	---	<4.6	<4.6	2.9
Pb-----------	15	19	38	30	18	17	31
Sc-----------	2.7	3.0	1.2	3	3.3	2.4	2.2
Sn-----------	2.8	3.9	1.9	---	---	---	---
Sr-----------	130	140	400	70	50	18	120
V------------	4.7	4.9	17	---	4.2	<3.2	2.2
Y------------	32	24	3.9	30	17	23	16
Yb-----------	3.9	5.4	2.1	2	1.7	3.8	2.8
Zn-----------	24	33	22	---	23	27	18
Zr-----------	66	55	83	100	54	36	69
			Instrumental neutron-activation analyses				
Ba-----------	572	540	2,200	---	1,050	352	900
Ce-----------	50	43	59	---	57	81	55
Co-----------	.9	1.0	.8	---	.8	.2	.4
Cr-----------	<6.7	3.7	<6.5	---	<3.0	<8.0	2.8
Cs-----------	1.2	5.3	4.6	---	5.4	4.7	1.2
Eu-----------	.43	.37	.94	---	.64	.31	.39
Gd-----------	4.9	3.8	2.0	---	7.2	7.4	5.0
Hf-----------	3.5	4.5	3.7	---	5.3	7.3	2.3
Ho-----------	.8	<.6	<.6	---	---	---	<.6
La-----------	25	16	28	---	25	43	30
Lu-----------	.4	.46	.28	---	.52	.60	.39
Nd-----------	22	14	21	---	25	32	21
Rb-----------	170	194	180	---	183	245	158
Sc-----------	2.48	2.6	1.53	---	3.61	2.50	1.96
Sm-----------	4.4	3.5	3.0	---	6.5	7.5	4.4
Ta-----------	1.9	2.95	.71	---	1.18	2.83	.48
Tb-----------	.85	.64	.30	---	1.10	1.08	.7
Th-----------	20.7	20	17.5	---	18.9	28.6	15.8
Tm-----------	.39	.46	.18	---	.54	.48	.34
U------------	4.6	4.3	2.5	---	3.9	8.1	2.1
Yb-----------	2.8	3.3	1.8	---	3.4	4.5	2.8
Zn-----------	29	26	16	---	35	50	14
Zr-----------	155	145	88	---	138	223	82

TABLE 34.—*Chemical data for felsic masses within the tonalite of Bear Valley Springs, the quartz diorite to tonalite of Antimony Peak, and the diorite to tonalite of Tehachapi Mountains*

[Chemical analyses in weight percent. Cross, Iddings, Pirrson, and Washington (CIPW) norms in volume percent. Emission-spectrographic analyses and instrumental neutron-activation analyses in parts per million]

Analysts for samples in tables 29 through 34:
Samples 784B, 3005, 3023, 3088, 3097, 3098A, 3158, 3181, and 3221: Hezekiah Smith (chemical analyses) and N. Rait (semiquantitative spectrographic analyses).
Samples 3356, 3401, 3427B, 3442, 3466, 3534B, 3575, 3631, 3670, 3695, 3733A, 3737, 3752, 3756A, 3791, 3794A, 3837, 3842A, and 3858: N. Skinner (chemical analyses), J.L. Harris (semiquantitative spectrographic analyses), and L.J. Schwarz (instrumental neutron-activation analyses).
Samples 4093, 4100A, 4102A, 4181, 4184, 4189, 4273, 4281, 4294, 4314, 4341, 4354, 4366, 4402, 4472, and 4528: J. Reid, P. Hearn, H. Rose, and J. Lindsay (chemical analyses), J.L. Harris (semiquantitative spectrographic analyses), and L.J. Schwarz (instrumental neutron-activation analyses).
Previously published analyses:
Chemical analyses: samples RWK-7, RWK-8, RWK-9, RWK-11, and RWK-13 (Kistler and Peterman, 1978, p. 2); samples 686, 690, 698, and 726 (Ross, 1972, p. 71).
Semiquantitive spectrographic analyses: samples 686, 690, 698, and 726 (Ross, 1972, p. 71).

	Felsic masses within the tonalite of Bear Valley Springs		Quartz diorite to tonalite of Antimony Peak			Diorite to tonalite of Tehachapi Mountains		
Sample	3695	3794A	784B	3023	3158	3097	3098A	3442
Chemical analyses								
SiO_2	68.9	73.6	62.3	61.3	62.3	59.7	55.2	61.6
Al_2O_3	15.6	13.7	20.8	20.2	16.2	17.2	17.2	15.5
Fe_2O_3	1.4	.72	1.3	1.3	1.5	.89	1.2	1.7
FeO	2.2	1.2	2.2	2.2	4.4	5.8	7.1	4.7
MgO	.96	.42	2.4	2.6	2.4	3.4	4.4	4.1
CaO	3.9	2.0	6.3	6.7	3.8	7.0	8.2	6.9
Na_2O	3.9	3.9	3.9	3.6	2.3	3.2	2.8	2.8
K_2O	2.4	3.6	.56	.63	3.4	.26	.59	1.4
H_2O^+	.82	.62	.44	.63	2.2	.58	1.0	1.1
H_2O^-	.16	.13	.40	.36	.22	.16	.27	.19
TiO_2	.50	.24	.33	.36	.86	.82	1.1	.67
P_2O_5	.13	.07	.16	.18	.21	.18	.21	.14
MnO	.07	.02	.08	.07	.07	.11	.01	.08
CO_2	---	---	.02	---	.19	.05	.04	.01
Total	100.94	100.22	101.19	100.13	100.05	99.35	99.42	100.89
CIPW norms								
Q	26.74	32.07	19.66	19.25	23.33	17.23	10.05	18.36
or	14.19	21.39	3.30	3.76	20.58	1.56	3.55	8.31
ab	33.01	33.18	32.89	30.73	19.93	27.46	24.14	23.79
an	17.98	9.29	29.98	32.34	16.67	32.25	33.23	25.69
wo	.22	.09	---	---	---	.61	2.74	3.21
en	2.39	1.05	5.96	6.53	6.12	8.59	11.17	10.25
fs	2.19	1.26	2.56	2.52	5.69	8.89	10.63	6.29
mt	2.03	1.05	1.88	1.90	2.23	1.31	1.77	2.48
il	.95	.46	.63	.69	1.67	1.58	2.13	1.28
ap	.31	.17	.38	.43	.51	.43	.51	.33
C	---	---	2.74	1.86	2.84	---	---	---
cc	---	---	.05	---	.44	.11	.09	.02
Total	100.01	100.01	100.03	100.01	100.01	100.02	100.01	100.01

TABLE 34.—*Chemical data for felsic masses within the tonalite of Bear Valley Springs, the quartz diorite to tonalite of Antimony Peak, and the diorite to tonalite of Tehachapi Mountains*—Continued

	Felsic masses within the tonalite of Bear Valley Springs		Quartz diorite to tonalite of Antimony Peak			Diorite to tonalite of Tehachapi Mountains		
Sample-------	3695	3794A	784B	3023	3158	3097	3098A	3442
Emission-spectrographic analyses								
B-----------	8.1	<4.6	4.1	4.0	33	9.8	17	<4.6
Ba----------	960	970	100	130	800	90	200	470
Be----------	1.7	1.3	.99	.92	1.6	<.68	.91	1.3
Ce----------	66	51	---	---	120	---	---	---
Co----------	6.7	3.3	7.4	8.2	11	15	19	23
Cr----------	4.6	3.2	13	17	16	18	67	120
Cu----------	<1.5	<1.5	<2.2	3.6	15	6.2	21	27
Ga----------	25	23	18	18	19	18	21	20
Gd----------	11	8.8	---	---	---	---	10	12
La----------	31	29	12	15	39	<10	<10	28
Mn----------	880	450	610	640	670	1,200	1,300	1,700
Mo----------	---	---	<2.2	2.4	2.8	3.6	3.9	<2.2
Nb----------	5.3	3.0	<3.2	<3.2	11	3.4	7.5	6.5
Nd----------	---	---	---	---	51	---	---	---
Ni----------	5.2	2.7	16	20	8.7	11	16	37
Pb----------	12	14	14	10	16	12	14	9.1
Sc----------	8.6	7.1	9.8	12	16	24	29	26
Sr----------	460	270	540	490	400	260	260	460
V-----------	49	24	47	56	79	130	140	170
Y-----------	17	21	5.5	7.6	23	18	20	30
Yb----------	2.5	2.9	1.3	1.3	3.2	2.7	4.5	3.6
Zn----------	47	13	35	24	79	60	75	92
Zr----------	97	31	29	33	110	52	43	29
Instrumental neutron-activation analyses								
Ba----------	1,103	959	174	212	---	---	264	439
Ce----------	40	29	14	14	---	---	27	25
Co----------	6.0	2.8	9.2	10.3	---	---	25.1	20.1
Cr----------	4.7	2.3	23.4	32.3	---	---	69.2	104.7
Cs----------	2.4	.7	.8	.5	---	---	1.1	1.2
Eu----------	1.03	.71	.78	.70	---	---	1.18	.84
Gd----------	3.3	2.8	2.0	2.4	---	---	5.1	2.8
Hf----------	4.3	3.8	2.7	2.9	---	---	4.0	2.6
Ho----------	.5	<.8	---	---	---	---	---	.5
La----------	16	13	7	7	---	---	11	10
Lu----------	.30	.25	.21	.18	---	---	.51	.32
Nd----------	14	16	13	<20	---	---	<40	18
Rb----------	86	99	9	16	---	---	17	48
Sb----------	---	---	.3	.3	---	---	.7	---
Sc----------	7.6	4.95	11.20	10.80	---	---	34.50	20.89
Sm----------	3.4	3.3	2.3	2.1	---	---	5.0	3.5
Ta----------	.51	.37	.14	.11	---	---	.46	.33
Tb----------	.68	.46	.34	.29	---	---	.87	.68
Th----------	7.4	3.0	.3	.4	---	---	2.5	.4
Tm----------	.34	.27	.19	.12	---	---	.53	.32
U-----------	1.6	1.7	<2.0	<2.0	---	---	1.6	<.9
Yb----------	1.9	1.6	1.3	1.1	---	---	3.4	2.1
Zn----------	62	41	54	47	---	---	101	94
Zr----------	94	138	120	140	---	---	150	<353

TABLE 35.—*Potassium and rubidium data for some plutonic-rock units of the southernmost Sierra Nevada*

Sample	K (wt pct)	Rb (ppm)	K/Rb
Tonalite of Bear Valley Springs			
3427	1.6	56	286
3575	1.6	87	184
3670	1.5	65	231
3791	1.2	49	245
3837	2.0	87	230
3858	1.6	67	239
4181	1.4	60	233
4184	1.3	60	217
4281	1.8	92	196
RWK-12	2.0	55	302
RWK-13	1.7	54	262
Average------	1.6	67	239
Granodiorite of Claraville			
3737	1.7	88	348
4093	2.2	76	289
4102A	2.3	104	212
4314	2.3	100	230
4341	2.4	82	293
4354	2.2	104	212
4402	2.7	92	293
4472	2.3	95	242
4528	2.9	112	259
RWK-7	2.4	98	243
RWK-8	2.0	82	242
Average------	2.3	94	260
Granodiorite of Lebec			
3088	2.2	96	229
3181	2.4	95	253
Average------	2.3	95	242
Granodiorite of Gato-Montes			
3356	2.7	83	325
3534B	2.6	129	202
3733A	1.7	88	193
3756	3.6	135	267
Average------	2.7	109	247
Tonalite of Mount Adelaide			
3631	0.7	15	467
4189	1.2	37	324
4273	2.0	69	290
4294	1.5	60	250
Average------	1.4	45	333
Granite of Tejon Lookout			
3401	4.0	170	235
3466	3.5	194	180
3752	5.7	180	317
Average------	4.4	181	244
Granite of Brush Mountain			
3005	3.7	183	202
3221	3.6	245	147
Average------	3.6	214	175
Granite of Sorrell Peak			
4366	3.2	112	286
Granite of Tehachapi Airport			
4100A	4.1	158	259
Felsic masses within the tonalite of Bear Valley Springs			
3695	2.0	86	233
3794A	3.0	95	316
Average------	2.5	91	275
Tonalite of Hoffman Canyon			
3842A	1.9	71	268
RWK-9	2.4	90	263
Average------	2.2	81	266
Quartz diorite to tonalite of Antimony Peak			
784	0.5	9	556
3023	.7	16	438
Average------	.6	12	497
Diorite to tonalite of Tehachapi Mountains			
3098	0.5	17	294
3442	1.2	48	250
Average------	.9	32	272
Unnamed unit			
RWK-11	2.0	80	250
Grand average.	---	---	266

TABLE 36.—*Average enrichment of light and heavy rare-earth elements relative to chondrite in some plutonic-rock units of the southernmost Sierra Nevada*

[REE, rare-earth element. Average chondrite abundances: light REE's, 1.81 ppm; heavy REE's, 0.26 ppm (Haskins and others, 1968). Ranges are for units with multiple samples]

Plutonic-rock unit	Light REE's (La+Ce+Nd)	Heavy REE's (Tm+Yb+Lu)
Tonalite of Bear Valley Springs-----	40 (27-64)	11 (5-14)
Felsic masses within the tonalite of Bear Valley Springs.	36 (32-39)	9 (8-10)
Granodiorite of Claraville----------	76 (55-104)	6 (3-11)
Granodiorite of Gato-Montes---------	82 (67-92)	7 (4-12)
Granodiorite of Lebec---------------	78	5 (4-7)
Tonalite of Mount Adelaide----------	37 (17-59)	3 (1.5-6)
Granite of Tejon Lookout------------	51 (40-60)	13 (9-16)
Granite of Brush Mountain-----------	62 (59-66)	20 (17-23)
Granite of Tehachapi Airport--------	59	13
Tonalite of Hoffman Canyon----------	70	6
Granodiorite of Sorrell Peak--------	72	5
Diorite to tonalite of Tehachapi Mountains.	30 (29-32)	13 (10-17)
Quartz diorite to tonalite of Antimony Peak.	19	6 (5-7)

TABLE 37.—*Heat production and content of heat-producing elements in samples of some granitic rocks of the southernmost Sierra Nevada*

Sample	K (wt pct)	U (ppm)	Th (ppm)	Th/U	Heat production (µcal/g/yr)
Diorite to tonalite of Tehachapi Mountains					
3442	1.2	<0.9	0.4	>0.4	<1.1
Average					<1
Tonalite of Bear Valley Springs					
3427	1.6	<1.1	5.4	<4.9	<2.2
3575	1.6	1.5	8.4	5.6	3.2
3670	1.5	<1.0	13.0	<13	<3.7
3791	1.2	<.9	.8	<.9	<1.2
3837	2.0	2.1	4.0	1.9	2.9
3858	1.6	1.5	6.0	4.0	2.7
4181	1.4	1.6	8.3	5.2	3.2
4184	1.3	1.7	8.4	4.9	3.3
4281	1.8	1.6	6.6	4.1	3.0
Average					<3
Tonalite of Mount Adelaide					
3631	0.7	<0.8	3.3	<4.1	<1.4
4189	1.2	.5	1.4	2.8	1.0
4273	2.0	3.0	10.1	3.4	4.8
4294	1.5	1.4	7.0	5.0	2.8
Average					<3
Felsic masses within the tonalite of Bear Valley Springs					
3695	2.0	1.6	7.4	4.6	3.3
3794	3.0	1.7	3.0	1.8	2.7
Average					3
Tonalite of Hoffman Canyon					
3842	1.9	1.4	8.2	5.9	3.3
Average					3
Granodiorite of Claraville					
3737	3.1	1.9	3.4	1.8	2.9
4093	2.2	1.8	8.4	4.7	3.6
4102A	2.2	2.7	6.3	2.3	3.8
4314	2.3	3.4	11.4	3.4	5.4
4341	2.4	2.6	10.0	3.9	5.6
4354	2.2	3.3	12.0	3.6	5.4
4402	2.7	4.9	16.7	3.4	7.6
4472	2.3	3.2	11.0	3.4	5.2
4528	2.9	2.3	8.6	3.7	4.2
Average					5
Granodiorite of Gato-Montes					
3356	2.7	2.0	8.4	4.2	3.9
3534	2.6	4.3	18.0	4.2	7.4
3733	1.7	2.1	13.8	6.6	4.8
3756	3.6	1.4	10.1	7.2	4.0
Average					5
Granodiorite of Sorrell Peak					
4366	3.2	3.2	13.6	4.3	5.9
Average					6
Granite of Tehachapi Airport					
4100A	4.1	2.1	15.8	7.5	5.8
Average					6
Granite of Tejon Lookout					
3401	4.0	4.6	20.7	4.5	8.6
3466	3.5	4.3	20.0	4.7	8.1
3752	5.7	2.5	17.5	7.0	6.9
Average					8
Grand average.	2.3	2.4	9.4	4.3	4.5
Standard deviation.	1.0	1.1	5.3	1.5	1.9

TABLE 38.—*Metallic-mineral deposits in the southernmost Sierra Nevada (south of lat 35°30'N.)*

[Only mines with known production are listed. Map numbers and data from Troxel and Morton (1962)]

Map No.	Mine name	Location	Mode of occurrence	Development and production
			Gold	
108	Amy (Gold State)---------	NE1/4 sec. 5, T. 29 S., R. 34 E.	Several parallel quartz veins, 5 to 100 cm thick, in sheeted zones in granodiorite. Rich ore in kidneylike masses	20-m-long shaft and several drifts, as long as 200 m, with stopes, now caved and flooded. Production of about $50,000 in gold.
119	Bella Ruffin (Ruffin, Berry).	Sec. 35, T. 29 S., R. 33 E.	Vertical vein, 1 m thick, in granitic rock; ore in pockets.	36-m-long shaft with short drifts and one stope. Production of 1 to 2 kg Au.
126	Black Bob----------------	SW1/4 sec. 10, T. 9 N., R. 20 W.	Quartz vein, 1 m thick, in granitic rock; some lead in ore.	46-m-long shaft with a few hundred meters of drifts. Production in 1932-34 of 800 t of ore (no grade reported).
136	Bright Star--------------	Ctr. sec. 18, T. 28 S., R. 34 E.	Quartz vein, 50 cm in average width, with an ore shoot 37 m long.	165-m-long shaft with three levels and more than 1,000 m of drifts and stopes; main shaft caved and covered. Production before 1900 of $600,000, some additional production in 1898-1903 from tailings, and small production in 1936-41 from mine.
140	Burning Moscow-----------	N1/2 sec. 25, T. 29 S., R. 31 E.	Quartz vein, 1 m thick, in quartz monzonite-----	46-m-long shaft with 168-m-long crosscut adit and more than 100 m of additional workings. Minor production in about 1934 of gold and copper.
167	Ellston (Producer)-------	SW1/4 sec. 9, T. 29 S., R. 31 E.	Quartz veins in quartz diorite; gold-bearing pockets.	24-m-long shaft with short drifts. Small production in 1913-14.
168	Esperanza----------------	Secs. 5, 6, T. 29 S., R. 37 E.	Quartz vein, 15 to 100 cm thick, in granitic rock. Ore shoots, as much as 30 m long, contain free gold and auriferous sulfides.	More than 100 m of workings. Production in 1932-35 of 1 to 2 kg Au.
170	Ferris (Golden, Jack Rabbit).	N1/2 sec. 24, T. 30 S., R. 32 E.	Quartz veins in schist and quartz diorite-------	67-m-long shaft with levels at 20, 33, and 61 m. Production in 1897-1914 of 15 to 20 kg Au.
174	French (Bowman, French Meadows, Trestle).	Ctr. SW1/4 sec. 29, T. 28 S., R. 34 E.	Quartz vein, as much as 1 m thick, in granitic rock, containing gold, pyrite, and marcasite; two main ore shoots.	200-m-long adit with unknown length of lateral workings (now caved). Production in 1906-7 and 1937-41 of 15 to 20 kg Au.
197	Gold Standard prospect---	NW1/4NW1/4 sec. 19, T. 28 S., R. 34 E.	Two quartz veins, as much 1/3 m thick, in phyllite, containing free gold and chalcopyrite.	Drifts, not extensive workings. Minor production in 1931-40 of gold and copper.
207	Gwynne (Jennette)--------	E1/2 sec. 21, W1/2 sec. 22, T. 29 S., R. 34 E.	Three northeast-striking quartz veins, as much as 3 m thick, containing free gold, pyrite, marcasite, arsenopyrite, and some scheelite.	More than 2,500 m of drifts, winzes, and stopes. Recorded production, mostly before 1933, of $770,000 in gold from ore grading $20 to $50 per ton.
210	Hart---------------------	Sec. 13, T. 30 S., R. 32 E.	Quartz vein, as much as 2 m thick, in quartz diorite.	Adit, at least 50 m long, with caved stopes. Production in 1932-41 of 15 to 20 kg Au.
222	Hub----------------------	Ctr. N1/2 sec. 4, T. 31 S., R. 35 E.	Quartz vein, less than 1 m thick, in rhyolite.	Shallow shaft and trenches. Minor production in 1939.
231	Jenette-Grant------------	NW1/4 sec. 18, T. 28 S., R. 34 E.	Quartz veins in metamorphic rock, containing gold, pyrite, and chalcopyrite.	290-m-long crosscut adit, with some other workings. Minor production of gold and copper.
236	Joe Walker---------------	E1/2 sec. 12, T. 29 S., R. 32 E.	Quartz vein, 2/3 to 6 m thick, in granitic rock; auriferous pyrite and arsenopyrite.	A few hundred meters of inclined shafts and drifts. Production in 1865-74 of $500,000 to $600,000 in gold, and in 1951 of about 28 kg Au.
237	Juan Dosie---------------	Sec. 2, T. 30 S., R. 33 E.	Quartz vein, 1/3 to 1 m thick, in granitic rock.	67-m-long shaft with levels at 24, 30, 46, and 61 m; a few hundred meters of level workings. Production of about 45 kg Au.
256	Lone Star----------------	NE1/4 sec. 18, T. 28 S., R. 34 E.	Quartz veins, about 1 m thick, in fine-grained metamorphic rock, containing free gold, pyrite, and arsenopyrite.	Veins developed by shallow shafts and drifts. Unknown production in 1896-98, 1912-13, 1940-48, and intermittently in the 1930's.
302	Pine Tree (American, Victoria).	NW. cor. sec. 3, NE. cor. sec. 4, T. 11 N., R. 15 W.	Quartz veins, as much as 1 m thick, in granitic rock, containing free gold, sparse sulfides, and, locally, scheelite.	Five adits, as long as 244 m, and a few thousand meters of drifts and stopes. Production in 1876-1907 of $250,000 in gold.
324	Retreat------------------	Secs. 4, 5, 8, T. 29 S., R. 34 E.	Gold- and silver-bearing quartz veins, 20 to 50 cm thick, in deeply weathered, sheared granitic rock. Ore in "kidneys" and irregular, discontinuous masses.	More than 200 m of adits and winzes. Production in 1930-39; amount unknown.
331	Ruby (Blue Bell, Curly Jim, Montezuma).	SW1/4 sec. 9, T. 29 S., R. 31 E.	Three gold-bearing quartz veins, 1 to 3 m thick, cutting across contact between quartz diorite and schist.	30-m-long shaft with drifts and stopes, and an older shaft with an undetermined length of workings. Production of more than 15 kg Au.
332	San Antonio--------------	NW1/4 sec. 23, T. 29 S., R. 36 E.	Quartz vein, 1/3 m thick, in crushed quartz monzonite.	50-m-long shaft and other shallow shafts and open cuts; some drifting from shaft. Small production in 1908-37.
343	Sky Line-----------------	Ctr. sec. 8, T. 31 S., R. 36 E.	Quartz vein, 1 to 3 m thick, in granodiorite, containing auriferous pyrite and other sulfides.	Three shafts, 30 to 77 m deep, with several levels of unknown extent. Production in 1937-38 of probably less than $25,000 in gold.
352	St. John-----------------	Sec. 33, T. 28 S., R. 35 E., and NE1/4 sec. 4, T. 29 S., R. 35 E.	Quartz "vein," as much as 1 m thick, in Mesozoic granitic rock; a quartz-filled gouge zone containing free gold, auriferous pyrite, stibnite, arsenopyrite, and galena.	Developed from 1860 to 1875 by several hundred meters of shafts and workings; intermittent operation until 1950. Production until 1875 of $700,000 in gold and silver, and in later operations of only a few tens of kilograms of gold.
383	Zenda--------------------	SW1/4 sec. 29, T. 30 S., R. 33 E.	Quartz vein, 9 to 15 m thick, at contact between quartz diorite and quartz porphyry.	Long crosscut and glory hole, with some other level workings. Production in 1909-58 of about $650,000 in gold.

TABLE 38.—*Metallic-mineral deposits in the southernmost Sierra Nevada (south of lat 35°30′N.)*—Continued

Map No.	Mine name	Location	Mode of occurrence	Development and production
			Tungsten	
551	Basin View	Sec. 31(?), T. 28 S., R. 33 E.	Scheelite in 1-m-wide tactite zone at contact between limestone and quartz diorite.	Shallow shaft and trenches. Minor production in 1944-45.
557	Blue Point	Cor. secs. 10, 11, 14, 15, T. 30 S., R. 36 E.	Crystals of wolframite and scheelite in small lenses in brecciated and sericitized granitic rock.	Small open cuts, shafts, and trenches. Minor production of tungsten ore and mineral specimens.
567	Donlevy (Orrell group)	Ctr. S1/2 sec. 14, T. 29 S., R. 34 E.	Two quartz veins, 1 to 2 m thick, occupying shear zones in granodiorite and containing scheelite and oxidized pyrite.	Two adits, 87 and 130 m long. Production of 11 t of ore in 1944 from kidneys containing 50 weight percent WO_3.
577	Good Hope (Tungsten Chief).	Ctr. S1/2 sec. 27, T. 28 S., R. 32 E.	Scheelite-bearing tactite in metasedimentary rock near a contact with quartz diorite.	Includes three separate mines (Good Hope, First Landing, and Rocky Point) a few hundred meters apart. Various modest-size shafts and level workings. Total production of about 4,300 t of ore that probably averaged no more than 1 weight percent WO_3.
582	High-Low	NE1/4NW1/4 sec. 29, T. 30 S., R. 36 E.	Scheelite in fractures in granodiorite	20-m-long shaft, with other shallow workings and pits. Minor production in 1954.
606	Minnehaha (Claude, Mayflower).	NE1/4 sec. 1, T. 31 S., R. 33 E.	Scheelite, free gold, and pyrite in quartz vein, as much as 1 m wide, blobby along contact between limestone and schist. Some scheelite crystals reportedly weighed as much as 45 kg.	Two adits, each about 100 m long, with other workings. Some production during World War I.
607	Moreland property	NW1/4 sec. 20, T. 28 S., R. 34 E.	Scheelite in tactite developed along bedding planes in metamorphic rock.	Shallow shaft, short drift, and some trenches. Probable small production in mid-1950's.
608	Mountain View	NW1/4SE1/4 sec. 26, T. 28 S., R. 32 E.	Scheelite-bearing tactite in silicified limestone in contact with quartz diorite.	Short adit and small open cut. Minor production in 1940-41.
609	Naja (Hobby, Naja Scheelite).	Ctr. E. side of sec. 36, T. 28 S., R. 33 E.	Scheelite disseminated in tactite and in small crystals in quartz veinlets cutting the tactite at contact between granitic rock and schist.	Shallow shaft and short underground workings. Small production in 1942, 1950, and 1955; one small pocket, 2 to 3 m long and less than 1 m thick, yielded ore grading 30 weight percent WO_3 worth $23,000 in 1950.
624	Summit Lime Co	Probably secs. 34, 35, T. 12 N., R. 15 W.	Podlike scheelite in quartz vein in granodiorite (one 275-kg pod of scheelite found).	Near an important limestone quarry; locality of vein not specified, nor are workings mentioned. Minor production in 1940-43.
			Antimony	
2	Amalia	Secs. 5, 8, T. 31 S., R. 36 E.	Stibnite-bearing quartz veins in granodiorite	Shallow shafts and open cuts. Some production during World Wars I and II.
3	Antimony Consolidated	Ctr. SW1/4 sec. 5, T. 31 S., R. 36 E.	Two parallel stibnite-bearing quartz veins in granodiorite, exposed for strike lengths of 100 and 200 m.	Several small shafts and trenches. Some production, amount unknown.
8	Jenette-Grant	NW1/4 sec. 18, T. 28 S., R. 34 E.	Stibnite-bearing quartz vein at contact between limestone and schist several tens of of meters north of gold camp.	21-m-long adit. Minor production in 1918 and 1943-44.
9	Maharg and Houghawott (Mahary and Houghawoth group).	Sec. 4(?), T. 10 N., R. 15 W.	Stibnite in vein at contact between limestone and "porphyritic rock."	10-m-long shaft. Minor production.
10	Mammoth Eureka	SE1/4 sec 33, T. 30 S., R. 34 E.	Quartz veins containing stibnite, "silver," pyrite, and arsenopyrite in a silicified zone, 300 m long by more than 100 m wide, in andesitic volcanic rocks.	120-m-long crosscut adit, as well as other adits and underground workings. Production during World War I, amount unknown.
12	San Emigdio (Antimony Peak, Bousby, Boushy, Padre).	Secs. 9-11, T. 9 N., R. 21 W.	Stibnite and antimony oxides in siliceous lenses along a poorly exposed shear zone in "quartz diorite," extending for more than 800 m.	Five adits as much as 200 m long. Production in 1882 to present of no more than 600 t of "metallic antimony" suggested by extent of workings.
13	Studhorse Canyon	NW1/4 sec. 4, R. 31 S., R. 33 E.	Stibnite and yellow antimony oxides in brecciated quartz vein in altered rhyolite dike.	24-m-long shaft and short drifts. Small production, amount unknown.
15	Top of the World	SE1/4 sec. 34, T. 30 S., R. 32 E.	Auriferous stibnite in kidney-shaped bodies in vein at contact between schist and quartz diorite; some scheelite present.	70 m of workings consisting of adit, short winze, and drift. Small production, amount unknown.
16	Wiggins	Sec. 8(?), T. 31 S., R. 33 E.	Stibnite in 1-m-thick vein in quartz diorite	23-m-long adit and open cuts. Production in 1918 of 62 t of ore.
			Tin	
546	Meeke (Hogan-Mallery, Meeke-Hogan).	SW1/4 sec. 25, T. 9 N., R. 18 W.	Several iron-rich tactite and gossan bodies at margin of marble in granitic rock. Cassiterite associated with scheelite, molybenite, tourmaline, and phlogopite in tactite replacing limestone. Later pyrite oxidized to form gossan.	Several shallow shafts and trenches; short level workings. Minor production in 1943-45.

TABLE 38.—*Metallic-mineral deposits in the southernmost Sierra Nevada (south of lat 35°30'N.)*—Continued

Map No.	Mine name	Location	Mode of occurrence	Development and production
			Silver	
514	Amalie (Amalia, Amelia; not to be confused with the Amalie Antimony Mine to the east).	NW1/4 sec. 22, T. 30. S., R. 33. E.	Three veins, 0.2 to more than 1 m thick, in rhyolite porphyry and schist.	Not described. Production included below with the Gold Peak and Cowboy mines.
516	Gold Peak and Cowboy (Zada and Old Cowboy).	SW1/4 and NE1/4 sec. 28, T. 30 S., R. 33 E.	Two parallel quartz veins, more than 1 m thick, in fractures or sheeting planes in rhyolite dikes that intrude metasedimentary rocks. Ore mineral are cerargyrite and bromargyrite (AgBr), with some free gold and ruby silver.	Cowboy vein: Five drifts, totaling about 500 m in length, on four ore bodies. Gold Peak vein: Four crosscut adits and other level workings, totaling several hundred meters in length, on two ore bodies. Production, mostly in 1901-6, from the Loraine district (mainly the Amalie, Gold, and Cowboy deposits) of more than $1,000,000 in silver.
			Zinc	
687	Blackhawk----------------	SW1/4 sec. 5, T. 31 S., R. 33 E.	Irregular replacement bodies in coarsely crystalline limestone between tactite and rhyolite. Various ore minerals, including sphalerite and galena.	Nearly 200 m of level workings accessible by three adits. Production in 1944-58 of about 16,000 kg Zn and 8,900 kg Pb from less than 300 t of ore.
688	Kelso (Condor, Cully, and Hoyes, Tejon Ranch).	SE1/2 sec. 24, T. 9 N., R. 18 W.	Sulfide replacement of limestone near granitic contact. Zinc mineralization accompanied by lesser copper, lead, silver, and iron (no minerals reported).	Three adits totaling a couple of hundred meters in length. Minor production in 1943 of several tens of tons of ore containing 17.5 weight percent Zn.
			Mercury	
451	Fickert-Durnal-----------	SE1/4SW1/4 sec. 26, T. 31 S., R. 32 E.	Brecciated zone in dike, containing cinnabar in seams and disseminated particles. Dike noted as "pegmatite," possibly volcanic(?).	Three short adits and a shallow shaft. Small production in 1917.
454	Walabu (Cuddeback, Walibu).	NE1/4 sec., 27, T. 31 S., R. 32 E.	Cinnabar as thin encrustations on fractures and filling small breccia "veins" near margins of altered rhyolite dike.	Six adits, from 12 to 100 m long, and 700 m of horizontal workings in an area 90 by 200 m. Production in 1916-20, 1929-31, and 1936-40 of about 1,300 flasks of mercury valued at about $150,000.
			Arsenic	
17	Contact------------------	SW1/4 sec. 10, T. 10 N., R. 15 W.	Arsenopyrite and pyrite in irregular lenses along contact of limestone schist pendant with quartz diorite.	15-m-long shaft with short drifts. Production in 1923-24 of about 50 t of ore containing as much as 40 weight percent As.

TABLE 39.—*Metallic-mineral occurrences in the Salinian block*

Map No. (fig. 80)	Commodity	Location	Mode of occurrence	Development and production	Reference
			Santa Lucia Range[1]		
1	Hg	Blue Rock fault zone, on the south slope of Long Ridge.	Cinnabar in silica carbonate rock associated serpentine.	?----------------------------------	Fiedler (1944).
2	Cu	Trampa Canyon, north of Cachagua Road.	Chalcopyrite and pyrite reported in garnet-rich tactite and gossan at granitic contact.	Pits and short adits----------------	Hart (1966).
3	Mo(?)	Near Jackhammer Spring, in SW1/4 sec. 13, T. 20 S., R. 4 E.	Molybdenite and a trace of powellite in garnet-rich tactite.	?----------------------------------	Do.
4	W	Sur River in Big Sur State Park.	"High-grade" scheelite-bearing boulder in streambed.	None--------------------------------	Salem Rice (in Hart, 1966, p. 113).
5	W	Uncertain; on the Sur River.	Scheelite concentrates panned from stream deposits.	None--------------------------------	O.P. Jenkins (in Hart, 1959, p. 113).
			La Panza Range		
6	Au	East slope of the La Panza Range; small streams that drain into the San Juan River.	Placer gold in stream deposits. The source of these placers has not yet been found (no gold-bearing quartz veins are known in the La Panza Range).	Production in 1840-1934 estimated at $100,000 in gold.	Franke (1935, p. 406-408, 420-423).
			Gabilan Range		
7	Ag	Uncertain; partly in W1/2 sec. 29, T 14 S., R. 4 E.	Argentiferous galena in small veins and disseminated in limestone.	Prospected in 1850; small adits and shafts now caved.	Trask (1854), Blake (1858), Hart (1966).
8	Mo	Westcott Ranch, 13 km E. of Soledad; probably in sec. 26 or 27, T. 17 S., R. 7 E.	Molybdenite in 80-cm-wide quartz vein in granitic rock.	15-m-long adit (1920)---------------	Hart (1966, p. 74).
9	Au	Uncertain; in Miners Gulch.	Gold reported in stream gravel; source not found.	Small prospects---------------------	Andrews (1936, p. 33), Hart (1966, p. 45).
10	Cu	Uncertain; north of Chualar Canyon.	Two mineralized zones, 3 to 4 m thick and more than 1.5 km long, in quartzite, limestone, and granitic rock. Assays reported to range from 1.67 to 19 weight percent Cu.	90-m-long adit and 27-m-long shaft, at least as old as 1879; no longer accessible(?).	Hart (1966, p. 122).
	Cu, As	Uncertain; on the Riley Ranch, at the head of Chualar Canyon, 19 km E. of Chualar.	Mineralized zone showing arsenopyrite, magnetite, azurite, and malachite at contact between limestone and granitic rock.	Open cut, short tunnel, and 9-m-long shaft.	Laizure (1925, p. 28), Hart (1966, p. 120).
[2]11	Barite	West of Fremont Peak, in SE1/4 sec. 34, T. 13 S., R. 4 E.	Barite as replacement pods and veins in silicified limestone and dolomite.	Three northwest-trending adits and small prospect pits. Production in 1916-20 of 2,700 t of ore valued at $31,000.	Bowen and Gray (1959, p. 39-40), Hart (1966, p. 28-29).
12	W	Fremont Peak, in S1/2 sec. 34, T. 13 S., R. 4 E.	Small patches of scheelite in garnet-epidote tactite.	None--------------------------------	Bowen and Gray (1959, p. 13).
13	Cu	Sec. 33, T. 16 S., R. 7 E.	Iron gossan, copper carbonate stains, and chalcopyrite in granitic rock.	Small adits and open cuts-----------	Wilson (1943, p. 266).
14	Au	Sec. 14, T. 17 S., R. 7 E.	"Gold-bearing material in Miocene rhyolite"-----	Two adits, 122 and 165 m long-------	Do.
			Point Reyes area		
[3]15	W	About 1 km W. of Sir Francis Drake Highway, on banks of stream that ends at Willow Point.	Scheelite in alluvium and limestone-------------	?----------------------------------	Ver Planck (1955, p. 265-266).
[3]16	W	About 1/2 km W. of junction of Sir Francis Drake Highway with Bear Valley Road.	Scheelite in limestone in stream valley---------	?----------------------------------	Do.
17	Au	Probably Avalis Beach, south of Tomales Point on Tomales Bay.	"Gold abounds but in small quantity"; presumably placer deposits, but no known source.	None--------------------------------	Trask (1856, p. 13).

[1]Many semiquantitative spectrographic analyses were made of altered rocks and stream sediment from the Ventana Wilderness area. However, nothing of ore grade was discovered, and no fluorescent tungsten minerals were seen (Pearson and others, 1967).
[2]Not a metallic-mineral deposit, but included here because it represents the only recorded production from a mineral deposit in the Salinian block. The only other production recorded is from the La Panza Range placer gold area, the lode source of which has not been found.
[3]Tactite is not mentioned in the published descriptions of these deposits.

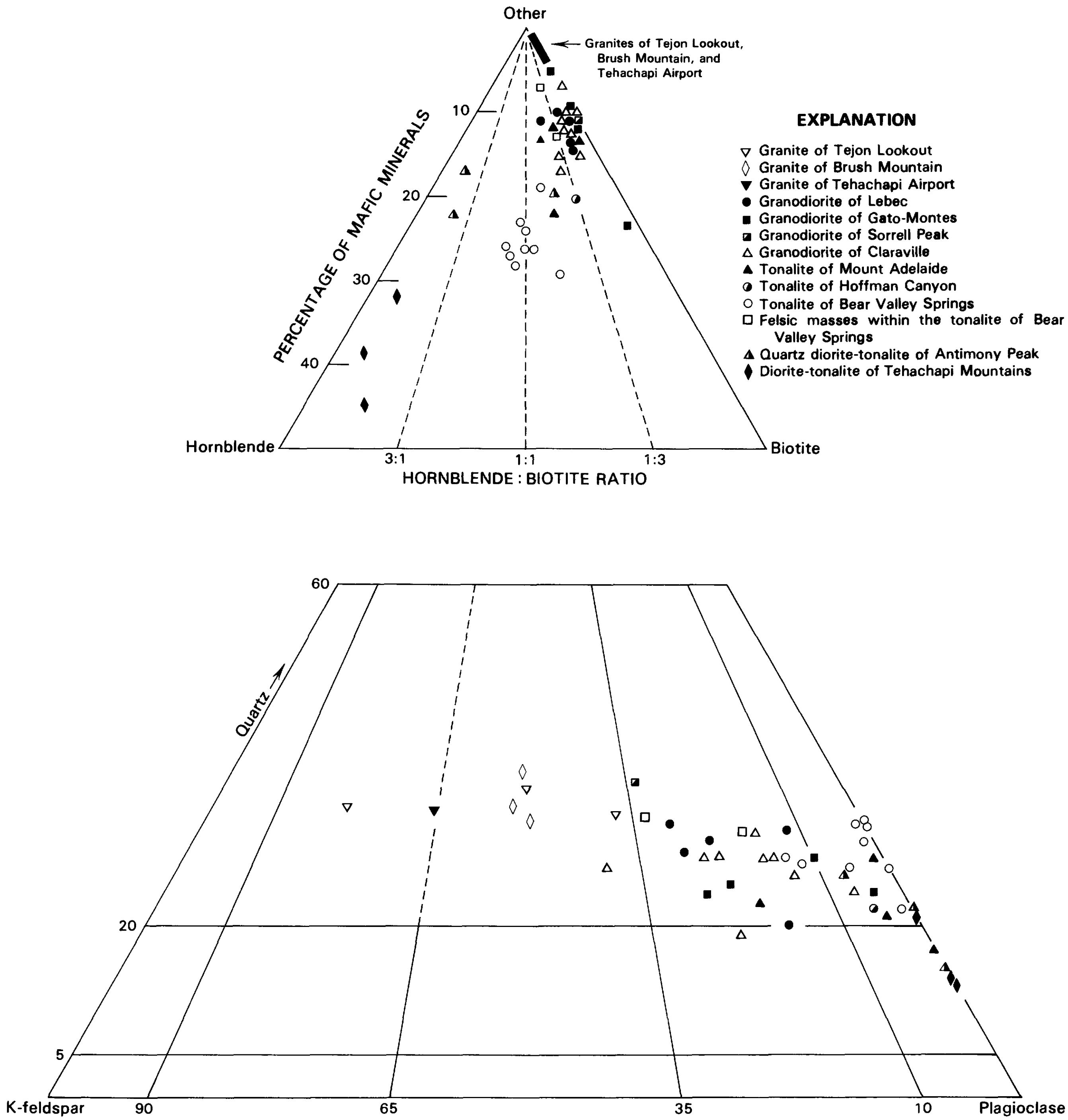


FIGURE 58.—Modal plots of chemically analyzed samples of plutonic rocks of the southernmost Sierra Nevada.

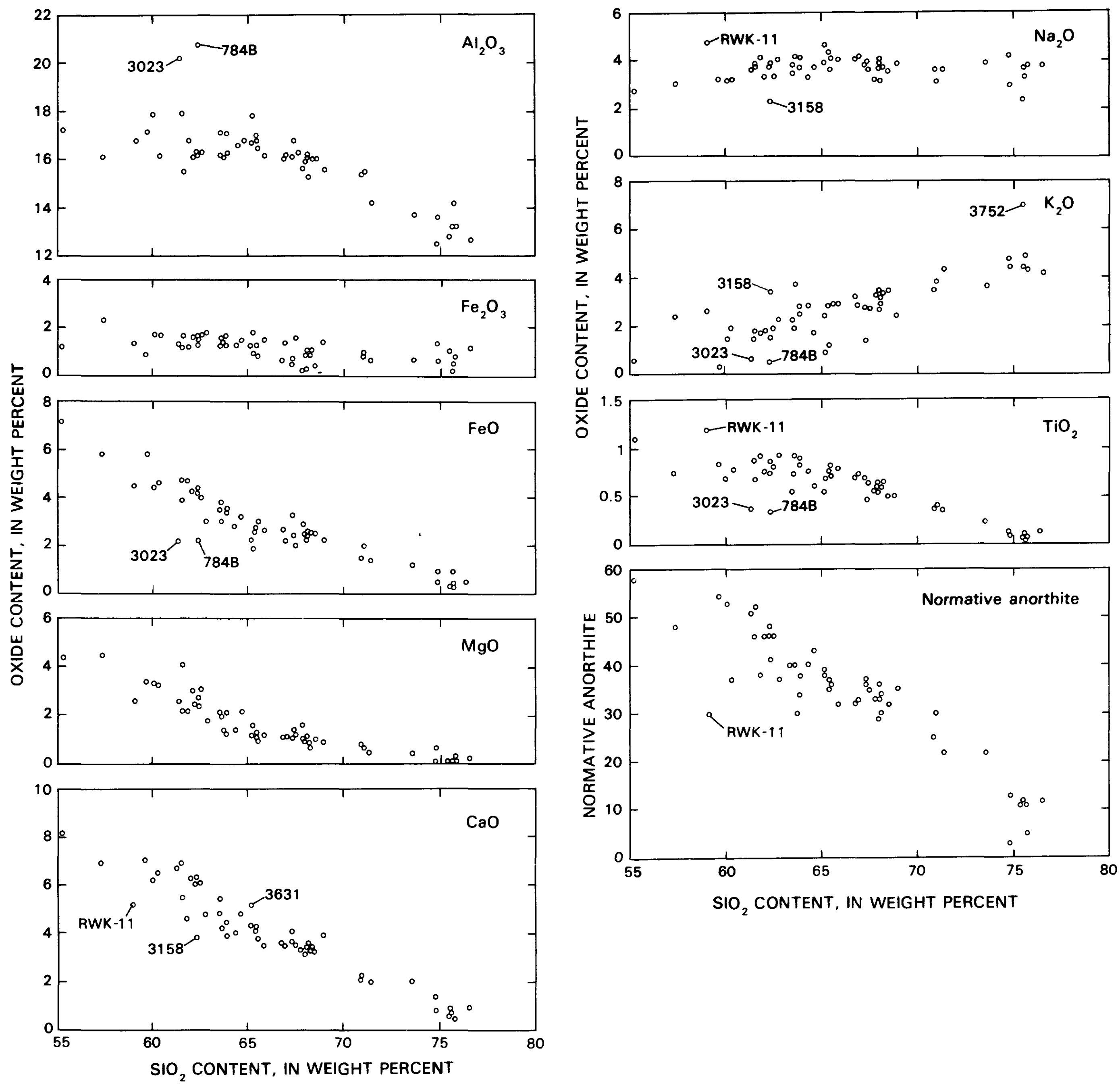


FIGURE 59.—Silica-variation diagrams. Numbered samples are discussed in text.

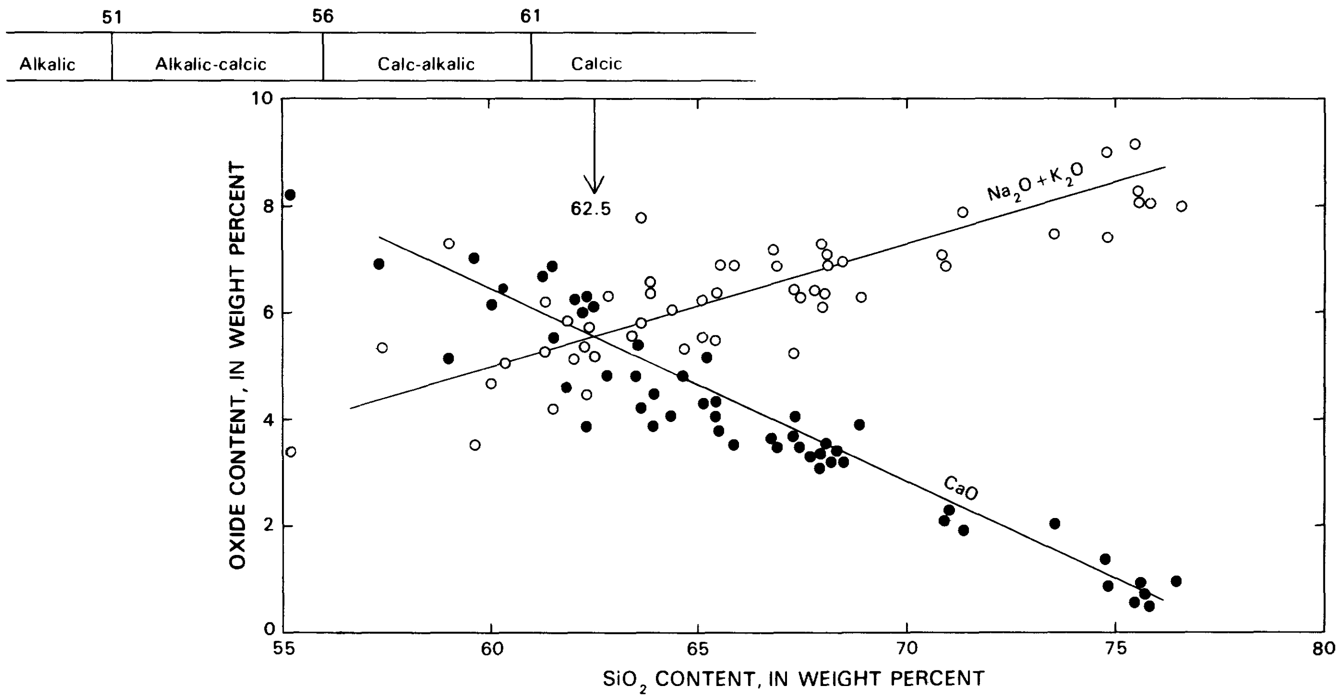


FIGURE 60.—Peacock index of chemically analyzed plutonic rocks of the southernmost Sierra Nevada.

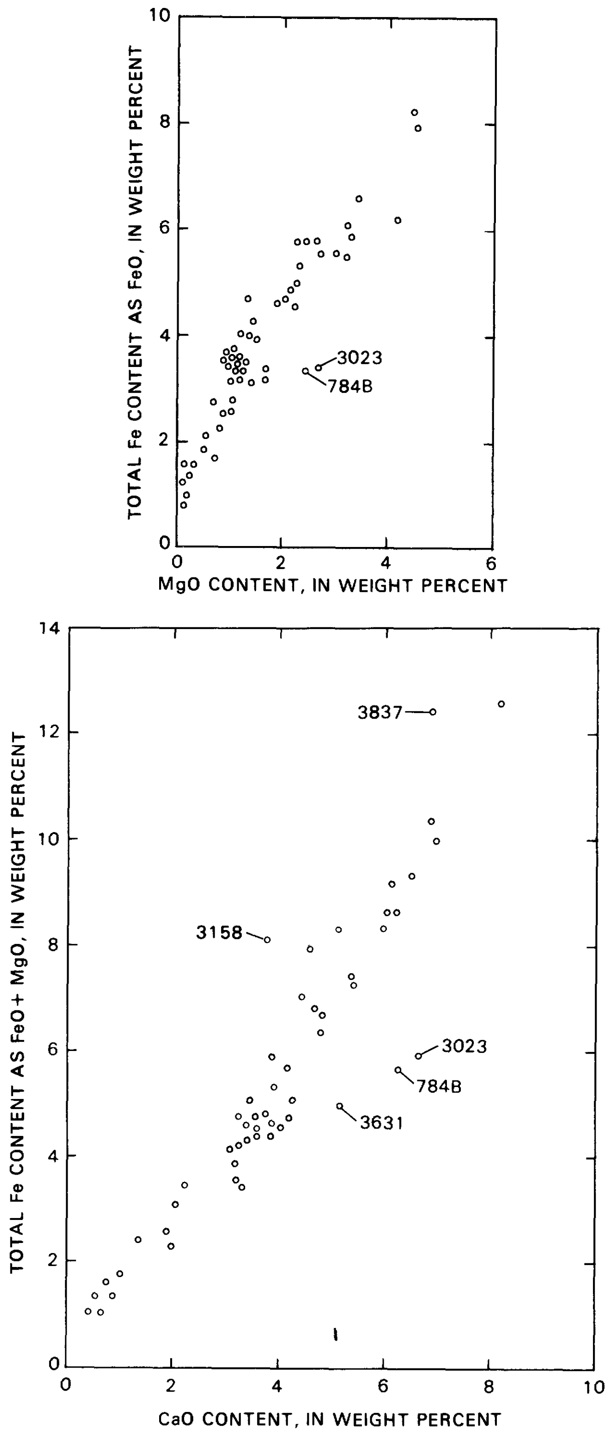


FIGURE 61.—Selected oxide ratios for chemically analyzed plutonic rocks of the southernmost Sierra Nevada. Total Fe (as FeO) is 0.9 Fe_2O_3 plus FeO. Numbered samples are discussed in text.

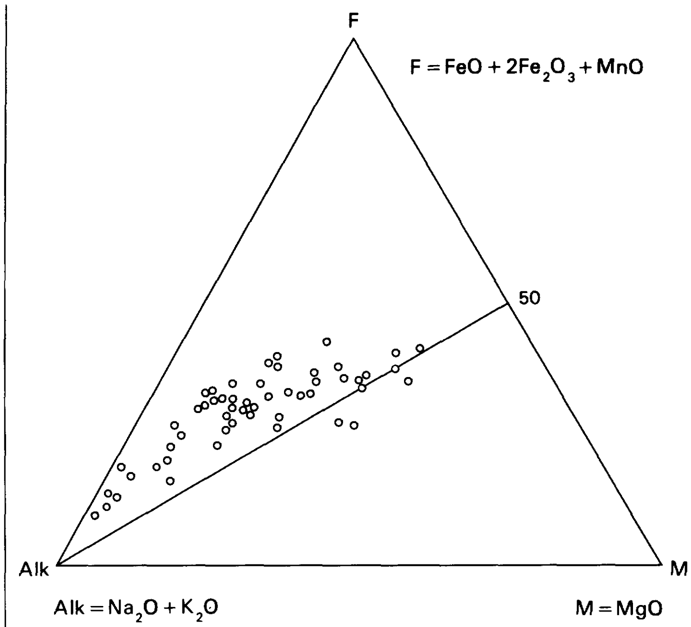


FIGURE 62.—Alk-F-M data for plutonic rocks of the southernmost Sierra Nevada.

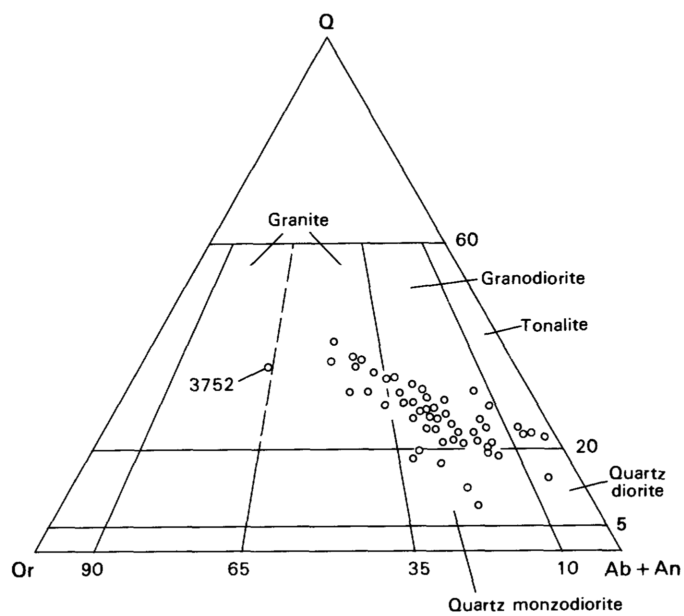


FIGURE 63.—Normative quartz (Q), orthoclase (Or), and albite plus anorthite (Ab+An) for plutonic rocks of the southernmost Sierra Nevada. Numbered sample is discussed in text.

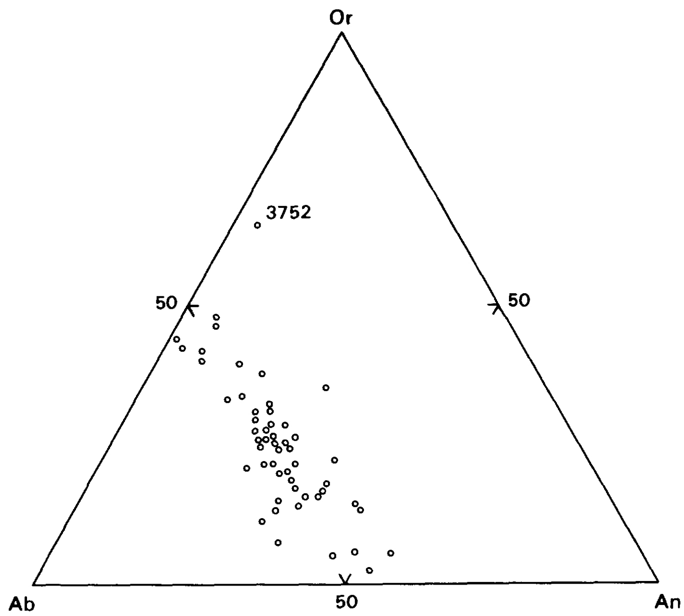


FIGURE 64.—Normative orthoclase (Or), albite (Ab), and anorthite (An) for plutonic rocks of the southernmost Sierra Nevada. Numbered sample is discussed in text.

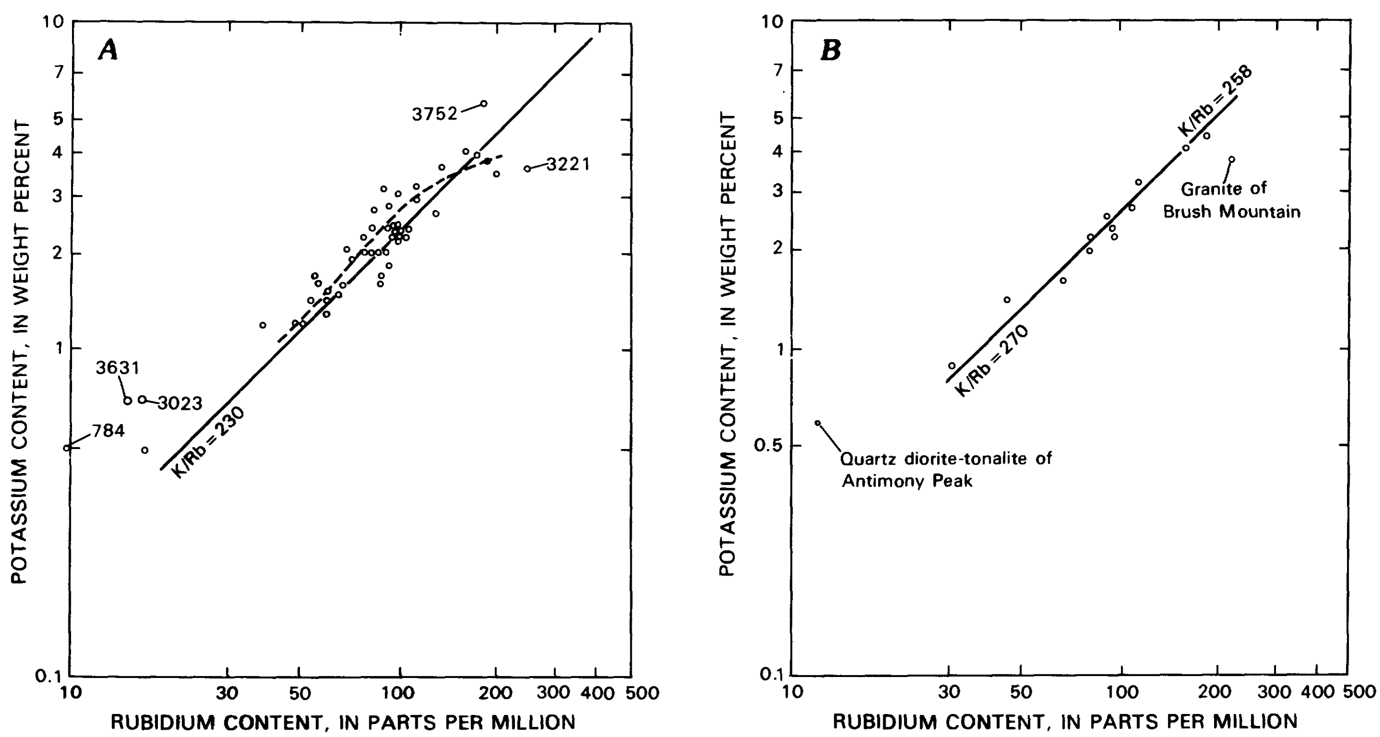


FIGURE 65.—Potassium versus rubidium content for samples of plutonic rocks of the southernmost Sierra Nevada. *A*, Values for individual rock samples. K/Rb-ratio line of 230 is the commonly accepted ratio for igneous rocks (Shaw, 1968). Dashed line is visual estimate of best fit for samples shown. Samples that plot well off main trend are identified by number (see table 35). *B*, Average values for some plutonic-rock units. Trendline is shown that best fits these data. Note that two plutonic-rock units plot well off the average trend.

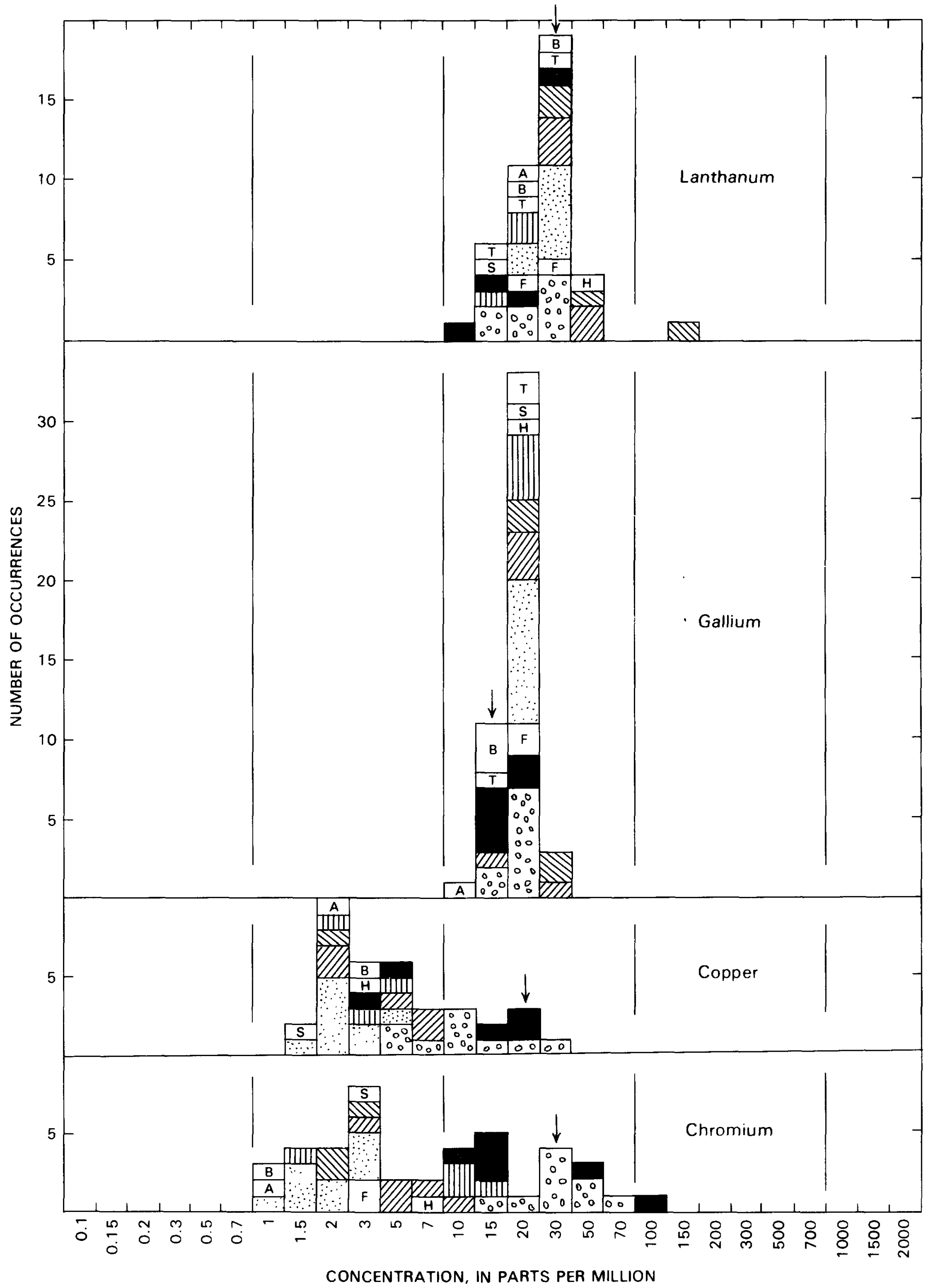


FIGURE 66.—Trace-element concentrations in samples of plutonic rocks of the southernmost Sierra Nevada. Data from emission spectrographic analyses. Arrows indicate average values for the central Sierra Nevada (Dodge, 1972). Concentration values indicate a range—for example, 70 includes all values from 70 to 99. See last part of figure for explanation.

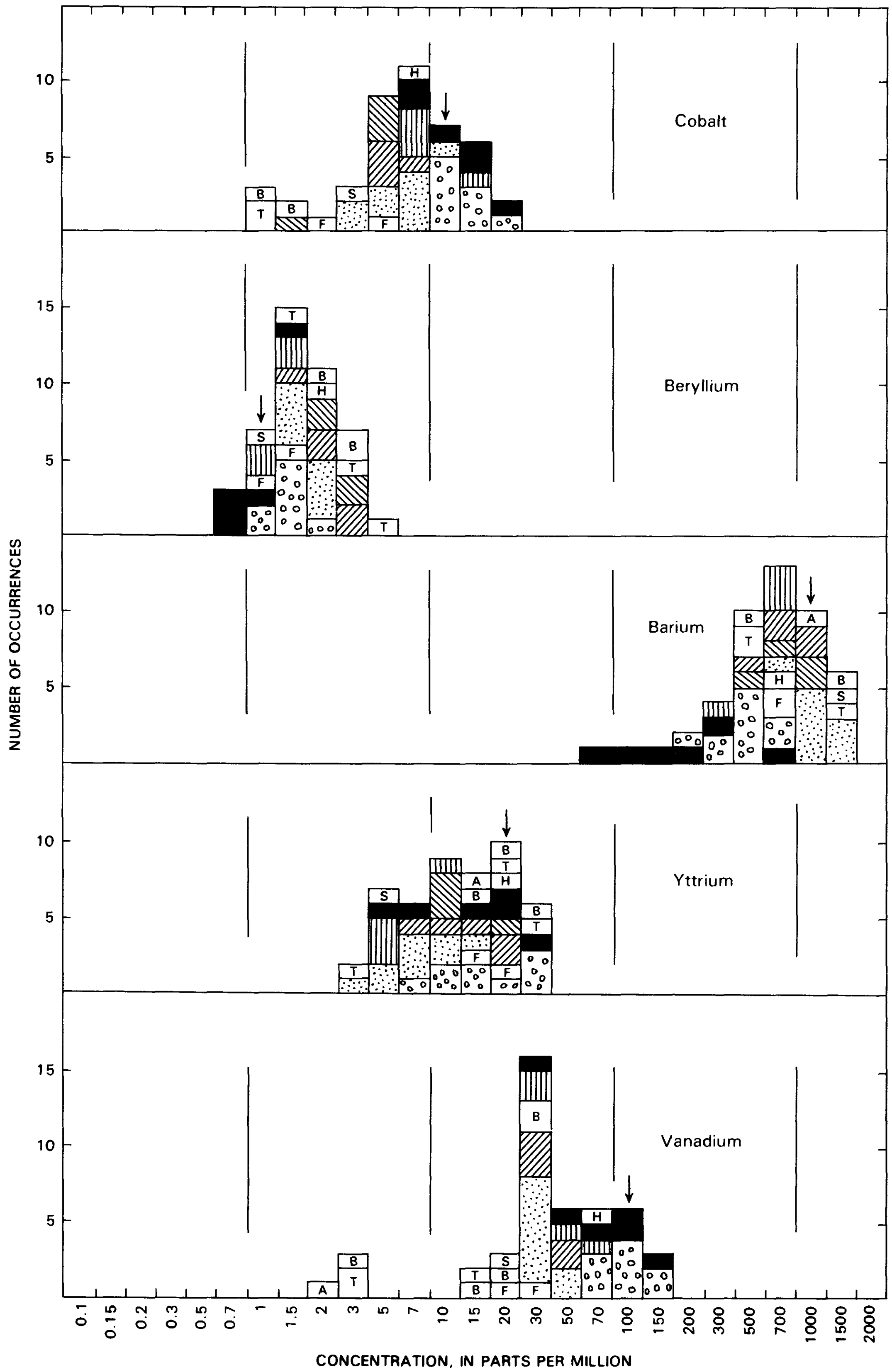


FIGURE 66.—Continued

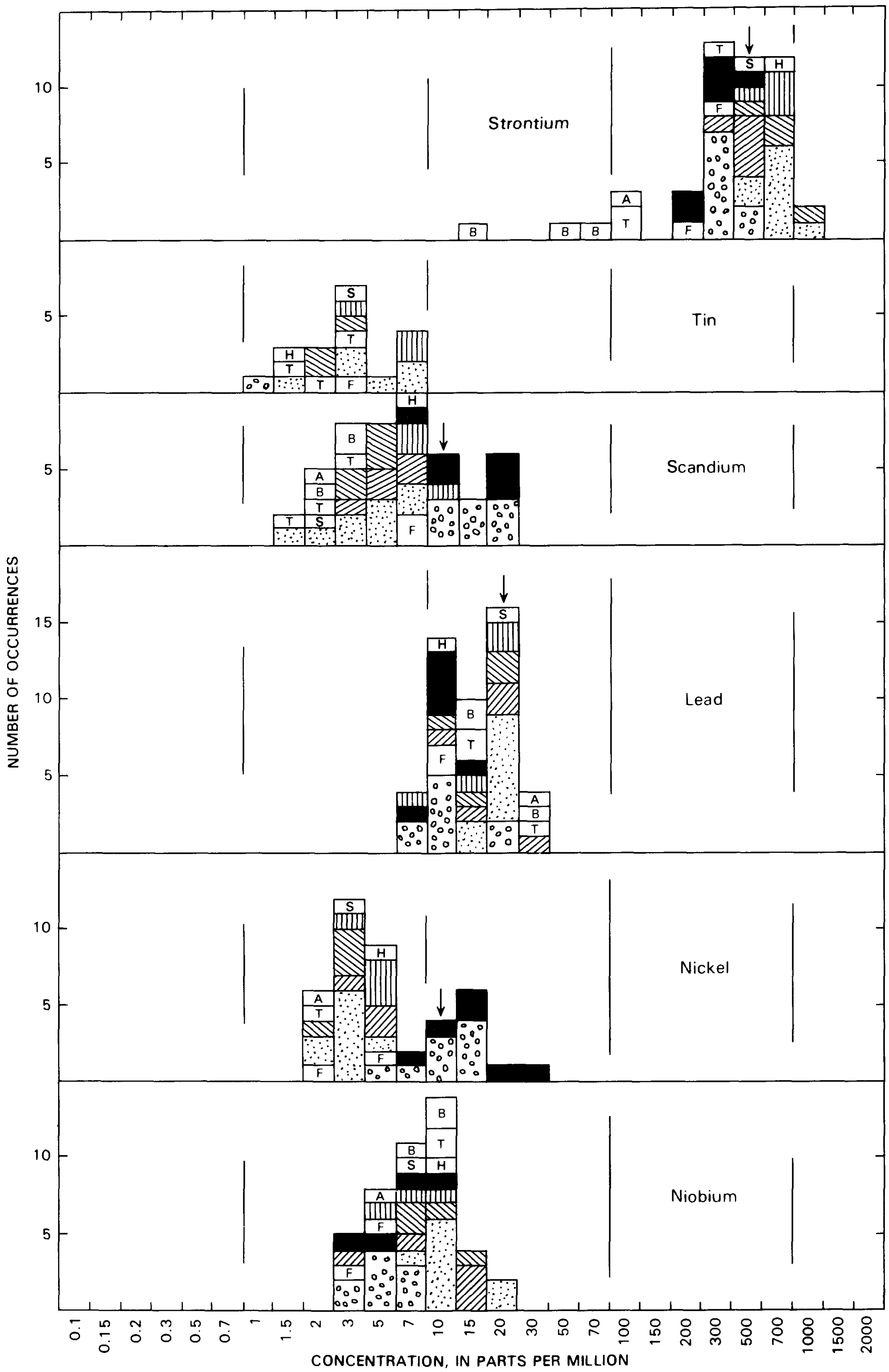
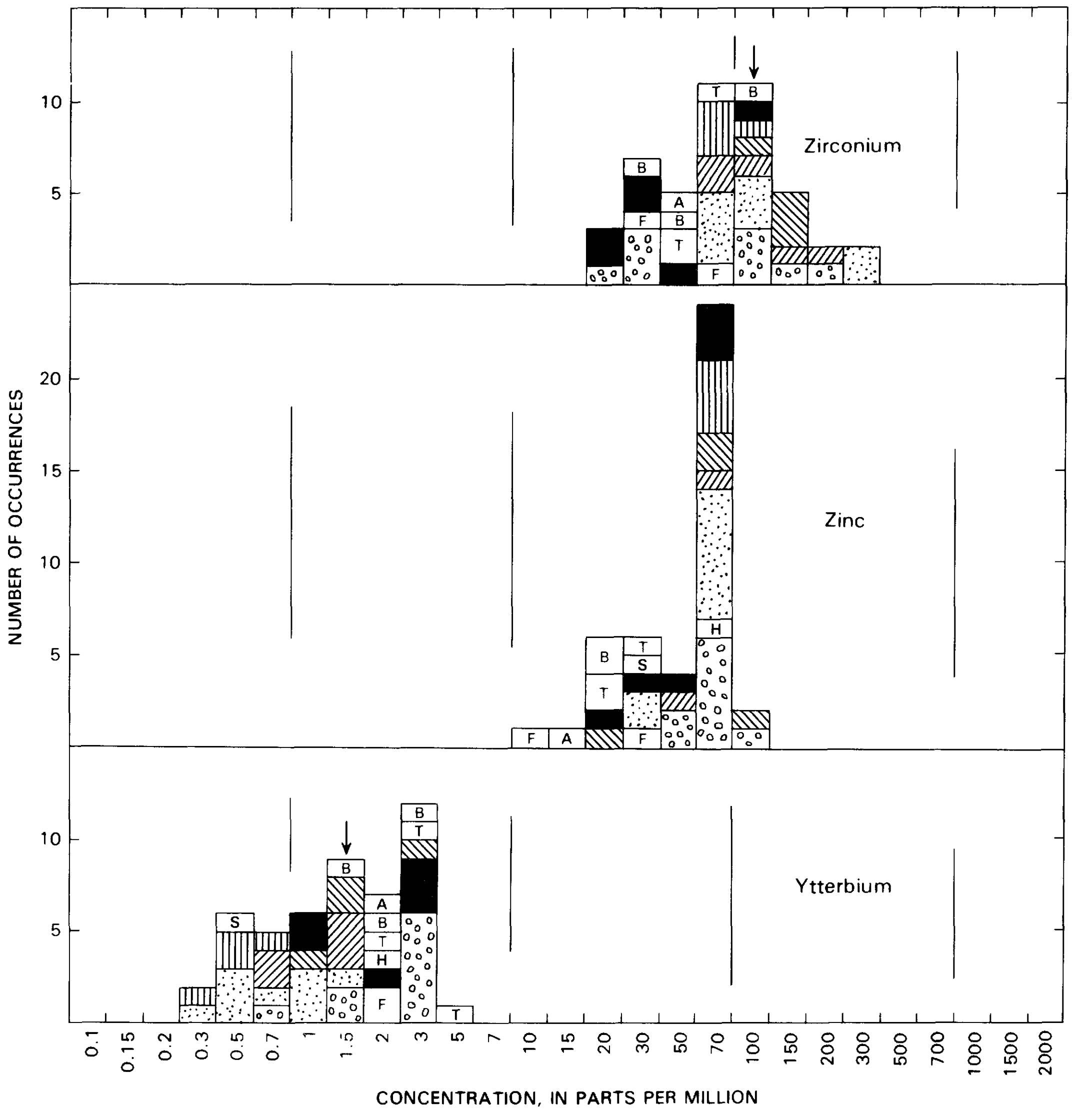


FIGURE 66.—Continued



EXPLANATION

Number of samples	Symbol	Unit	Number of samples	Symbol	Unit
1	A	Granite of Tehachapi Airport	4		Granodiorite of Gato-Montes
1	S	Granodiorite of Sorrell Peak	5		Granodiorite of Lebec
1	H	Tonalite of Hoffman Canyon	4		Tonalite of Mount Adelaide
2	F	Felsic masses within the tonalite of Bear Valley Springs	6		Diorite-tonalite of Tehachapi Mountains
3	B	Granite of Brush Mountain	9		Granodiorite of Claraville
3	T	Granite of Tejon Lookout	9		Tonalite of Bear Valley Springs

FIGURE 66.—Continued

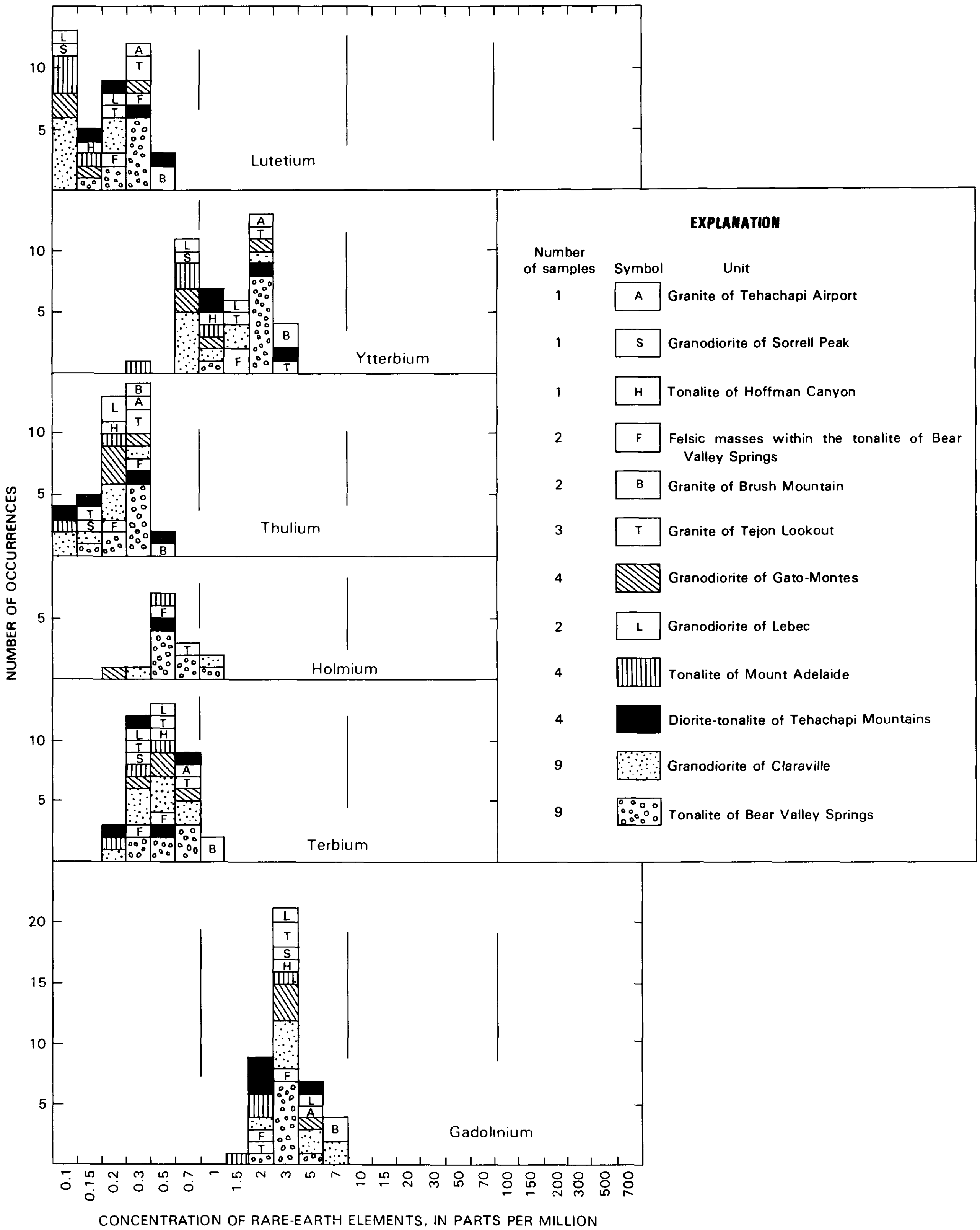


FIGURE 67.—Rare-earth- and selected other trace-element concentrations for samples of plutonic rocks of the southernmost Sierra Nevada. Data from instrumental neutron-activation analyses. Concentration values indicate a range—for example, 70 includes all values from 70 to 99.

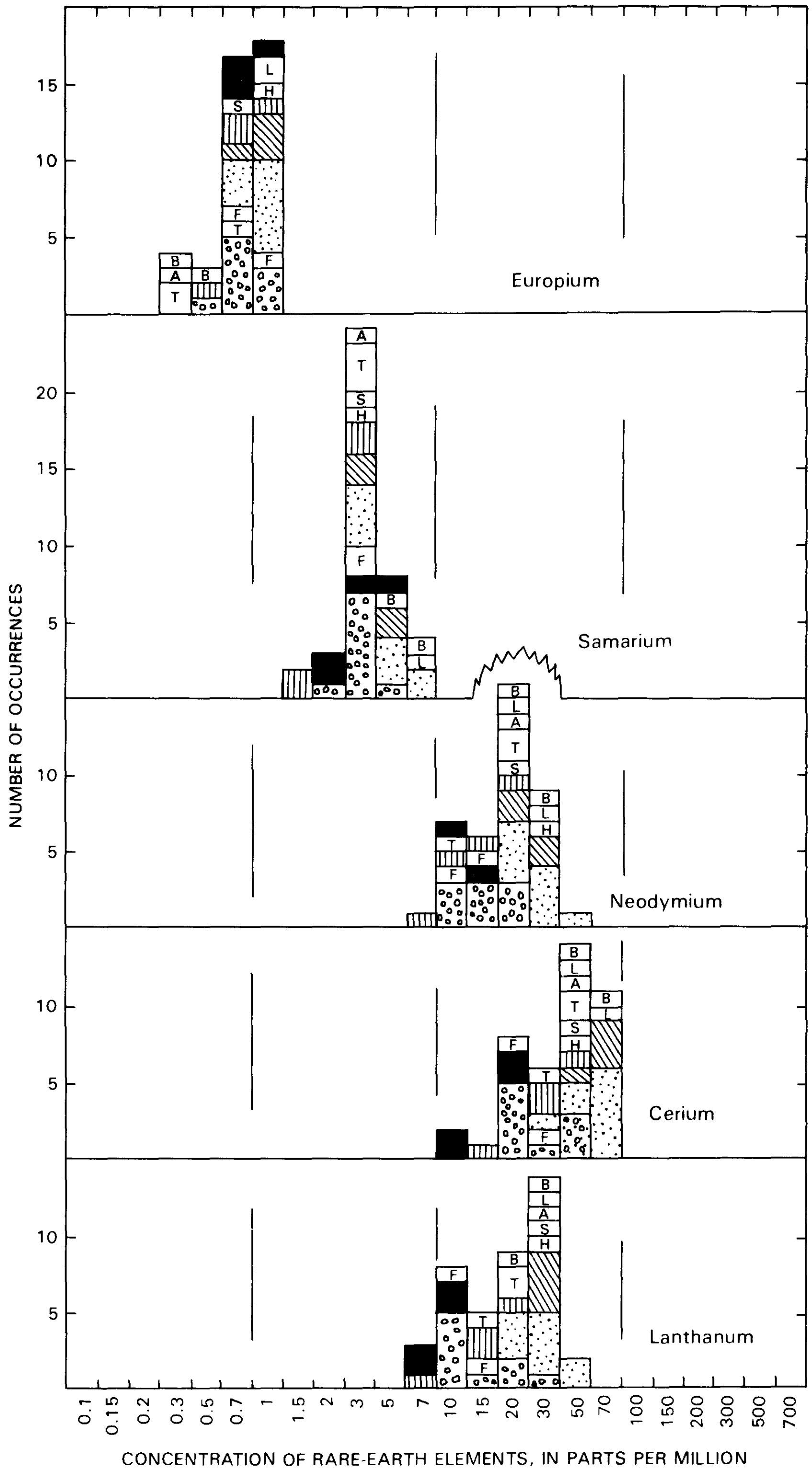


FIGURE 67.—Continued

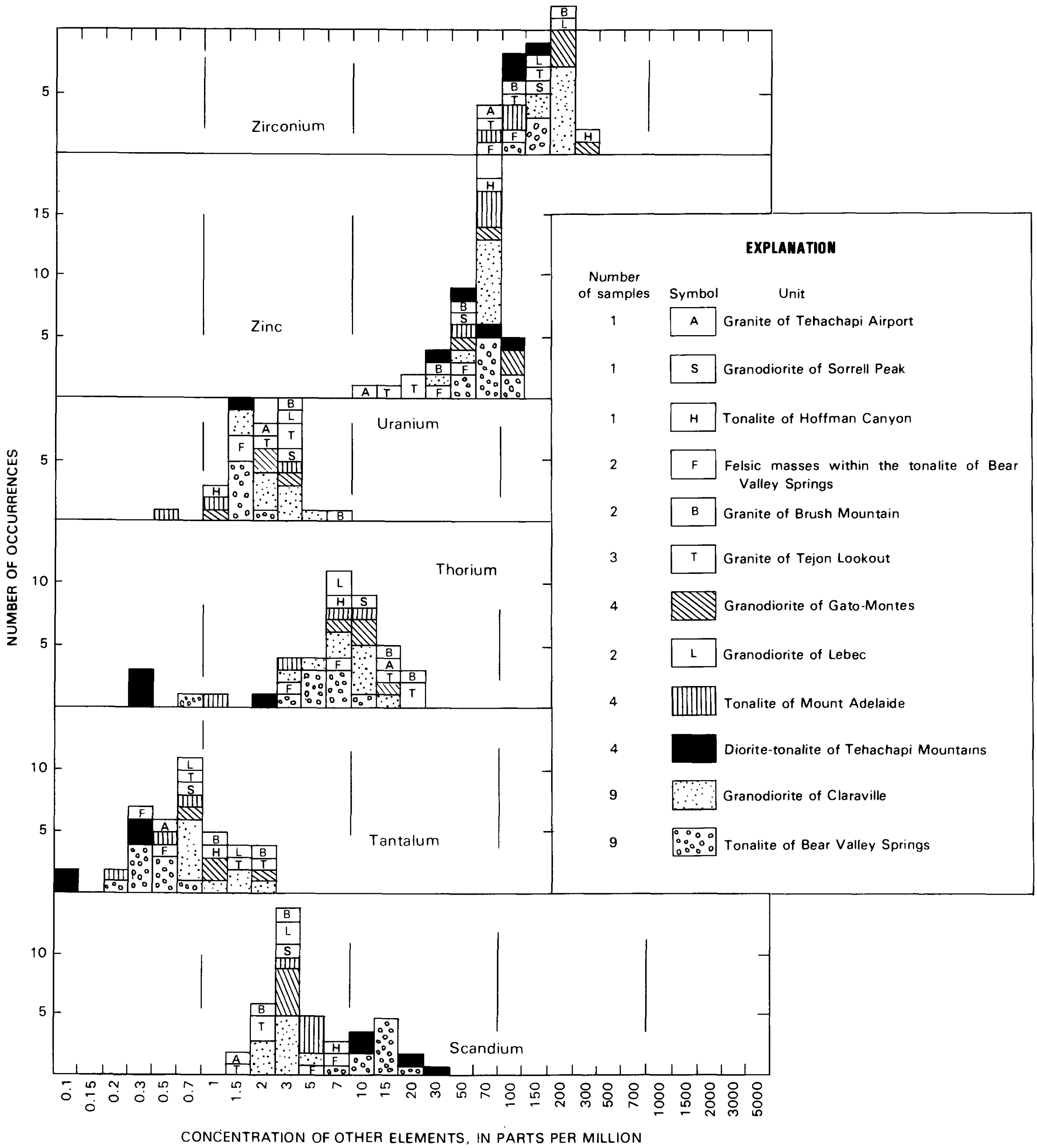


FIGURE 67.—Continued

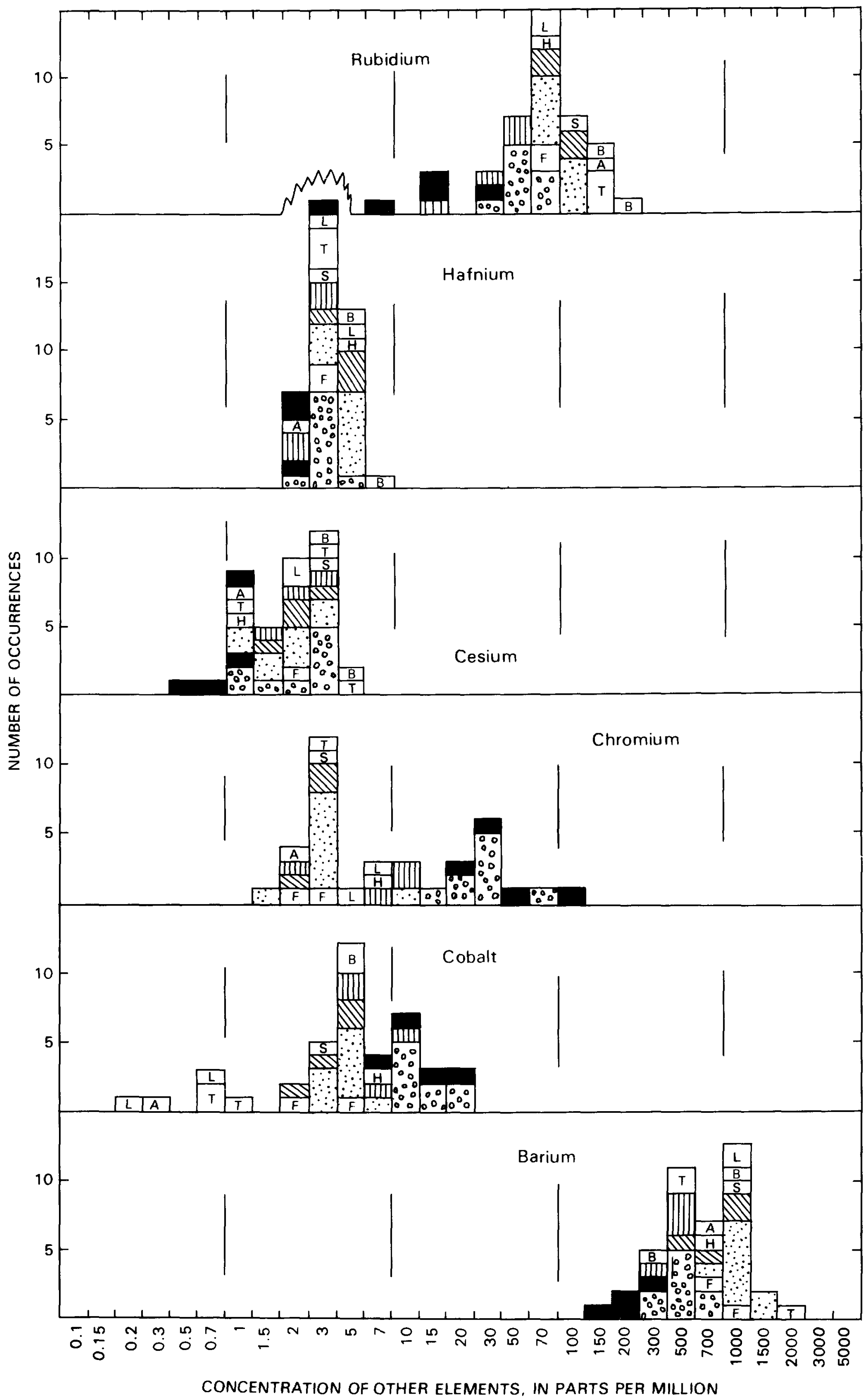


FIGURE 67.—Continued

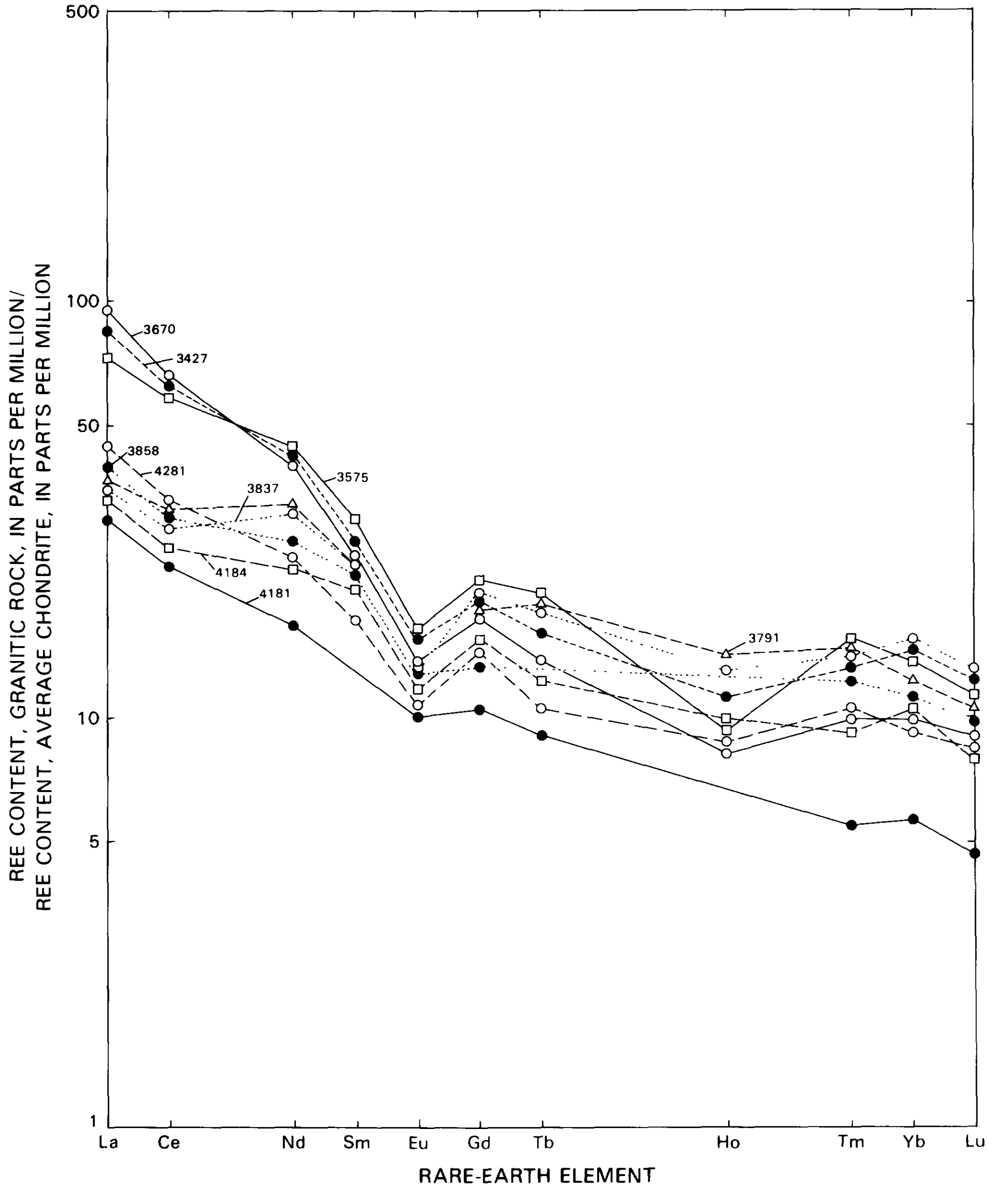


FIGURE 68.—Rare-earth-element abundances in samples of the tonalite of Bear Valley Springs relative to average chondrite.

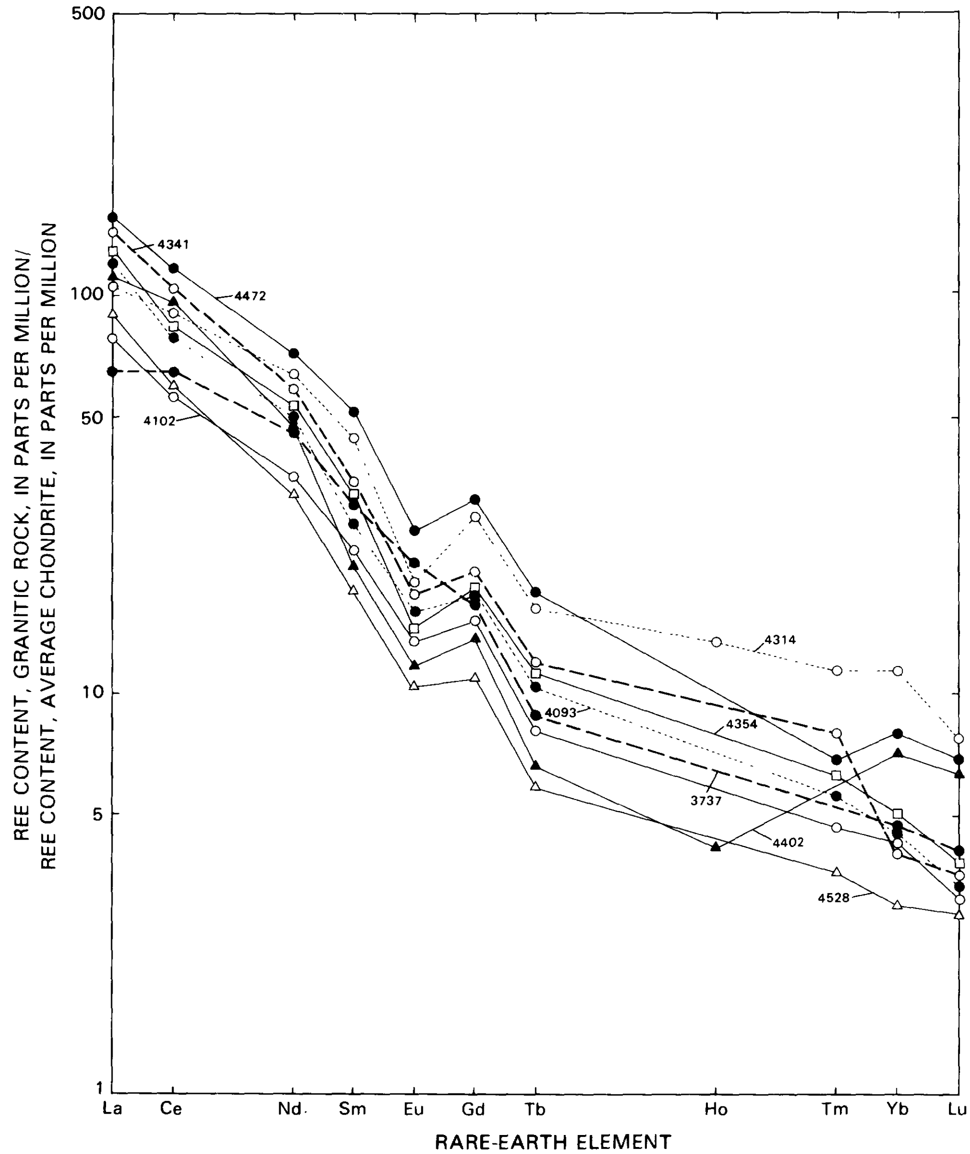


FIGURE 69.—Rare-earth-element abundances in samples of the granodiorite of Claraville relative to average chondrite.

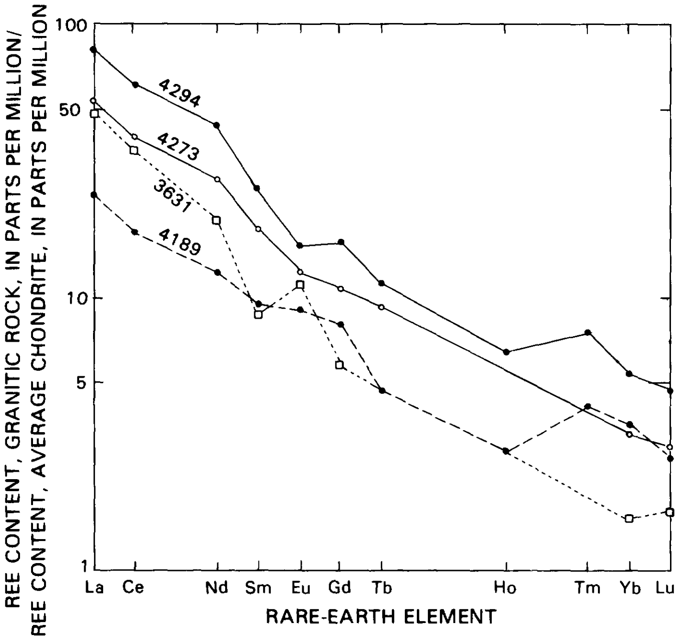


FIGURE 70.—Rare-earth-element abundances in samples of the tonalite of Mount Adelaide relative to average chondrite.

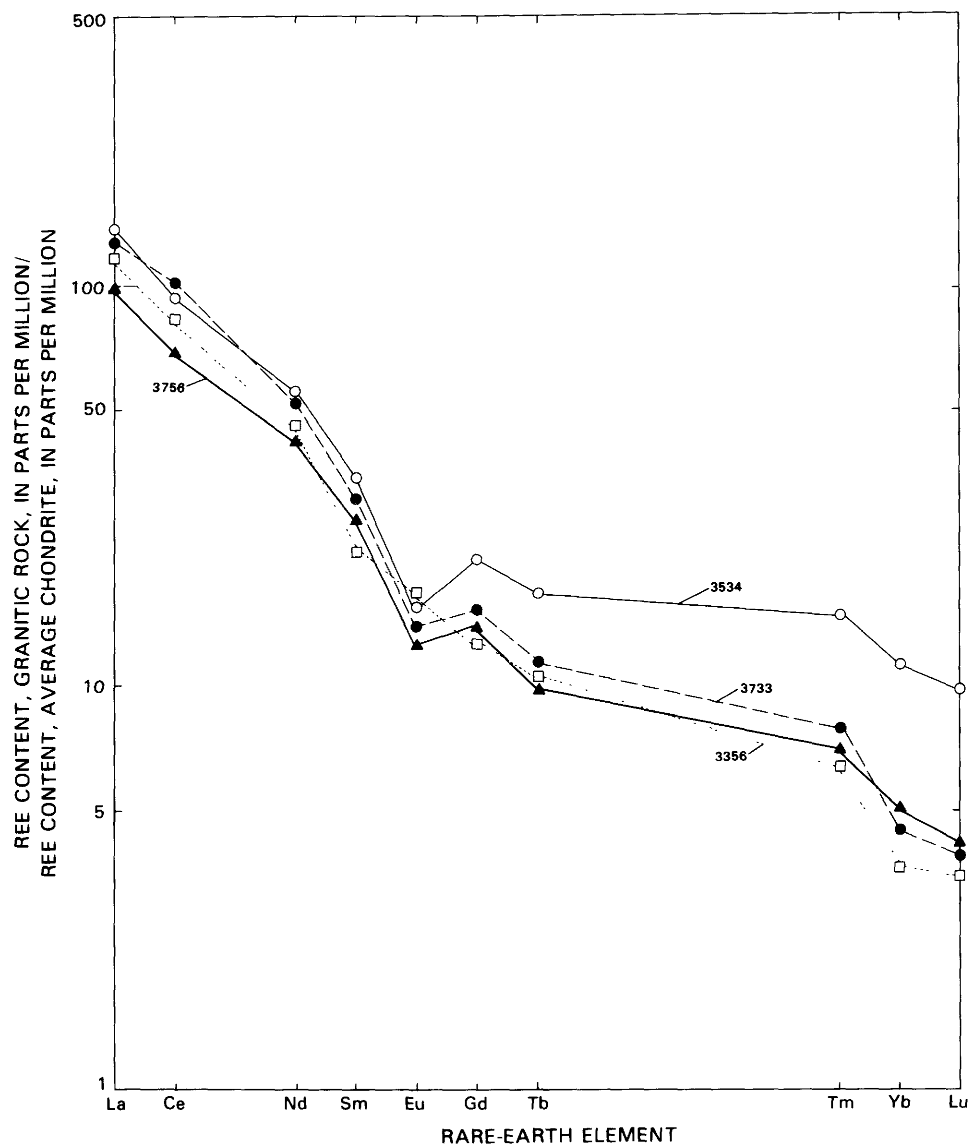


FIGURE 71.—Rare-earth-element abundances in samples of the granodiorite of Gato-Montes relative to average chondrite.

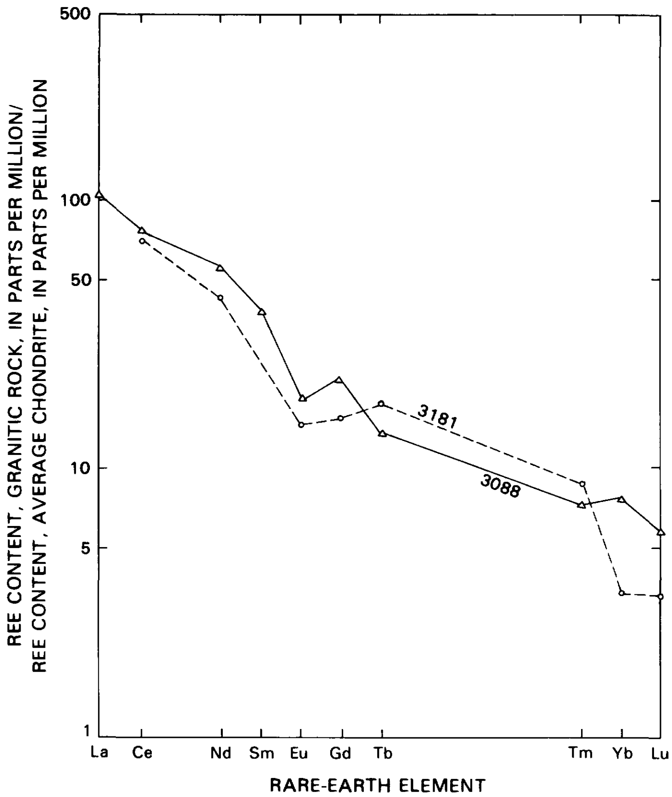


FIGURE 72.—Rare-earth-element abundances in samples of the granodiorite of Lebec relative to average chondrite.

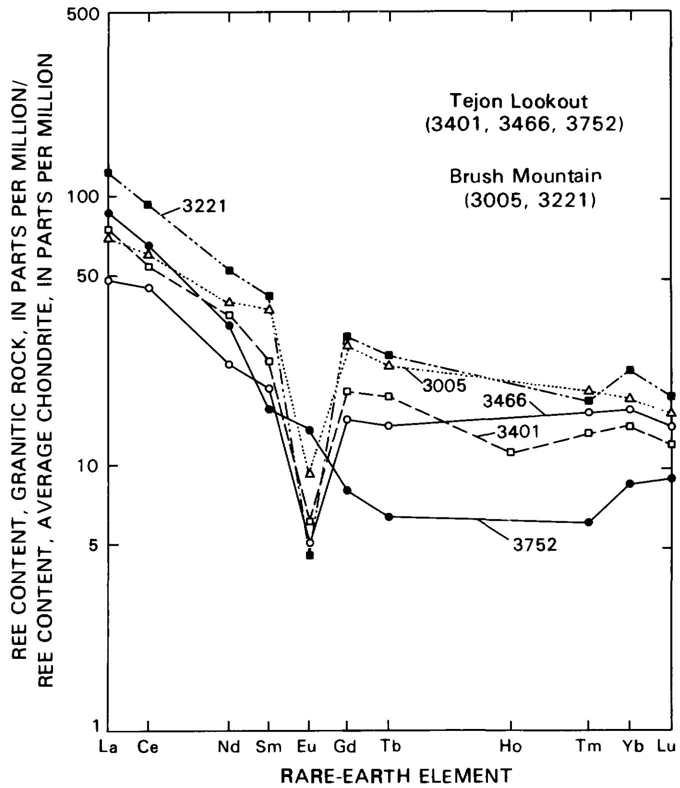


FIGURE 73.—Rare-earth-element abundances in samples of the granites of Tejon Lookout and Brush Mountain relative to average chondrite.

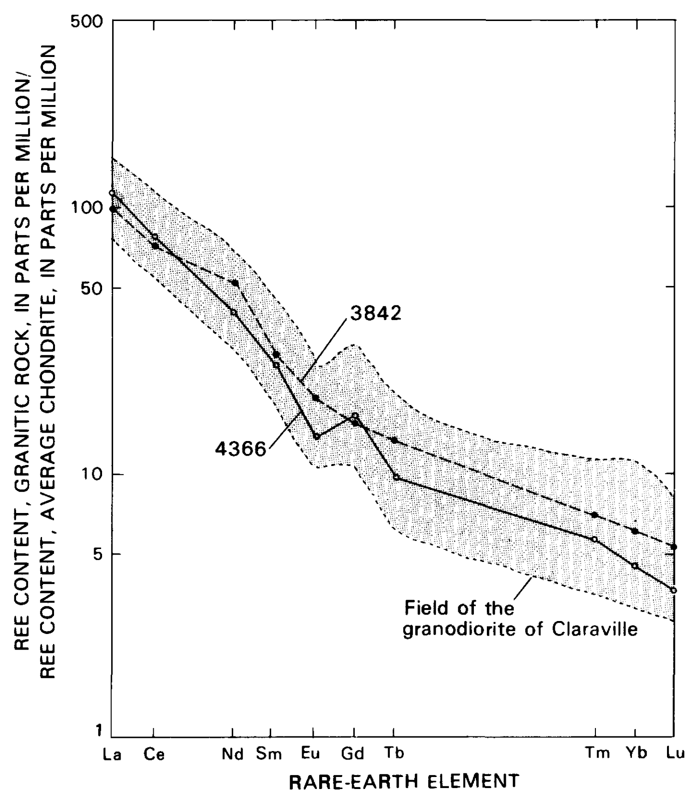


FIGURE 74.—Rare-earth-element abundances in samples of the tonalite of Hoffman Canyon (3842) and the granodiorite of Sorrell Peak (4366) relative to average chondrite. Field of the granodiorite of Claraville (fig. 69) is shown for comparison.

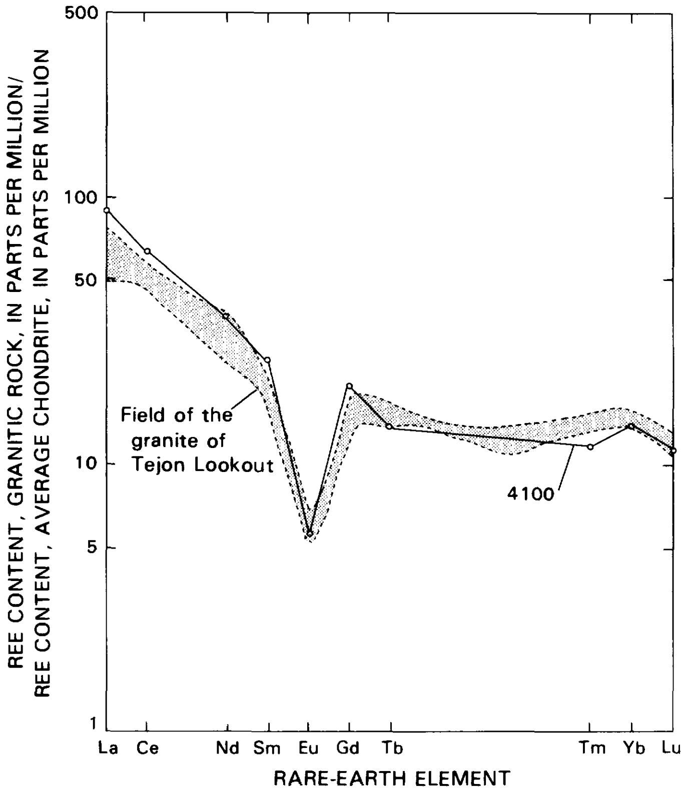


FIGURE 75.—Rare-earth-element abundances in sample of the granite of the Tehachapi Airport relative to average chondrite. Field of the granite of Tejon Lookout (fig. 73) is shown for comparison.

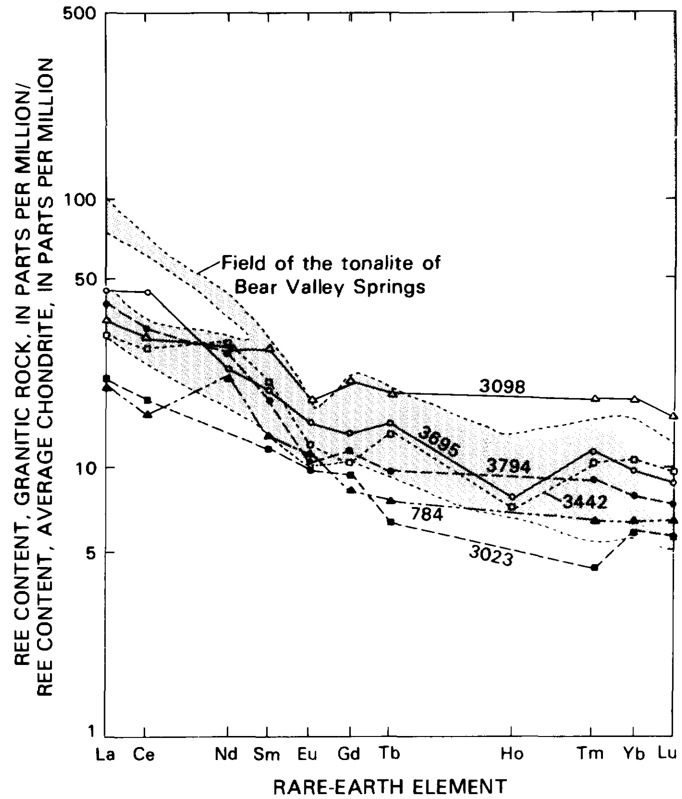
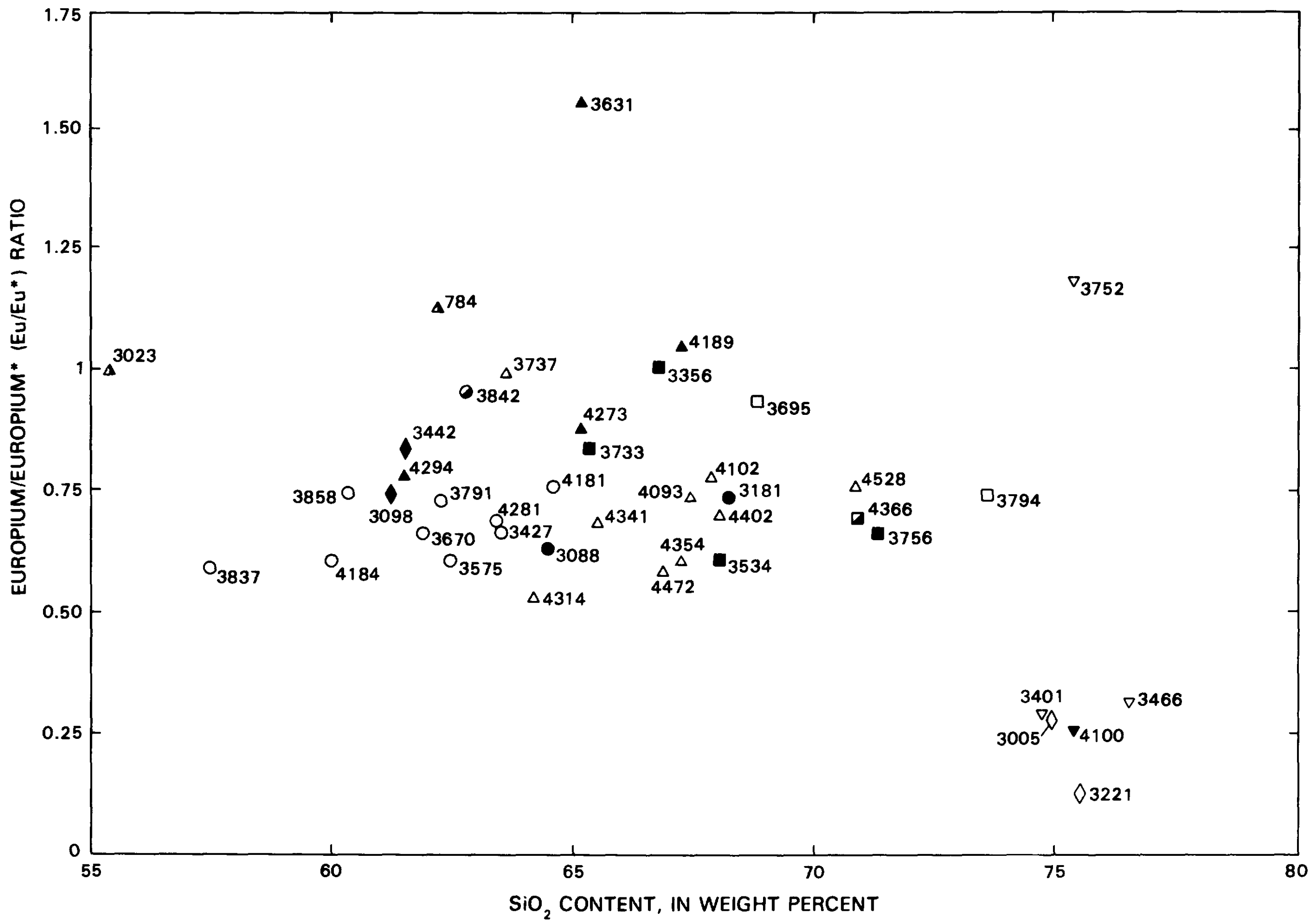
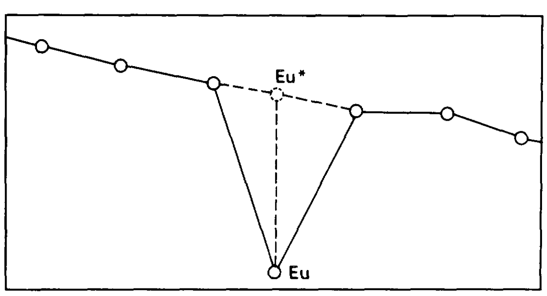


FIGURE 76.—Rare-earth-element abundances in samples of felsic masses within the tonalite of Bear Valley Springs (3695, 3794), the diorite to tonalite of the Tehachapi Mountains (3098, 3442), and the quartz diorite to tonalite of Antimony Peak (784, 3023) relative to average chondrite. Field of the tonalite of Bear Valley Springs (fig. 68) is shown for comparison.





EXPLANATION OF EU*

Eu* is the hypothetical position of the europium value on an REE plot if there is no europium anomaly. The ratio of Eu to Eu* is the measure of the size of the europium anomaly.

FIGURE 77.—Europium/europium* (Eu/Eu*) ratio versus silica content for selected plutonic-rock samples. Symbols refer to specific plutonic-rock units (see fig. 6).

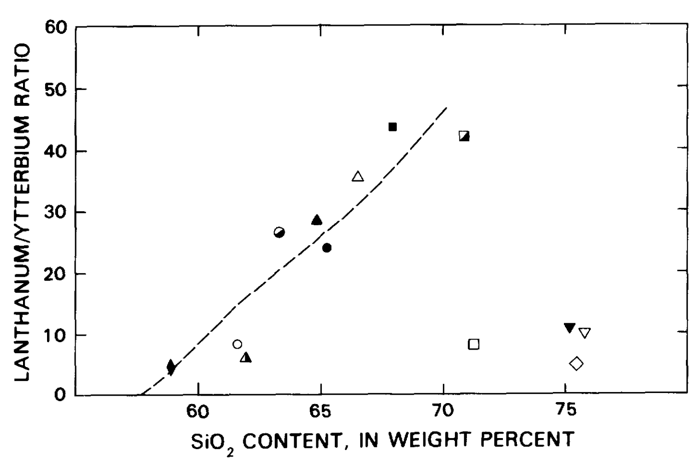


FIGURE 78.—Average La/Yb ratio versus silica content for some plutonic-rock units (see fig. 6 for explanation of symbols). Dashed trendline is a visual estimate.

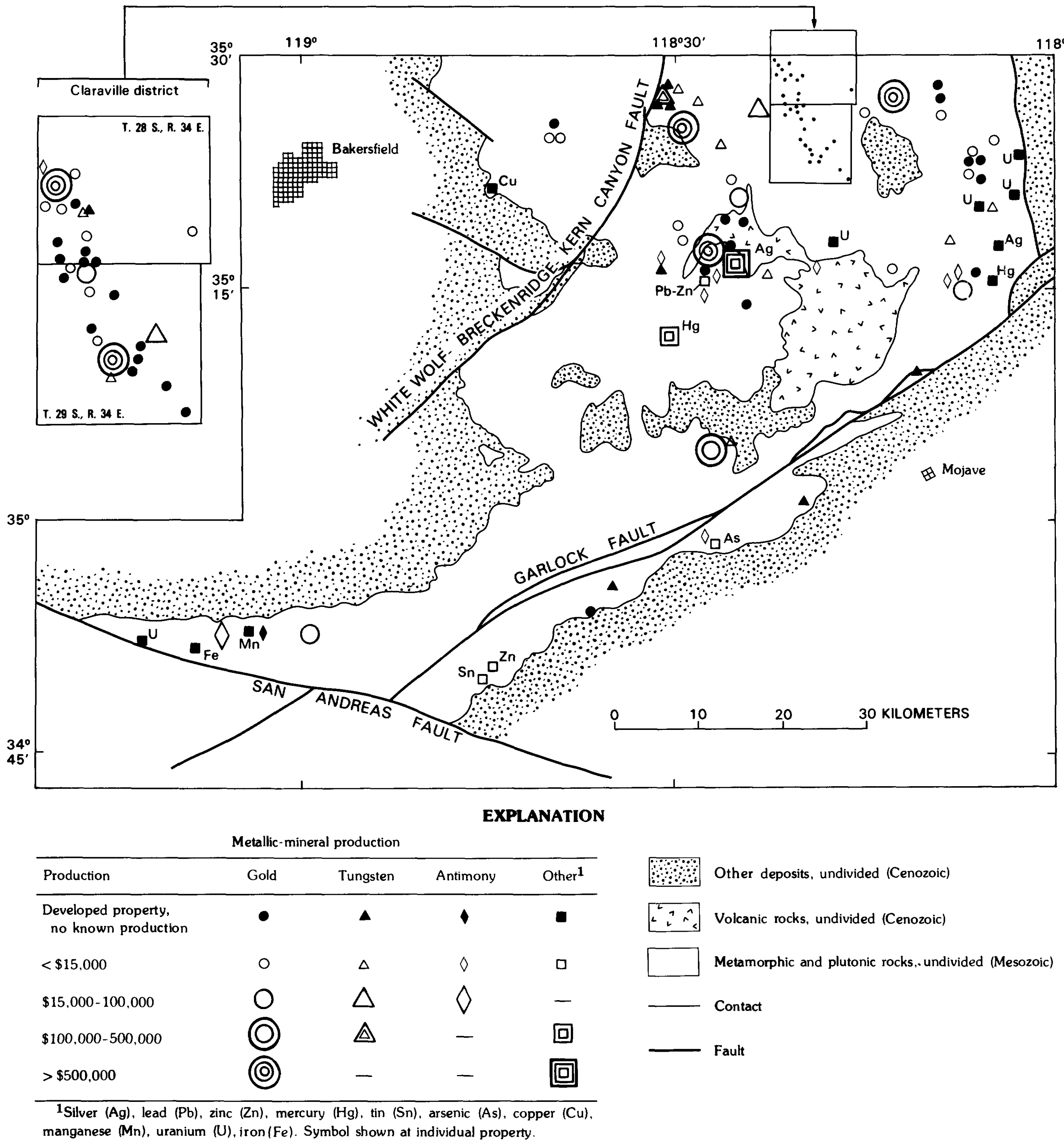


FIGURE 79.—Locations of metallic-mineral mines and developed prospects in the southernmost Sierra Nevada (see table 38).

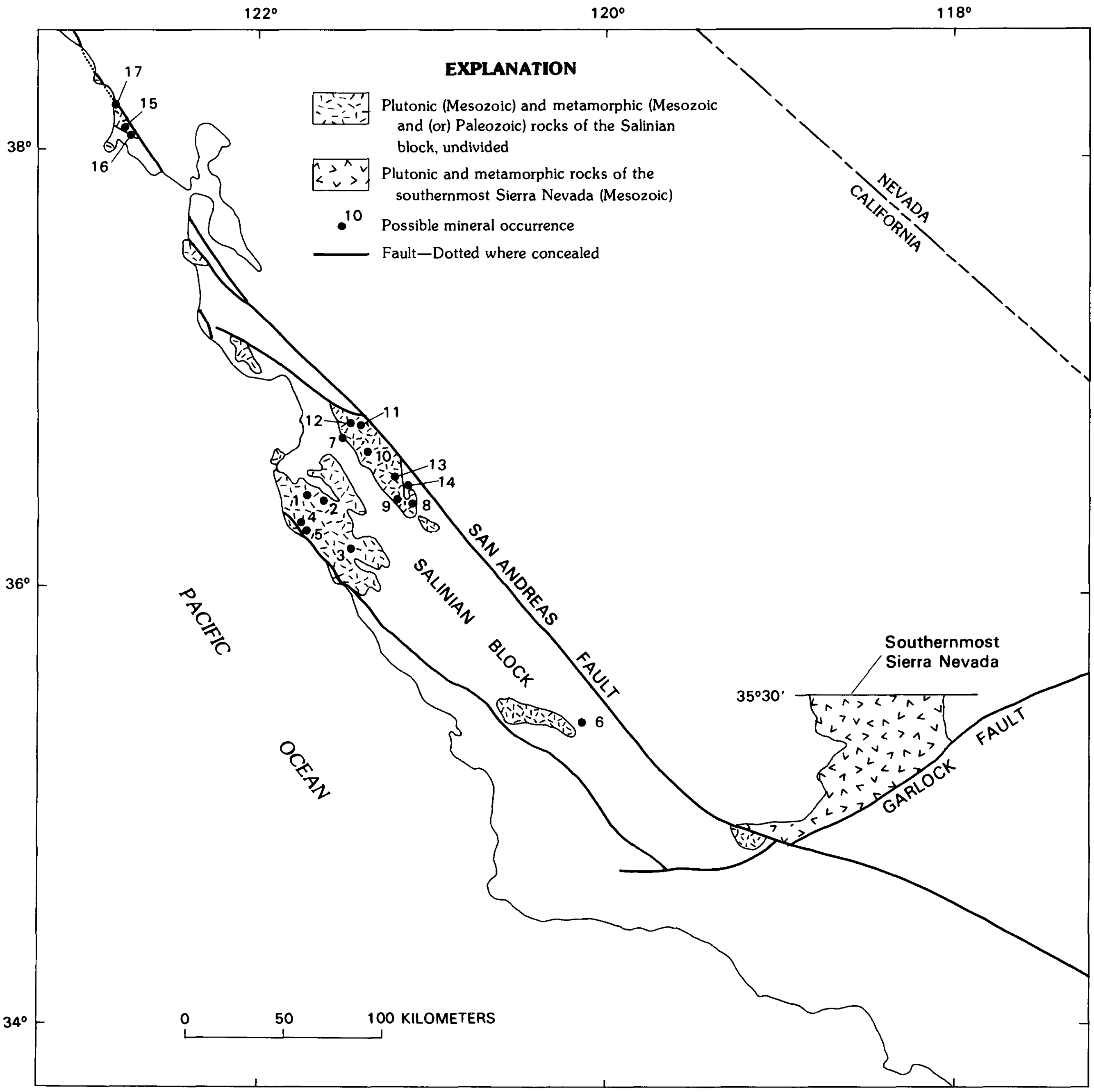


FIGURE 80.—Locations of possible metallic-mineral occurrences in the Salinian block (see table 39).