RARE EARTHS ABUNDANCES AND FRACTIONATIONS AND THEIR IMPLICATIONS FOR BATHOLITHIC PETROGENESIS IN THE PENINSULAR RANGES BATHOLITH, CALIFORNIA, USA, AND BAJA CALIFORNIA, MEXICO

Thesis by

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To a lovely lady, my wife Carol

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ABSTRACT

Rare Earth Element (REE) patterns of plutonic rocks across the Peninsular Ranges batholith vary systematically west to east, transverse to the long axis and structural trends of the batholith. Three major parallel elongate geographic regions are each defined by distinct REE pattern types. Rocks from the western region display slight light REE enrichment, flat heavy REE, and negative Eu anomalies. An abrupt transition to rocks with middle and heavy REE fractionated and depleted REE patterns with no or positive Eu anomalies occurs in the central region of the batholith. Further to the east a second transition to strongly light REE enriched rocks some of which have positive or negative Eu anomalies occurs. Some gabbros may show divergent patterns.

These large variations are observed even in similar lithologies across the three regions and notably in tonalites, the major rock type of the batholith. The slopes of the REE patterns within rocks of each region are largely independent of rock type, and no consistent variations in REE abundances and Eu anomalies with lithology are noted with the exception of some gabbros. Most of the leucogranodiorites of the western region have larger negative Eu anomalies than nearby tonalites. Granodioritic rocks of the central and eastern regions may have positive, negative, or no Eu anomalies.

These results are the first report of systematic variations in REE characteristics across a granitic batholith whose geologic setting at a convergent plate boundary has been established. Some similarities and contrasts to REE variations across modern volcanic arcs are noted. Along the westernmost margin of the batholith in northern Baja California, Mexico, leucotonalitic rocks of the San Telmo pluton display essentially flat REE patterns strongly resembling those observed for near-trench volcanic rocks. The REE patterns of quartz gabbros and tonalites of the western region correspond closely to those of circum-Pacific high-alumina basalts. The heavy REE depleted and fractionated patterns observed in the rocks of the central and eastern regions of the batholith do not have counterparts in oceanic island volcanic arcs, and few counterparts in continental margin volcanic arcs.

The REE variations generally correlate with other transverse asymmetries in major petrologic and geochemical characteristics. The abrupt depletion and fractionation in the middle to heavy REE and elimination of negative Eu anomalies appear coupled to an increase in Sr concentration and a marked restriction in lithologic diversity. This transition occurs over a range of initial $^{87}Sr/^{86}Sr$ ratios. The light REE enriched rocks of the eastern region are distinguished from the central and western regions by higher initial ratios. Geographic discontinuities in $\delta^{18}O$ and age

distributions in the batholith correlate approximately with the REE discontinuities, but locally diverge by the dimensions of one or two plutons.

Determinations of REE abundances in major and trace phases of a representative eastern region granodiorite indicate accessory sphene and allanite are the major reservoirs of REE in this rock. Hornblende is the only significant REE site in the major minerals, and in some batholithic lithologies it may be the dominant site. High-level crystal fractionations involving hornblende and accessory phases do not appear capable of producing the observed geographic characteristics. Contamination processes including upper crustal material also seem ruled out.

The REE and other geochemical variations across the batholith appear to originate in deep-seated sources. Partial melting in source rocks in which assemblages rich in plagioclase give way laterally to garnet-bearing assemblages in source regions of broadly basaltic composition which are already zoned in light REE abundances, $^{87}Sr/^{86}Sr$, $\delta^{18}O$, and possibly Sr content appears to account for most of the observed features. The geologic context of the source material remains largely undefined and may include mantle and crustal components. However, the source regions for all parts of the batholith must have bulk compositions and phase assemblages capable of producing the dominant

tonalite and low-K_2O granodiorite lithologies. This major constraint appears to strongly limit the amount of sialic crustal material permitted to be present in the source regions. The geometry of the convergent boundary appears to have determined the elongate form of the batholith, and, probably, the general alignment of all the geochemical variations along its length. The results of this study may be useful in comparing possibly related crust-forming processes and products in other orogenic-plutonic terrains.

TABLE OF CONTENTS

CHAPTER 1 INTRODUCTION AND PREVIOUS WORK

CHAPTER 1

INTRODUCTION AND PREVIOUS WORK

THE PROBLEM OF THE ORIGIN OF BATHOLITHS AT CONVERGENT PLATE BOUNDARIES

Large batholiths exposed by erosion in continental margin orogenic provinces, such as the Cordilleran batholiths of western North America, are characterized by many features which serve to define their tectonic setting. These batholiths form elongate magmatic belts or arcs whose external geometry and internal structures parallel the regional tectonic grain. They occur in close proximity to and follow the trend of known or inferred continental margins or plate boundaries. They are often paralleled by oceanward belts of complexly deformed rock masses in large part composed of ocean floor sediments and basalts which appear to be tectonically added to the continent at the time of batholithic emplacement, and seem analogous to modern trench materials. The batholithic arcs are more closely flanked by thick accumulations of near coeval greywacke-turbidite deposits composed of little weathered volcanic and plutonic detritus.

Modern plate tectonic theory holds that the deformation and mountain building of orogenic episodes occur principally along convergent boundaries. These junctures between rigid lithospheric plates are identified in ocean basins and at continental margins as active elongate volcanic chains closely associated with oceanic trenches and dipping seismic zones. This dynamic tectonic environment is the site of plate consumption, intense magmatic activity, crustal thickening, deformation, and metamorphism. The rapid accumulation of volcanic materials builds an emergent arc which sheds debris, forming thick clastic deposits on its flanks. The resulting elongate belts of volcanic sequences, related plutonic intrusives, and derivative greywacke-rich sedimentary suites are recognized as the dominating components of many orogenic terrains, young and old. The correspondence in lithologies and structures of older orogenic belts and modern volcanic arcs allow such belts and the processes fundamental to their formation to be related to convergent plate boundaries.

Some of the most convincing arguments relating batholiths to convergent plate margins center about the close temporal, spatial, and chemical association of many batholiths and orogenic volcanic rocks. In northwestern Baja California, field, paleontologic, and geochronologic studies (Silver, Stehli, and Allen, 1956; 1963; Silver, Allen, and Stehli, 1969) have established an intimate temporal correspondence between a thick volcanogenic sequence and the plutons of the Peninsular Ranges batholith that intrude them. Jurassic plutonism in the Alaska-Aleutian Range overlapped in time with the accumulation of a thick Lower Jurassic volcanic sequence (Reed and Lanphere, 1973); volcanics that preceded intrusion of phases of the Sierra Nevada batholith are preserved in wall rocks and pendants (Bateman, Clark, Huber, Moore, and Rinehart, 1963).

Petrologic and geochemical asymmetries across modern volcanic arcs are now widely recognized and documented, and are a fundamental feature of these arcs. In a profound early work, Kuno (1966) observed a systematic variation in basalt type from tholeiitic to high-alumina to alkali-olivine across the axis of the Quaternary Japanese arc. Similar variations have been observed with stratigraphic position in other arcs as well (e.g. Gill, 1970). Kuno (1966) noted a relation between the chemistry of the volcanic rocks and the depth to the seismic zone, and later workers have established the generality of this concept for oceanic and continental margin arcs. The most emphasized chemical variable is the increase in K_2O relative to SiO_2 traversing away from the trench, and many trace elements appear to vary sympathetically with the major elements, as summarized by Jakes and White (1972). The petrogenesis of these volcanic rocks is still not well understood, but a clear relation between their chemical composition and the inherent geometric asymmetry of the subduction zone exists.

Many workers have emphasized that the volcanics associated with batholiths, from Precambrian to Cenozoic, are similar in composition to modern island arcs. The volcanic and plutonic suites overlap in composition, although the plutonics are commonly more silicic. Most importantly, the chemical variations across volcanic arcs have batholithic analogues. In fact, petrologic and chemical variations across batholithic belts were recognized long before those in volcanic arcs; for example, the west to east zonation of the plutonic rocks of the North American Cordillera (Lindgren, 1915; Moore, 1959). Chemical trends across the central Sierra Nevada have demonstrated that K_2O increases absolutely and relative to SiO_2 eastward (Bateman and Dodge, 1970). Similar variations have also been shown in the Alaska-Aleutian Range batholith (Reed and Lanphere, 1974), and in the batholithic rocks of southern California (Baird, Baird, and Welday, 1974).

Many aspects of the genesis of batholiths and how they relate to the subduction process continue to be actively investigated. What is the nature of the source materials of the continental margin batholiths and how do they differ from those of oceanic volcanic arcs? What are the physical conditions of and processes triggering the production of the batholithic magmas? How are they tied to the subduction process? Are these magmas all highly evolved differentiates or are they derived directly from their source regions? Were these magmas largely melt or mobilized crystal-rich mixtures?

Trace element abundances and variations are sensitive indicators of magmatic processes, making them a powerful petrogenetic tool. Interpretations of trace element characteristics, particularly rare earth elements (REE), have provided important contributions to petrogenetic models of volcanic arcs (e.g. Gill, 1974; Ringwood, 1974). While REE have been studied in several granitic terrains, the purpose of this investigation has been to determine for the first time the broad REE characteristics of a well studied batholith and to use this information to help interpret the genesis of the magmatic arc in a petrologic and tectonic context.

The characteristics of the Peninsular Ranges batholith which have established its setting as a magmatic arc, its general petrologic features, and a rationale for the study of this particular batholith are presented in the following section. The introduction is concluded with a review of REE properties, their behavior in magmatic systems, and distribution in the earth's crust. Observations on REE variations in whole rocks and minerals and the ensuing discussion of the origin of these variations in the light of other petrological and geochemical systematics form the core of this thesis. This work is concluded with a discussion of inferences and speculations concerning the nature of the source materials and the petrogenetic processes, and the broader problems of magmatic arc petrogenesis and continental growth.

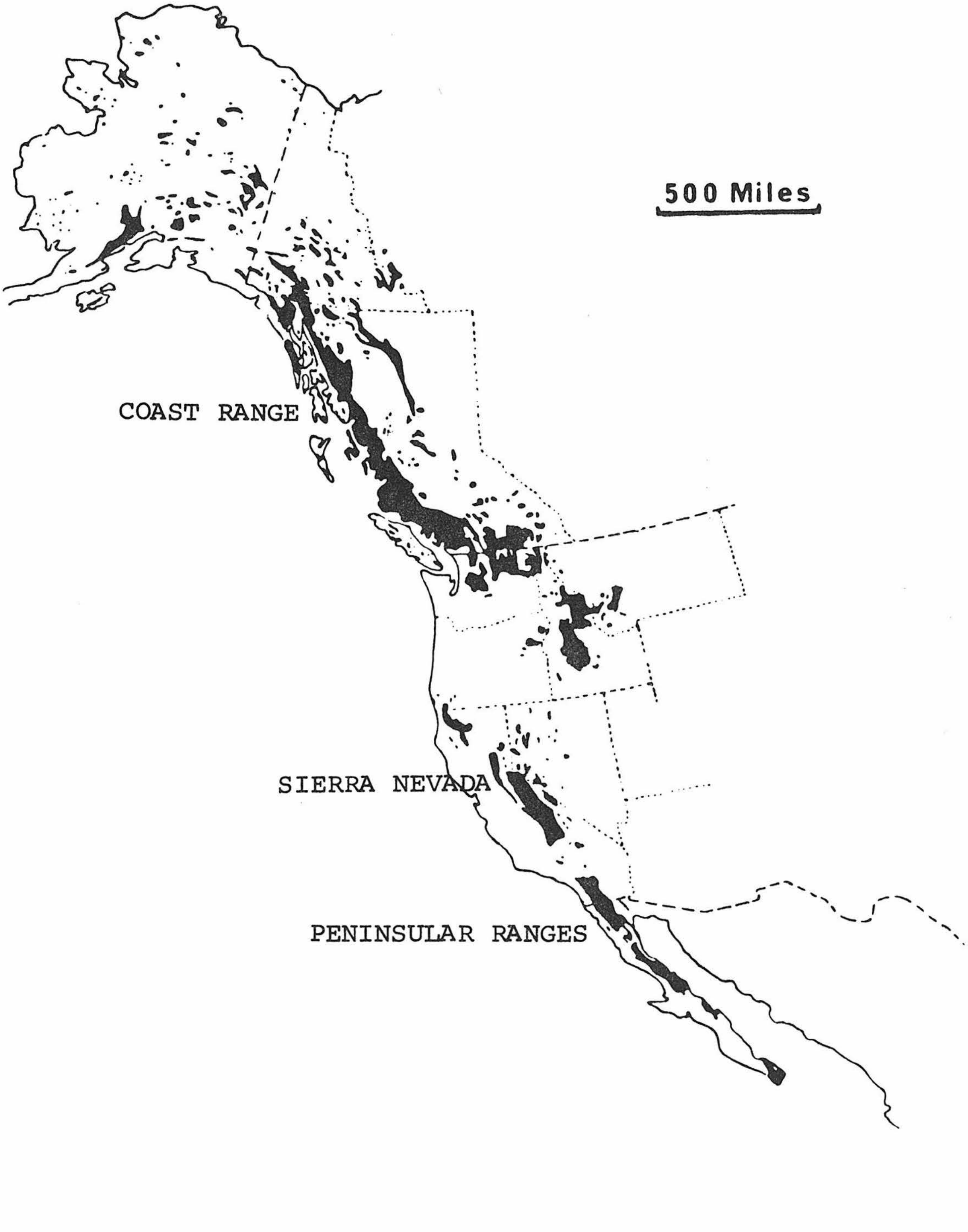
PENINSULAR RANGES BATHOLITH

OVERVIEW

The Cretaceous Peninsular Ranges batholith of southern and Baja California is one of the largest composite batholiths of the Cordilleran province of western North America. It is an elongate belt comprised of hundreds, perhaps thousands, of plutons paralleling the Pacific coast from the Los Angeles basin to the tip of Baja California (Fig. 1). The plutons of the batholith were emplaced in thick metavolcanic and metasedimentary sequences of Mesozoic and, at least locally, Late Paleozoic age. Mesozoic structural elements in the belt of prebatholithic rocks and in the plutons themselves tend to parallel the elongation direction of the batholith. They help define the trend of the magmatic arc across which major petrologic variations occur in both prebatholithic and batholithic rocks. Uplift and erosion quickly followed the Cretaceous emplacement of the batholith, many nearby localities receiving Upper Cretaceous sands and gravels containing its granitic debris.

Over the past several million years, the peninsula has been rifted from continental North America. Translation to the northwest along the Gulf of California-San Andreas fault system has and continues to occur as a result of the impingement of the East Pacific Rise spreading ridge. The Peninsular Ranges are therefore a tectonic fragment, trun-

Figure 1. The major well-exposed segments of the Cordilleran batholiths of western North America. The Peninsular Ranges, Sierra Nevada, and Coast Range batholiths display strong alignment to the northwesterly Cordilleran tectonic trend.


500 Miles
COAST RANGE
SIERRA NEVADA
PENINSULAR RANGES

cated on three sides: the San Andreas fault to the northeast, Salton Depression-Gulf of California on the east, and East Pacific Rise on the south. Geochronologic and geologic studies by Silver, Early, and Anderson (1975) have indicated that the eastern tectonic boundary of the Peninsular Ranges is not the limit of the batholith; emplacement progressed across Sonora, Mexico in the Late Cretaceous at least as far as Chihuahua.

In the present study, consideration of the batholith is focused on about 400 km of the northern Peninsular Ranges proper in southern California and northern Baja California. This area has received more study than the province to the south. Larsen (1948), in the northwestern portion of the Peninsular Ranges, provided the first thorough field and petrographic description of a part of this batholith. The reconnaissance study of Gastil, Phillips, and Allison (1975) has provided an important geologic summary and map of the state of Baja California. Geochronologic, geologic, and isotopic investigations of Silver and his coworkers have provided a progressive and unifying context for petrological and geochemical studies north and south of the border.

PREBATHOLITHIC ROCKS

The prebatholithic rocks of the Peninsular Ranges are almost if not entirely comprised of metamorphosed stratified rocks. There is, as yet, no place within the Peninsular

Ranges where it has been demonstrated that these rocks are underlain near surface by older crystalline basement.

In the western or coastal areas, volcanic, volcaniclastic, and other immature sedimentary rocks of Mesozoic age are predominant. In southern California, these include the argillitic and quartzitic rocks of the Bedford Canyon formation (Larsen, 1948), quartzofeldpspathic schists, arkosic quartzites, and amphibolites of the French Valley Formation (Schwarcz, 1969), and the Black Mountain (Hanna, 1926) and Santiago Peak (Larsen, 1948) Volcanics, predominantly of intermediate composition. The complete range of age of these rocks is uncertain, but the sedimentary formations include some Triassic and Jurassic strata in the Santa Ana Mountains (Silberling, Schoellhamer, Gray, and Imlay, 1961), and the Black Mountain Volcanics in San Diego County contain Upper Jurassic fossils (Fife, Minch, and Crampton, 1967). Stratigraphic relationships among all these rocks vary locally from unconformable to interlayered.

In northern Baja California the prebatholithic section of the western Peninsular Ranges is almost entirely of volcanic derivation. An example is the thick volcanic and volcaniclastic section of the Alisitos Formation, south of Ensenada. Of Aptian age (Allison, 1955), it has no recognized bottom and possesses features indicating rapid accumulation and intense tectonic activity (Silver et al., 1963; 1969). Volcanic activity closely preceded and possibly overlapped plutonic emplacements.

To the east, deformation and the grade of metamorphism increases (Silver et al., 1963; Schwarcz, 1969). Dominantly metasedimentary rocks, derived from feldspathic sandstones and shales, such as the Julian Schist of eastern San Diego County (Merriam, 1946; Everhart, 1951) and similar rocks in the Sierra Juarez and Sierra San Pedro Martir (Woodford and Harriss, 1938; Gastil et al., 1975) are common. In the vicinity of the San Jacinto fault zone, the prebatholithic rocks are at almandine-amphibolite facies and consist largely of quartzofeldspathic paragneisses and schists with minor interlayered marble and amphibolite (Sharp, 1965; 1967). A high density of intrusives and extensive deformation have largely obscured the age and original stratigraphic relationships in these rocks. In the desert ranges in Baja California, the only Upper Paleozoic fossils observed in place in the Peninsular Ranges are found in a section consisting of quartzofeldspathic metasedimentary rocks and mafic schists with thin interbedded marbles (McEldonney, in Gastil et al., 1975).

The above quoted works have established that a volcanic arc was built upon an unknown basement in the western Peninsular Ranges, and that a contemporaneous sedimentary basin may have existed to the east which received volcanic and probably continental detritus. This inferred basin may have been floored by Precambrian craton in Sonora (Anderson and Silver, 1977), but no such basement has yet been established in the Peninsular Ranges.

INTRUSIVE ROCKS

The Peninsular Ranges batholith is a composite batholith whose individual plutons range in lithology from gabbro (locally peridotite) to granite. Tonalites and low-K_2O granodiorites are the most common rock types (Larsen, 1948; Silver et al., 1975; Gastil et al., 1975). Individual plutons vary in size, but most are a few kilometers in diameter, and dimensions of tens of kilometers are not uncommon (Gastil et al., 1975).

Intrusive contacts are typically sharp (Larsen, 1948) and many examples of both passive and forceful emplacement are found. In the western Peninsular Ranges, plutons commonly intrude preexisting rock discordantly (Larsen, 1948) while to the east a deeper level of emplacement has been exposed by erosion and concordant intrusive contacts are the norm (e.g. Sharp, 1967; Murray, 1978). Larsen (1948) inferred the general sequence of intrusion to be from mafic to felsic compositions. Although, at least locally, field relations indicate exceptions (Silver, per. comm.) and the zircon U-Pb isotopic age studies of Banks and Silver (1966; 1969) require some modification of this generalization, most subsequent work supports this sequence.

Many plutons are compositionally uniform within modest limits limits, the gabbros being the principal exception (Miller, 1937; Larsen, 1948). Texturally, most of the rocks are medium to medium-coarse grained, hypidiomorphic granular, and grain size distributions may be equigranular or seriate, with porphyritic textures less common (Larsen, 1948). The minerals plagioclase, quartz, hornblende, biotite and alkali feldspar make up all but a few percent of most rocks. A common feature of many tonalite and granodiorite plutons is the presence of abundant mafic inclusions. These are usually a fine to medium grained, allotriomorphic-granular, more mafic rock with mineralogy similar to that of the host intrusive but in different proportions (Hurlbut, 1935; Larsen, 1948). Most of the plutonic rocks possess uniquely igneous characteristics such as euhedral oscillatory zoning in plagioclase, the character of the mineral assemblages, and lack of normative corundum despite high Al_2O_3 contents. Hybrid rocks are sometimes found near contacts (Larsen, 1948), indicating that contamination by assimilation is probably only a local, short range phenomenon.

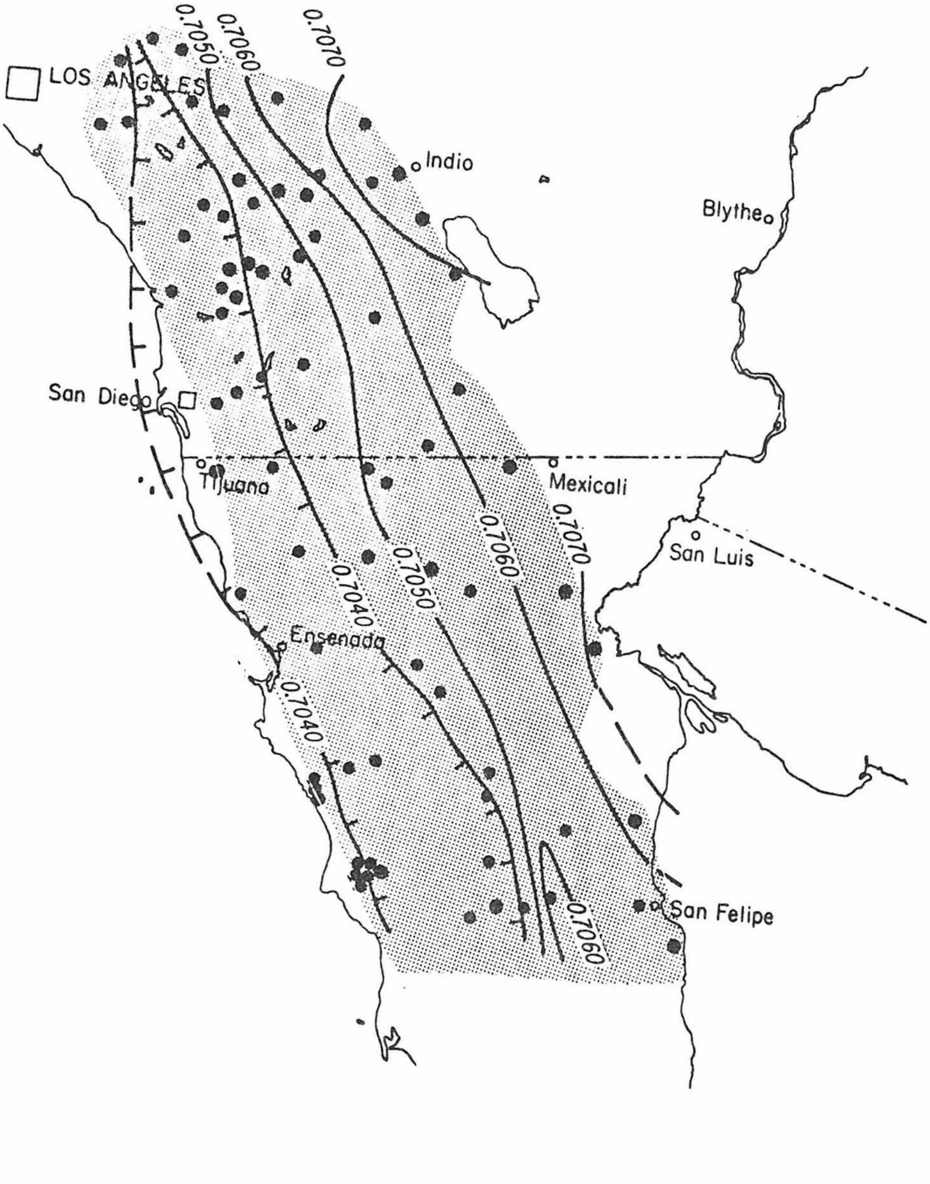
REGIONAL ASYMMETRIES

A most remarkable characteristic of this batholith is regionally developed asymmetries in many features and parameters. These include geochronologic, geochemical, and petrologic characteristics, the trends of which are transverse to the structural trends of the batholith and therefore appear to bear some fundamental relationship to its origin.

A space-time progression of plutonic emplacements eastward from the Pacific margin (130-110 my) to the east side of the peninsula (100-90 my) has been reported by Silver et al. (1975), utilizing U-Pb isotopic ratios in zircons. These data demonstrate that emplacements were concentrated in narrow but continuous linear zones that progressed eastward across the batholith region during the Cretaceous. The Peninsular Ranges batholith was therefore the product of a single major intrusive cycle of limited duration. K-Ar ages also are progressively younger eastward but with steeper space-time gradients, and have been interpreted as apparent ages reflecting uplift and cooling during the latest Cretaceous (Krummenacher, Gastil, Bushee, and Doupont, 1975).

Early and Silver (1973) and Silver et al. (1975) reported a strong geographic variation in initial $^{87}Sr/^{86}Sr$, rising to the ENE from less than 0.7035 to more than 0.7075 with at least two significant reversals (Fig. 2). They have noted that these initial ratios are independent of rock type, with the exception of some gabbros which are slightly less radiogenic than adjacent plutons. Sr concentrations also show a well defined increase to the east, largely independent of lithology. Recently, Taylor and Silver (1978) re-

FIGURE 2. Isopleths of constant initial $^{87}Sr/^{86}Sr$ ratio in the northern Peninsular Ranges (Early and Silver, in preparation). Note the parallelism of the contours to the northwest-trending axis of the peninsula. The gradient rises to the north east, but two reversals are present: one along the western margin of the peninsula, and one along the southernmost eastern margin.


0.7050
0.7060
0.7070
LOS ANGELES
Indio
Blythe
San Diego
Tijuana
Mexicali
0.7060
0.7070
San Luis
0.7040
0.7050
Ensenada
0.7040
San Felipe
0.7060

ported regional variations in whole rock $\delta^{18}O$, values for tonalites and granodiorites ranging from about 6.0 to 7.0 in the west to in excess of 11.0 in the east.

Prior to describing regional variations in petrologic character, it is emphasized that a single rock type - tonalite - is a major and usually dominant rock type for every region of the batholith in the Peninsular Ranges (Larsen, 1948; Gastil et al., 1975). The tonalites across the batholith have consistent major mineralogic and bulk compositions (Appendix I), yet display all of the striking variations discussed above.

Within the context of the prominence of the tonalites across the batholith, there is considerable variability in the presence and relative proportions of other lithologies. The lithologic diversity of the western Peninsular Ranges, where gabbro and siliceous leucogranodiorite join tonalite as major rock units, gives way to the east to a more restricted range of compositions strongly dominated by tonalites and granodiorites low in K_2O (Larsen, 1948; Silver et al., 1975; Gastil et al., 1975). Larsen also noted that the boundary between regions with this contrasting lithologic character is about parallel to the structural trends of the batholith.

In at least some of the desert ranges to the east of the Peninsular Ranges proper, granodiorites appear to be the most abundant rock type (Miller, 1946; Hirschi and de

Quervain, 1933). Tonalites appear to drop in abundance and be largely supplanted by granodiorites low in alkali feldspar (Silver, per. comm.). These granodiorites generally have color indices as high as the tonalites and comparable amounts of quartz. Adamellites are present but not abundant.

Gastil, Krummenacher, Doupont, and Bushee (1974) have described the west-east petrographic variation in terms of gabbro, tonalite, and adamellite sub-belts.

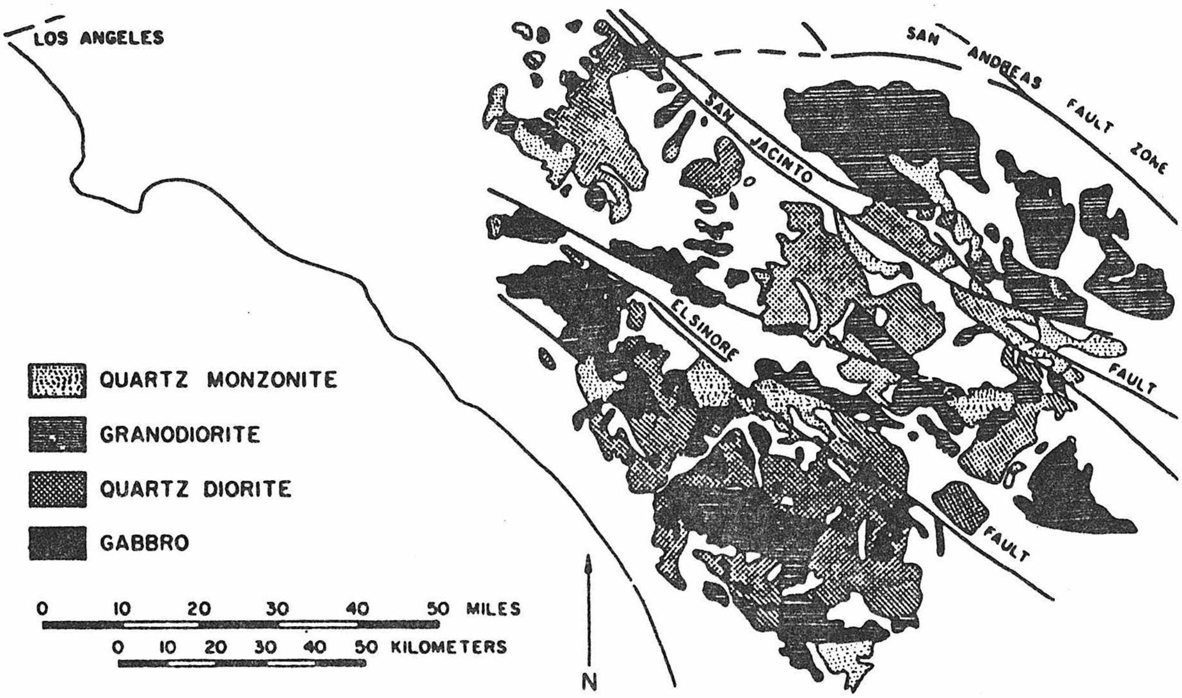
These petrographic observations have been complemented by recent chemical studies by Baird, Morton, Baird, and Woodford (1974) and Baird, Baird, and Welday (1974). Batholithic rocks have been sampled in a limited part of the northern Peninsular Ranges and in the Transverse Ranges to the north and east, across the San Andreas fault. An asymmetrical distribution of lithologies within the Peninsular Ranges is indicated in Figure 3.

TECTONIC SETTING

The restoration of the peninsula to its pre-Cenozoic position with respect to the continent, as reconstructed by many workers (e.g. Moore, 1973), is assumed in the following discussion.

To the east of the Peninsular Ranges, in Sonora, lies the Precambrian craton (Anderson and Silver, 1977). Late Precambrian and Paleozoic deposits, intruded by Mesozoic plutons including many related to the Peninsular Ranges

Figure 3. The distribution of major plutonic rock types in the northernmost Peninsular Ranges, from Baird et al. (1974b). Tonalites are present across the batholith, but abundant gabbros are restricted to the western part of the peninsula.



from Baird, et al., 1974

batholith, have been found in this area.

To the west, little material older than Mesozoic can be identified. Off the present Pacific coast, an assemblage of greywacke, bedded chert, serpentinite, and glaucophane-bearing schists are spottily exposed from beneath the cover of Late Cretaceous and younger sedimentary rocks of the continental borderland (Stuart, 1974). These materials of oceanic character probably correlate with the Franciscan of central and northern California, and are assumed to be approximately as old as the batholith.

Therefore at the time of its formation, the batholith appears to have been bracketed to the west by oceanic trench and arc-trench materials and to the east by the Precambrian craton and younger continental margin deposits. This establishes the setting for this batholith at a continental margin. The pervasive tectonic grain of the batholith, the volcanic pile that accumulated in association with it, and the oceanic trench-type materials flanking it to the west are most satisfactorily interpreted as the result of convergent motion with an associated subduction zone and magmatic arc.

Integration of this information with broad-based regional geologic and other studies allows the following assessment of the tectonic history of the Peninsular Ranges up to batholithic emplacement. The generally quiescent continental and continental margin environment in which the sediments of the Cordilleran geosyncline were deposited during the Late Precambrian and Paleozoic was terminated during the first half of the Mesozoic (e.g. Hamilton, 1969). In California, a northwest trending Jurassic and possibly other older Mesozoic arcs were developed at a high angle to the preexisting trends (Hamilton, 1969; Kistler, Evernden, and Shaw, 1971; Silver and Anderson, 1974), and gave rise to prominent mountain chains which shed much debris off their flanks. Large lateral shears may have juxtaposed diverse terrains (Silver and Anderson, 1974; Saleeby, 1977), which was followed by resumption of subduction giving rise to a NNW trending Cretaceous arc. In the case of the Peninsular Ranges, this arc appears to have developed to the west of the Jurassic arc on a site composed of diverse oceanic and continental margin crustal materials (Silver and Anderson, per. comm.). The complexity of the prebatholithic crust and mantle into which it was emplaced must be addressed in the consideration of possible source materials for the batholith.

RATIONALE FOR STUDY

Geochronologic investigations in several of the Cordilleran batholiths up to 1973, summarized by Lanphere and Reed (1973) and Kistler (1974), have indicated that many are composed of several spatially discrete belts of plutons emplaced in several distinct cycles. The Jurassic and Cretaceous belts are the most thoroughly documented as they appear to account for the majority of plutons. These arcs are long, most probably through-going features stretching along much of the continental margin of North America during the Mesozoic (e.g. Hamilton, 1969; Silver and Anderson, 1974). The batholiths, as they are geographically known, are well exposed continuous segments of these arcs which commonly overlap in distribution. It has been inferred by many workers that an arc is the product of subduction at a plate boundary; spatial divergences and transgressions in magmatic arc activity in the Cordillera may be in response to progressive changes and reorientations in plate boundaries and plate motions (Silver, per. comm.). Petrologic and geochemical trends across batholiths composed of multiple arcs (e.g. Bateman and Dodge, 1970; Kistler and Peterman, 1973; Reed and Lanphere, 1974) can be expected to reflect the composite nature of these batholiths rather than asymmetries within a single batholithic arc.

The Peninsular Ranges batholith has been argued to have been emplaced during a single intrusive cycle with no spatial overlap with older or younger plutonic arcs (Silver et al., 1975). In contrast to most of the other batholiths of the Cordillera, the less geochronologically and structurally complex Peninsular Ranges batholith offers the opportunity to examine the petrologic and geochemical characteristics of a plutonic belt closely analogous to modern volcanic arcs. The systematic geographic zonation of many properties across this batholith as discussed above is one of the most significant and intriguing set of observations in batholithic research. The origin of these variations is fundamental to the understanding of the generation of batholiths, the characterization of their source regions, and the growth and evolution of the continents.

The basic design of this dissertation from its inception has been the study of the REE characteristics of this batholith in light of the above considerations. The REE are a sensitive and powerful petrogenetic tool which will complement previous geologic, geochronologic, petrologic, and geochemical studies in this batholith. The large body of previously collected data, many on the same samples subsequently used in this study, provide insight for sampling design and are available for integration with and extension of the capabilities of the present study.

RARE EARTH ELEMENTS

PRELIMINARY STATEMENT

The purpose of this section is largely to review for the reader the chemistry, geochemical behavior, and distribution of the REE, and their application to petrologic problems. A discussion of the controls influencing partitioning of REE among minerals and melts is provided, as well as a review of previous studies in granitic terrains. Readers who are knowledgeable about these subjects are invited to bypass all or parts of this section. However, note that certain essential premises for some of the conclusions of this study are developed here.

REE AS TRACE ELEMENTS

In most rock systems, the REE are present at the parts per million level and behave as trace elements. Trace elements are defined here as those elements whose presence in a rock does not significantly alter the stability fields of its major minerals. This usually requires very low concentration levels, and Henry's Law behavior is typically observed. A particular element may behave as a trace element in some phases, but not in others.

Trace element abundances in igneous rocks are now commonly used to evaluate petrogenetic processes and source region characteristics. Compared to major elements, trace elements display much larger ranges in concentration, often exceeding an order of magnitude in related rocks. Such large variations have been observed even among rocks with limited bulk compositional differences, indicating trace element abundances may record potentially valuable information not necessarily available from major elements. If their distributions among phases are known, trace elements may provide insights to petrogenetic mechanisms independent of phase equilibria considerations of the major components. Combined, these two approaches may often strongly constrain possible solutions.

In practice, the full potential value of trace element studies frequently is not realized. Perhaps the most significant factor limiting their effectiveness is the large range in observed absolute values of apparent crystal/melt distribution coefficients, usually measured in volcanic rocks. This range can be in part due to analytical difficulties (imprecise mineral analyses; impure mineral and glass/groundmass separates) and several aspects of nonequilibrium in volcanic rocks (mineral zoning due to Rayleigh fractionation and kinetic disequilibrium; melt inhomogeneities; exotic crystals, etc.), but some and perhaps much of the observed effect may be real differences in partitioning dependent on other factors. Trace element partitioning between melt and crystal is a measure of the relative affinities of an element for these phases. These affinities may be described thermodynamically, but at an atomistic level, they are a consequence of the structural characteristics of both the crystal and melt, and charge balance requirements. Temperature, pressure, and composition affect partitioning to the extent which they modify these properties. At the present time, little quantitative information is available on the dependence of partitioning on these variables. In many mineral-melt systems, the qualitative sense of response can only be surmised.

Among trace elements, the rare earth elements (REE) have added power as information can be obtained from the relative fractionation of this series of elements, independent of absolute values. This has been an important factor influencing the selection of the primary investigative tool in this study. The properties of the REE, their behavior in magmatic systems, and their distribution in crustal lithologies are summarized in the following sections.

SOME PROPERTIES OF RARE EARTH ELEMENTS

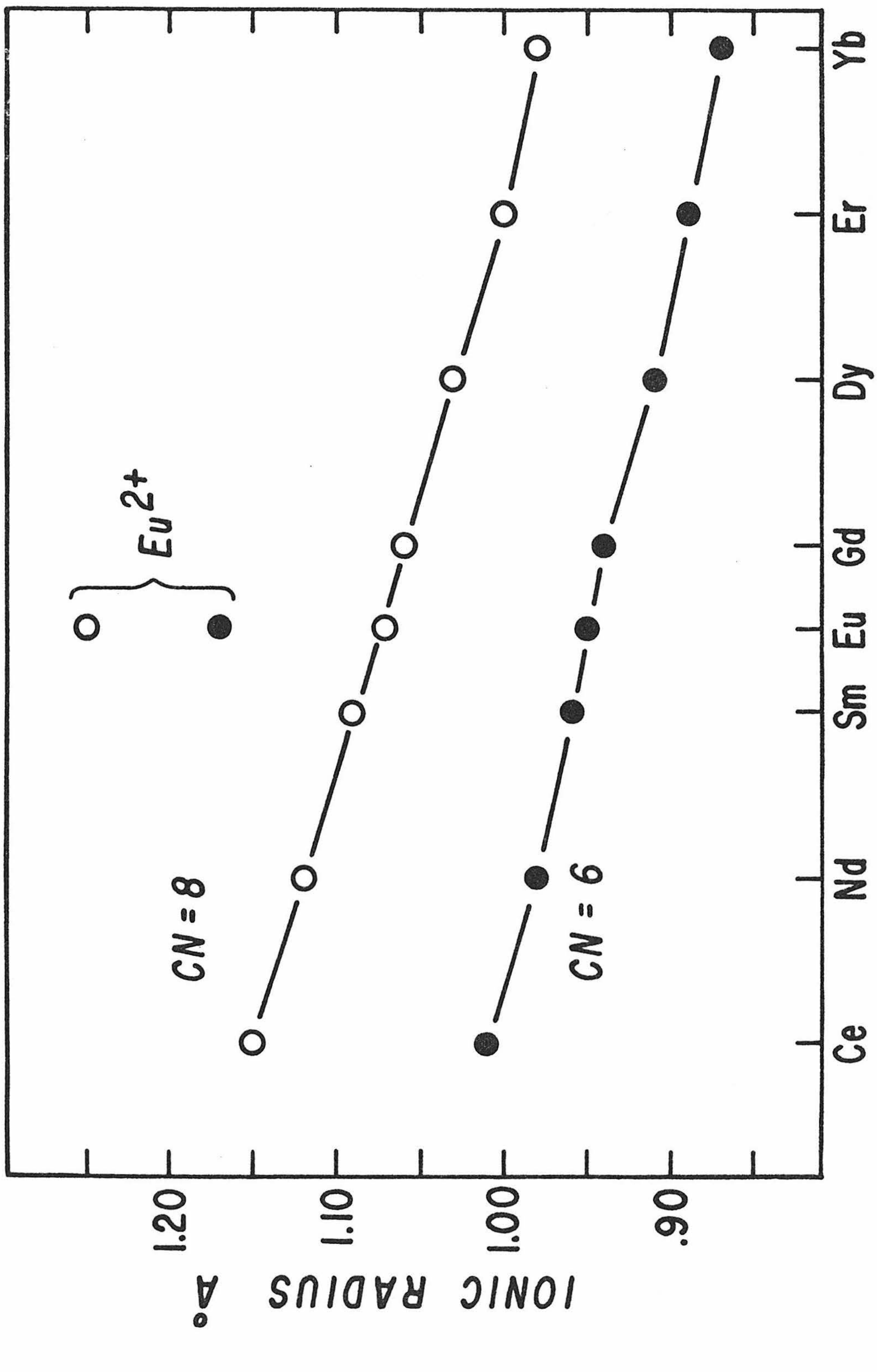
The REE are a closely related group of 14 metallic elements, atomic numbers 58 through 71, that make up a very small part of the matter of the universe. These elements display a uniformity in their chemical behavior, a result of the similarity in their size and electronic structure. The trivalent oxidation state is commonly observed in nature for almost all of the REE, the exceptions being Ce and Eu which can exist partly as Ce^{4+} and Eu^{2+}. With increasing atomic number, the 4f electron energy level is progressively filled. This is accompanied by small but systematic decreases in atomic and ionic radii, called the lanthanide contraction, which is responsible for most of the variations in chemical behavior among the REE. Ionic radii of eight of the REE for both six and eight fold coordination (Fig. 4) show this smooth decrease in radii with increasing atomic number. Eu^{2+} is about 15% larger in ionic radius than Eu^{3+}, an important difference.

The REE are strongly lithophile in their geochemical behavior. Their moderate size and high charge allows them to enter many common and rare rock-forming minerals and to form strong bonds. They do not appear to be as susceptible to mobilization and exchange as alkali and alkaline earth elements. The closeness of their ionic radii to Ca^{2+} among the major elements supports observations that REE appear to follow calcium preferentially into minerals where the difference in charge can be compensated.

NORMALIZATION OF RARE EARTH ELEMENT NOTATION

Early analyses of REE in natural materials, particularly chondritic meteorites and shales, revealed a complex but recurring dominant saw-tooth pattern in plots of abundance versus atomic number. This basic pattern is believed to have been produced from nucleosynthetic processes; its effects are seen in all natural materials. In order to simplify the complexity of graphical displays, several workers

FIGURE 4. Ionic radii of eight rare earth elements in six- and eight-fold coordination (Shannon and Prewitt, 1969; 1970). Trivalent radii are plotted for all elements; divalent Eu is also shown.


IONIC RADIUS Å
1.20
1.10
1.00
.90
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
CN = 8
CN = 6
Eu^{2+}

in the early 1960's suggested normalizing REE abundances, element by element, to a standard of comparison (e.g. Coryell, Chase, and Winchester, 1963). This standard has most often been taken as an average of chondritic meteorites, as these materials show only a limited variation in relative REE abundances, and also are thought to have possibly undergone fewer fractionating processes than most terrestrial materials. On a chondrite normalized REE diagram, a material with the same relative abundances as chondrites will plot as a flat line. Absolute abundances are compared along the ordinate of such a diagram, which is usually logarithmic. Not everybody uses the same normalization values. In this work, I have followed the suggestion of Sun and Hanson (1975) and have used the precisely determined concentration values of the meteorite Leedy (Masuda, Nakamura, and Tanaka, 1973) divided by 1.20 to bring these in close agreement to the more commonly used but less precisely known values for a chondritic average by Haskin, Haskin, Frey, and Wildman (1968). These adjusted values are given in Table I. Of course when REE abundances are ratioed as in the case of calculating a distribution coefficient between a mineral and melt, the normalization factor is eliminated.

MINERAL-MELT RARE EARTH FRACTIONATIONS

Presentation of Data From Literature

The mineral/melt distributions of REE for minerals

TABLE I

ARBITRARY CONCENTRATION VALUES IN PPM USED FOR NORMALIZATION

Ce	.813
Nd	.597
Sm	.192
Eu	.072
Gd	.259
Dy	.325
Er	.213
Yb	.208

commonly observed in granitic rocks and which might possibly be present in their source regions are considered. Almost all the distributions summarized here are from phenocryst/matrix measurements in volcanic rocks. For most minerals, measurements on rocks covering a large range in bulk composition are presented. Errors are contributed from both natural and laboratory sources as discussed above, but in most cases the general level and shapes of the REE distributions are reproducible suggesting they are correct. Presentation is in approximate order of increasing polymerization of silicon tetrahedra in the structures, with nonsilicates considered last.

In all the following discussions, the terms light, middle, and heavy REE refer approximately to Ce-Nd, Sm-Eu-Gd, and Dy-Er-Yb, respectively.

Olivine (Fig. 5) rather effectively excludes all REE relative to melt with distribution coefficients (DC) much less than 1. Garnet and zircon (Fig. 6a,b) accept the heavier REE readily (DC > to >> 1), but discriminate against the light REE relatively, producing strongly fractionated patterns. Orthopyroxenes (Fig. 5) have low contents of REE relative to melt (DC < 1) with light REE-poor distribution patterns. In contrast, calcic clinopyroxenes (Fig. 7a) have higher mineral/melt ratios (about 1) reflecting their Ca content, and a distinct concave-down shape. A very similar shape is uniformly displayed by calcic hornblendes (Fig.

7b), ranging to higher absolute values (DC > 1 to >> 1) for heavy REE. Biotites (Fig. 7b) discriminate against all REE (DC < 1) and have fairly flat patterns. Plagioclase feldspars (Fig. 8a) are always low in REE relative to melt (DC > 1 to >> 1) except for Eu, giving rise to the well known positive Eu anomalies characteristic of feldspar. Apatites (Fig. 8b) are enriched in all REE (DC >> 1), with some preference for the middle REE but negative Eu anomalies.

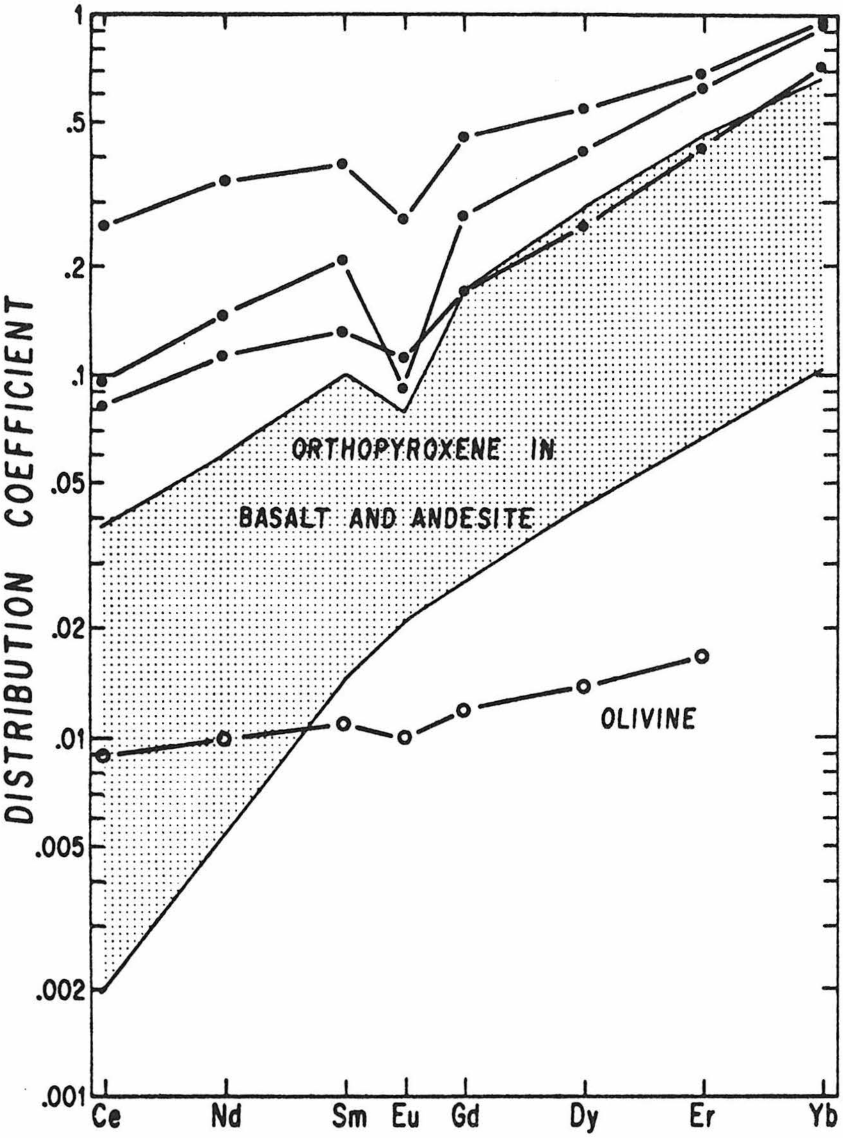
Discussion

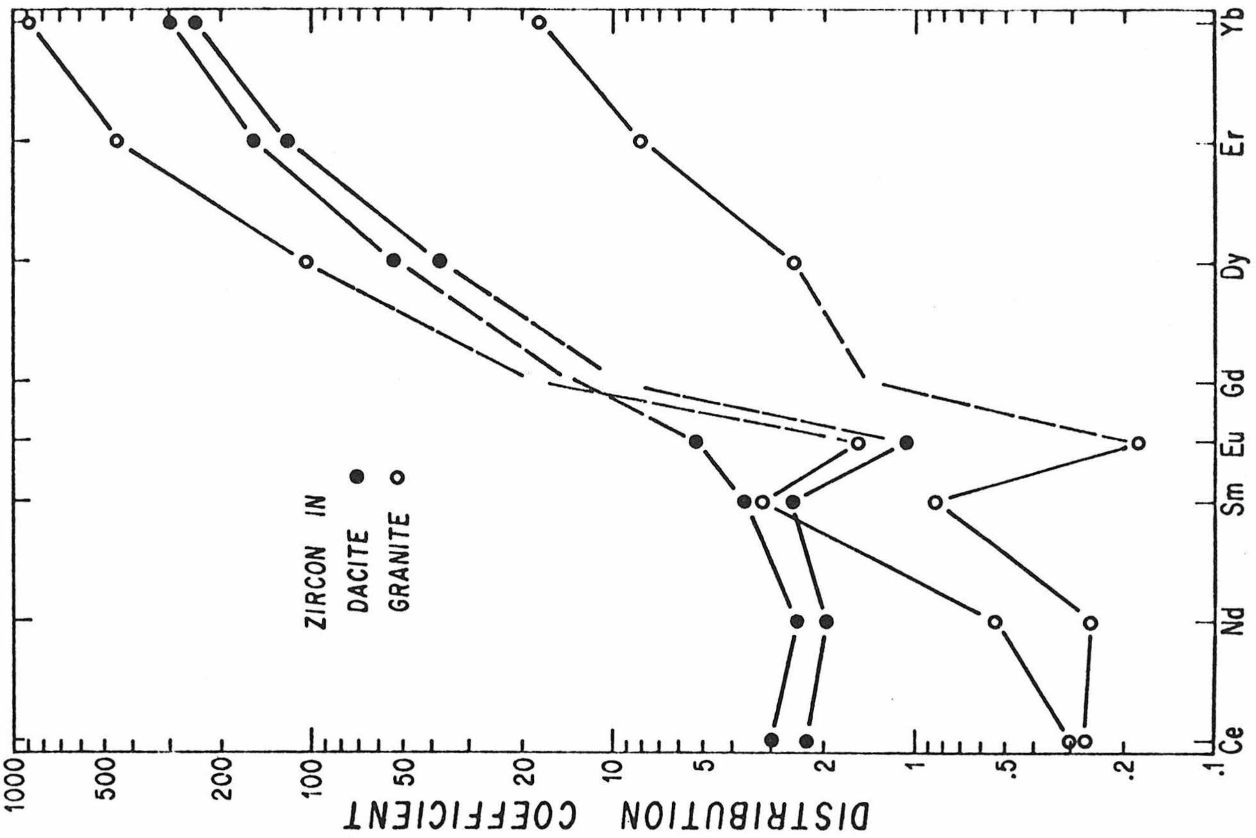
The general form and levels of the distribution patterns of mineral/melt ratios can be reconciled with aspects of the crystal structures for most of these minerals. The very low levels of REE in olivine are a consequence of the small octahedral sites of the tightly knit olivine crystal structure. The acceptability of the heavy REE in garnet and zircon, minerals also composed of independent SiO_4 tetrahedra, appears related to the presence of eight-fold coordination polyhedra in which the REE almost certainly reside. Garnet has a three dimensional framework of alternating tetrahedra and octahedra which, in response to substitutions, can expand or contract the eight-fold site. The strong exclusion of the lighter REE in both garnet and zircon indicates that the eight-fold sites are very sensitive to the size of the substituting REE ion.

The larger mineral/melt ratios observed for REE in Ca-bearing clinopyroxenes compared to orthopyroxenes is the

FIGURE 5. Selected distribution coefficients for olivine and orthopyroxene. A field for orthopyroxene in basalts and andesites is shown, as well as a few individual determinations in dacites. Sources: Schnetzler and Philpotts, 1970; Nagasawa and Schnetzler, 1971.

FIGURE 6. a) Selected distribution coefficients for garnet. Values for garnet in eclogite and peridotite calculated from REE distribution with coexisting clinopyroxene, assuming a clinopyroxene/melt REE distribution. Sources: Shimizu, 1975; Philpotts et al., 1972; Schnetzler and Philpotts, 1970. b) Distribution coefficients for zircon in dacites and granites. Values for zircon in granites calculated from whole rock rather than melt REE concentrations. Source: Nagasawa, 1970.


DISTRIBUTION COEFFICIENT
1
.5
.2
.1
.05
.02
.01
.005
.002
.001
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
ORTHOPYROXENE IN
BASALT AND ANDESITE
OLIVINE


ZIRCON IN
DACITE ●
GRANITE ○
DISTRIBUTION COEFFICIENT
1000
500
200
100
50
20
10
5
2
1
.5
.2
.1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

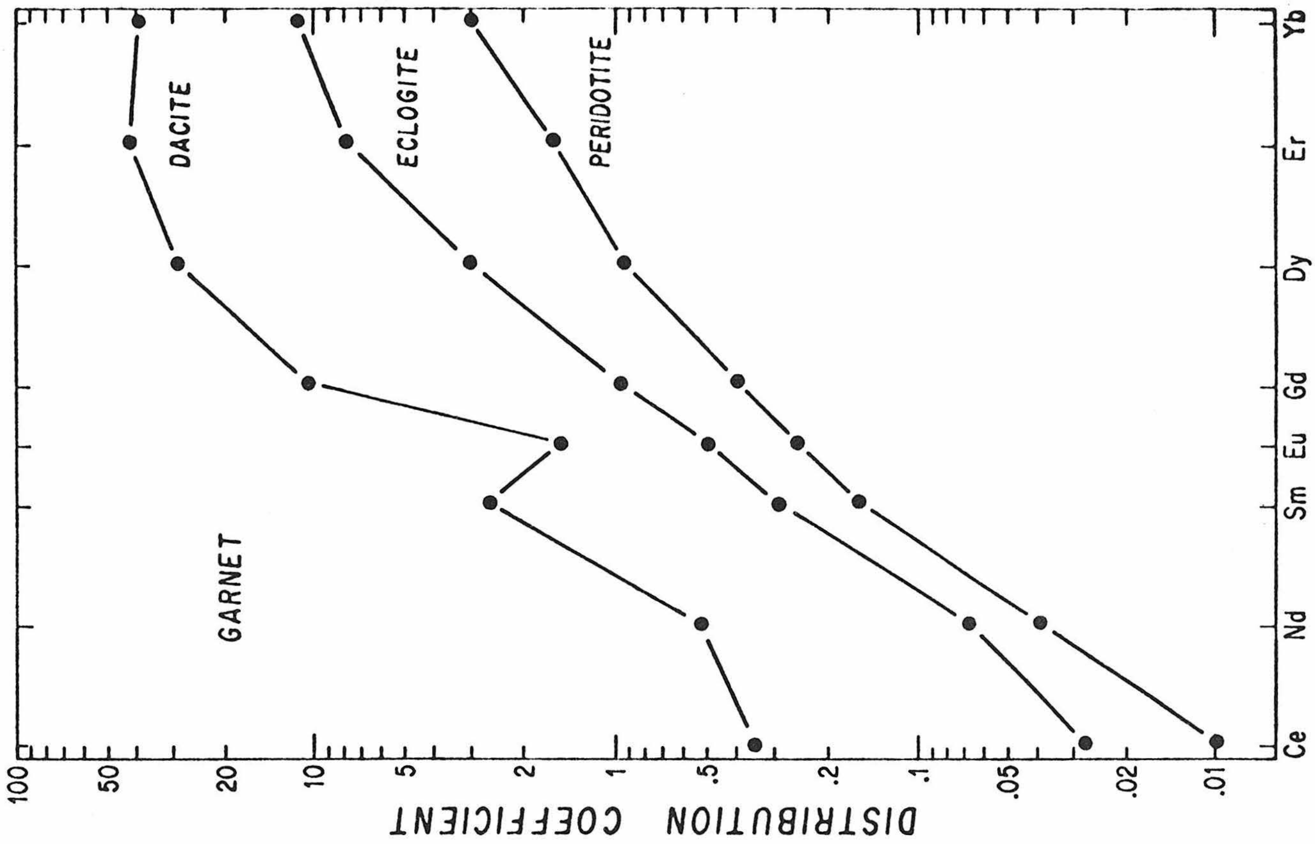
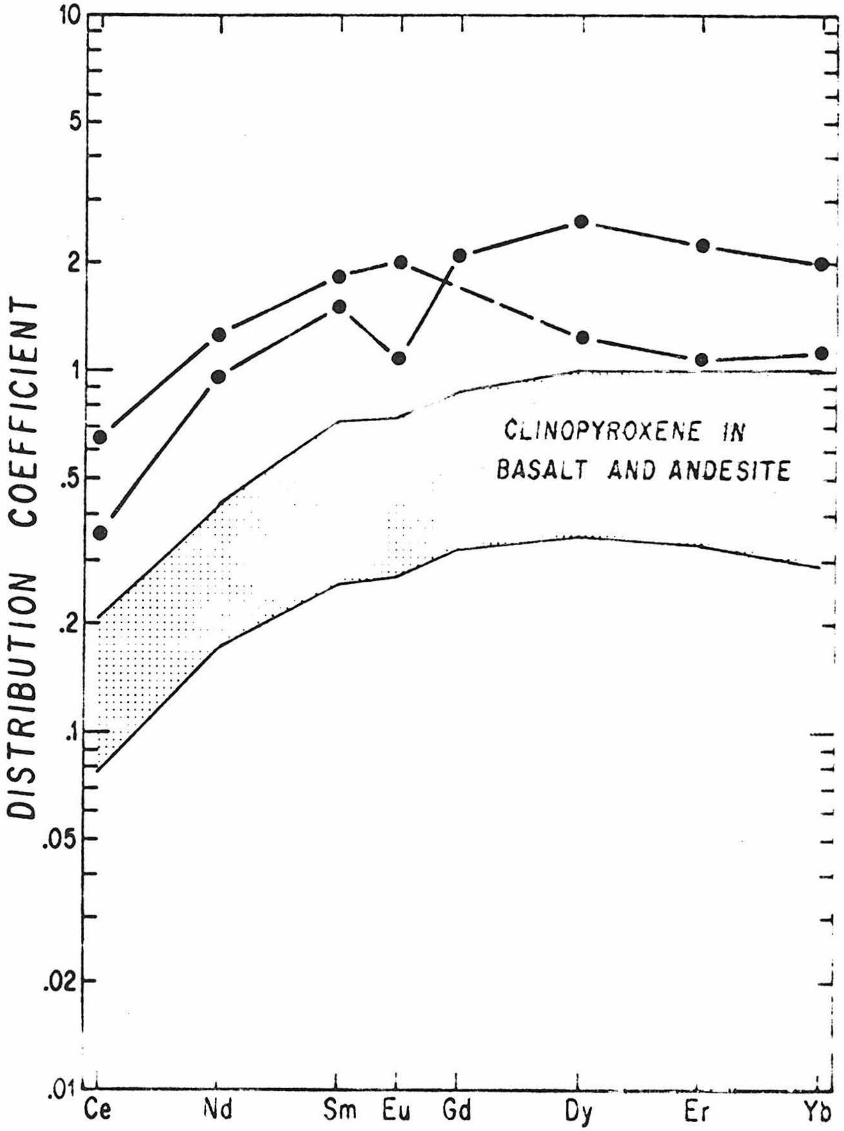
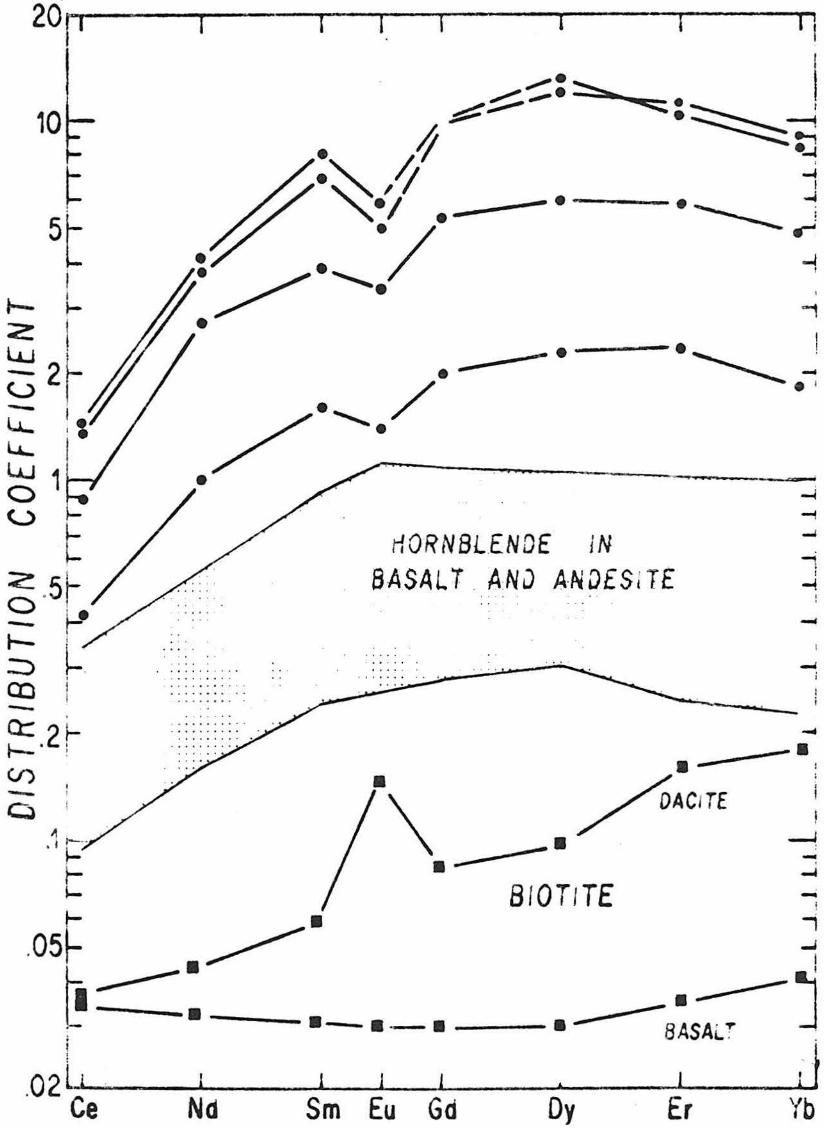
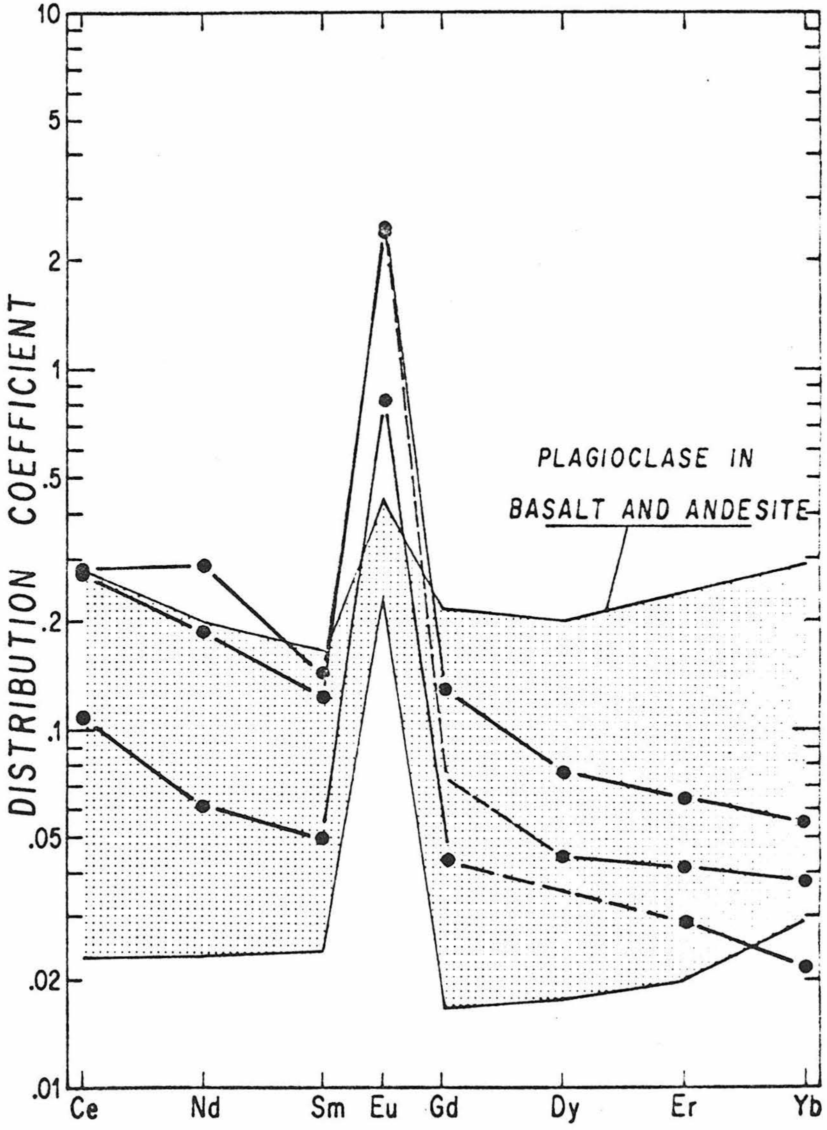

GARNET
DACITE
ECLOGITE
PERIDOTITE
DISTRIBUTION COEFFICIENT
100
50
20
10
5
2
1
.5
.2
.1
.05
.02
.01
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

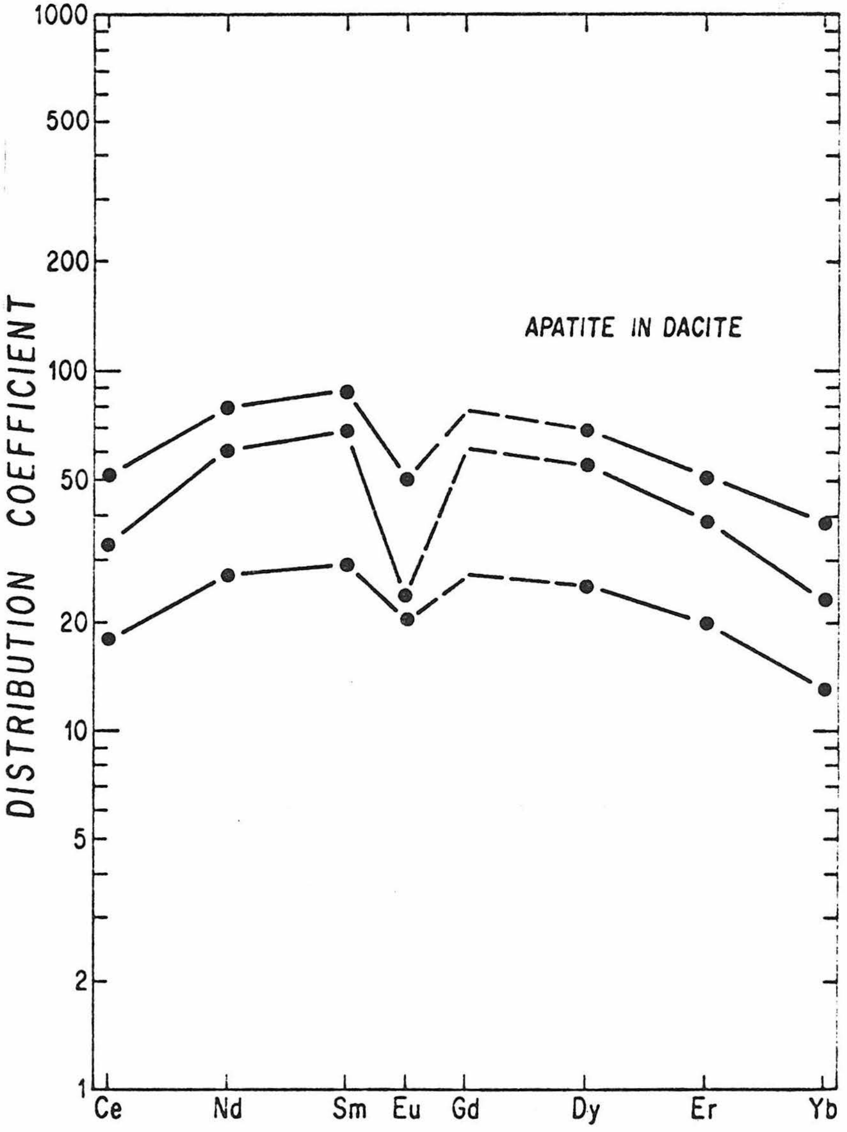
FIGURE 7. a) Selected distribution coefficients for clinopyroxene. Field for a representative range for clinopyroxene in basalts and andesites is shown, as well as two determinations in dacites. Sources: Schnetzler and Philpotts, 1970; Nagasawa and Schnetzler, 1971. b) Selected distribution coefficients for hornblende and biotite. The range of values for hornblende in basalts and andesites is shown, as well as additional determinations in dacites. Sources: Philpotts and Schnetzler, 1968; Higuchi and Nagasawa, 1969; Schnetzler and Philpotts, 1970; Nagasawa and Schnetzler, 1971.

FIGURE 8. a) Selected distribution coefficients for plagioclase. A field for plagioclase in basalts and andesites is shown, along with a few determinations in dacites. Sources: Schnetzler and Philpotts, 1970; Nagasawa and Schnetzler, 1971. b) Selected distribution coefficients for apatite in dacites. Source: Nagasawa, 1970.


DISTRIBUTION COEFFICIENT
10
5
2
1
.5
.2
.1
.05
.02
.01
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
CLINOPYROXENE IN
BASALT AND ANDESITE


DISTRIBUTION COEFFICIENT
20
10
5
2
1
.5
.2
.1
.05
.02
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
HORNBLENDE IN
BASALT AND ANDESITE
DACITE
BIOTITE
BASALT


DISTRIBUTION COEFFICIENT
10
5
2
1
.5
.2
.1
.05
.02
.01
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
PLAGIOCLASE IN
BASALT AND ANDESITE


DISTRIBUTION COEFFICIENT
1000
500
200
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
APATITE IN DACITE

result of the larger Ca-bearing M2 site (whose coordination is usually considered to be eight-fold when occupied by Ca) in clinopyroxene. The small, regular M1 octahedral site in both pyroxene structures is equivalent to the octahedral sites in olivine, and are not expected to hold appreciable amounts of REE. The parallelism in the patterns of the hornblendes and calcic pyroxenes are consistent with the crystallographic similarity of the M4 and M2 sites, respectively, in these minerals. The alkali-bearing A site of the amphibole structure is not expected to accommodate the usually smaller and highly charged REE ions. This is supported by the low REE distributions in the biotites, whose interlayer alkali site is similar to the amphibole A site. The absence of a site in biotite equivalent to the amphibole M4/pyroxene M2 site, and the presence of octahedral sites analogous to the octahedral or six-fold sites in those minerals further explains the observed biotite REE uptake.

The behavior of REE in the feldspars must be viewed somewhat differently. While some discrimination related to size is evident in the slightly light REE enriched patterns, the gross exclusion of trivalent REE from feldspar might be attributed to charge balance constraints. Only Eu^{2+} appears readily accepted into the feldspar structure. Feldspar might be incapable of providing coupled substitutions allowing the charge balance of 3+ cations for Ca^{2+}, Na^{+}, or K^{+}.

A large range in the values of mineral/melt ratios for several of these minerals is observed, most notably hornblende and garnet. In general, the observed values increase with increasing SiO_2 content of the host magma. The constancy in the shape of the patterns over these large ranges suggests that the size of the REE sites in the minerals are not changing and therefore not controlling these values. Diffusion-limiting effects, due to mineral growth rates outstripping the diffusion rate of REE in the melt, are not considered to have controlled these patterns. Such effects would tend to bring mineral/melt ratios closer to unity, yet the variations in the data clearly show transgressions of this value while slopes do not change.

Alternatively, the composition of the melt may exert important controls on mineral/melt distributions. The ability of melts to have compositionally dependent preferences for REE and other trace elements has been demonstrated experimentally by partitioning between immiscible melts (Watson, 1976). Ca-Fe-rich mafic melts are observed to have several times the REE concentrations of SiO_2-rich felsic melts they are in equilibrium with, suggesting that the composition of melts may have a first order effect on the partitioning of REE. The qualitative sense of this partitioning is in agreement with concepts of melt structure (e.g. Hess, 1977). In mafic melts the low degree of polymerization of SiO_4 tetrahedra makes available nonbridging oxygens to coordinate divalent and trivalent cations such as Ca and the

REE in the melt. With increasing SiO_2 content, the degree of polymerization in the melt increases and reduces the number of available nonbridging oxygens and will probably also affect the energetics of coordination. As appropriate sites are still available in some minerals crystallizing from the melt, partitioning into these minerals would be enhanced. This will be true for Ca and other divalent cations as well as the REE.

It is not anticipated that melt structure could influence the shape of REE mineral/melt distributions as the configuration of sites in melts is much less restricted than sites in minerals, and would not be able to discriminate among the small differences in the REE. This is supported by the limited experimental work of Watson (1976).

It would seem that melt structure may control the absolute value of REE mineral/melt ratios without changing the shapes of the patterns, and therefore provide an explanation for the observed variations in general distribution coefficients. The shapes of the patterns would be controlled by the mineral only.

Small to moderate Eu anomalies are observed in the REE mineral/melt ratios of phases other than feldspar. If a significant proportion of the total Eu in a magma is present as Eu^{2+}, anomalous behavior for Eu is to be expected in many phases. This reflects both the difference in charge between Eu^{2+} and the rest of the trivalent REE (including Eu^{3+}), and the larger size of Eu^{2+} compared to Eu^{3+}. The effects of this difference in size can in general be qualitatively estimated from the shape of the REE distribution. Phases that progressively exclude the lighter REE can be expected to discriminate against Eu^{2+}, and those which prefer the larger, lighter REE will accept Eu^{2+} over Eu^{3+}. This argument appears to be generally supported by the data presented in Figures 5 through 8.

The magnitude of Eu anomalies in mineral/melt distributions is a complex topic, dependent on many parameters. The magnitude of Eu anomalies in plagioclase have been shown experimentally to vary with the partial pressure of O_2 (which affects Eu^{2+}-Eu^{3+} equilibria) and temperature (Drake and Weill, 1975), and can also be expected to be a function of bulk composition as this affects the redox equilibria of Eu (Morris and Haskin, 1974). All of these dependencies appear to result from variations in the oxidation states of Eu, and points out that while many minerals have the potential to have Eu anomalies, the extent to which they exhibit them will depend strongly on physical and chemical conditions in the system in which they form.

Summarizing, the great value of REE compared to most other trace elements is their relative constancy of fractionation behavior for any one mineral. Factors controlling the absolute values of mineral/melt distributions can be rationalized in part, but quantitative determinations of the actual dependencies are few. Convincing measurement and application for detailed analysis still appears remote.

The relative fractionations of minerals thought to play critical roles in the petrologic development of plutonic rocks of batholiths commonly have distinctive patterns. Plagioclase, the dominant mineral in most of these rocks, displays the large positive Eu anomaly unique to feldspar. Differentiation processes involving this mineral will induce in the rock products characteristic Eu variations. The broad, concave-down REE distributions for common calcic chain silicates are distinctive. Although garnet is not generally observed to be a primary phase in granitic rocks, it has broad stability fields for such compositions at high pressure (e.g. Green and Ringwood, 1968). The very strongly fractionated REE distributions for garnet will impart strong fractionations to derived materials. A number of minor or trace minerals which concentrate REE may also play significant roles. They have not been investigated systematically prior to this work. The study of REE variations in orogenic igneous suites, when integrated with broader based investigations, promises to provide important constraints on petrogenetic processes controlling their evolution.

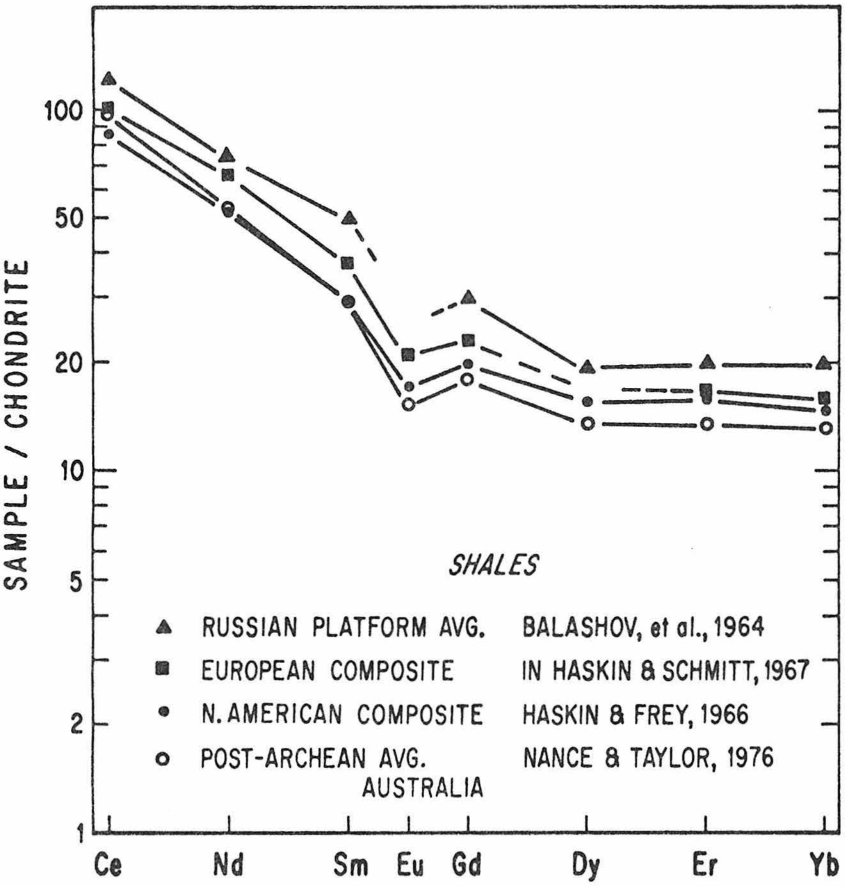
DISTRIBUTION OF REE IN THE CRUST

The principal source materials of granitic batholiths are within the upper mantle and crust. Evaluation of the roles which known crustal sedimentary and igneous materials might play in the generation of batholiths necessitates the characterization of their REE properties, and these are addressed below. This includes a brief discussion of selected studies of granitic rocks which are useful for comparison to the observations of this thesis.

Crustal Sedimentary Rocks

Much of the early work involving REE has focused on sedimentary rocks on the assumption they were representative of the earth's crust, or at least the part of the crust exposed to weathering. The characteristic REE fractionation pattern of shale composites analyzed in the early investigations has been confirmed by subsequent studies of composites of rocks from continental areas over the entire globe (Fig. 9). The relative abundances of heavy REE are similar to chondrites, but at levels 10 to 20 times higher. The light REE are strongly enriched to approximately 100 times chondrites for Ce, and a distinct negative Eu anomaly is present. Studies of individual sedimentary rocks, including such diverse lithologies as carbonates, sandstones, and greywackes as well as shales have shown considerable variation in absolute abundances in the large majority of samples analyzed, but only minor deviation from the relative abundances found in composite shales (Haskin, Wildeman, Frey, Collins, Keedy, and Haskin, 1966). Haskin and his coworkers conclude that REE fractionations in sedimentary rocks have little depen-

FIGURE 9. REE patterns of shale composites and averages from four continents. Note the similarity in their REE abundances and fractionations.


SAMPLE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
SHALES
▲ RUSSIAN PLATFORM AVG. BALASHOV, et al., 1964
■ EUROPEAN COMPOSITE IN HASKIN & SCHMITT, 1967
● N. AMERICAN COMPOSITE HASKIN & FREY, 1966
○ POST-ARCHEAN AVG. AUSTRALIA NANCE & TAYLOR, 1976

dence on bulk composition, and that the REE patterns of shale composites are representative of average crustal sedimentary rocks. These shales represent mature sedimentary materials derived from areally extensive source terrains. The uniformity of their REE characteristics must be attributed to either remarkably uniform source terrains or to contributions from diverse but very well mixed sources which have similar proportions in continental masses.

Several workers have observed that Eu may be systematically enriched in Precambrian sedimentary rocks, particularly Archean rocks (Wildeman and Haskin, 1973; Jakes and Taylor, 1974; Nance and Taylor, 1977). This feature has been interpreted by some of these authors as indicative of a different crustal composition during the Archean. However, in as much as most of the Archean samples are rather immature sedimentary rocks, comparison to post-Archean shale composites may not be appropriate. The more immature a sediment is, the more likely that it will reflect the chemical characteristics of its source materials as chemical weathering and physical sorting have had minimal effects. For example, the Devonian greywackes of the Baldwin Formation, Australia, appear to reflect their volcanic sources (Chappell, 1968; Nance and Taylor, 1977).

Crustal Igneous Rocks

The flat to somewhat light REE depleted patterns characteristic of mid-ocean ridge tholeiites (Frey, Haskin, Poetz, and Haskin, 1968; Kay, Hubbard, and Gast, 1970) also appears in early, close to trench volcanics of modern oceanic island arcs (Gill, 1970; Jakes and Gill, 1970; Ewart, Bryan and Gill, 1973). Similar patterns (compare in Fig. 10) are observed from the oceanward parts of arcs developed on continental crust (Japan: Philpotts, Martin, and Schnetzler, 1971 and Yajima, Higuchi, and Nagasawa, 1972) and rocks that may have formed in a comparable environment but are now incorporated into a continent (e.g. Arth and Hanson, 1972; Jahn, Shih, and Murthy, 1974).

Considerable variation in light REE characteristics have been observed across island and continental margin arcs, some of which have been closely correlated to the well established petrologic variations. The light REE depleted patterns of the tholeiitic portions of the arcs give way to the moderately to strongly light REE enriched patterns of high alumina (calc-alkaline) (Fig. 11) and alkalic types (Figi: Gill, 1970; Japan: Masuda, 1968 and Philpotts et al., 1971; Chilean Andes: Lopez-Escobar, Frey and Vergara, 1977 (tholeiites not represented)). It is important to note, however, that little variation in abundance or fractionation of the heavy REE is observed. Exceptionally, two heavy REE depleted and fractionated Chilean andesites were reported by Lopez-Escobar et al. (1977) (also Fig. 11) from the very northern part of the arc they studied. These samples appear to be in

FIGURE 10. REE patterns of some tholeiitic rocks.

Horizontal pattern: Field for most mid-ocean ridge basalts from several ocean basins (Kay et al., 1970), which may be considered characteristic of these basalts.

Vertical pattern: Field for young tholeiitic basalts and andesites of the Japanese arc (Philpotts et al., 1971).

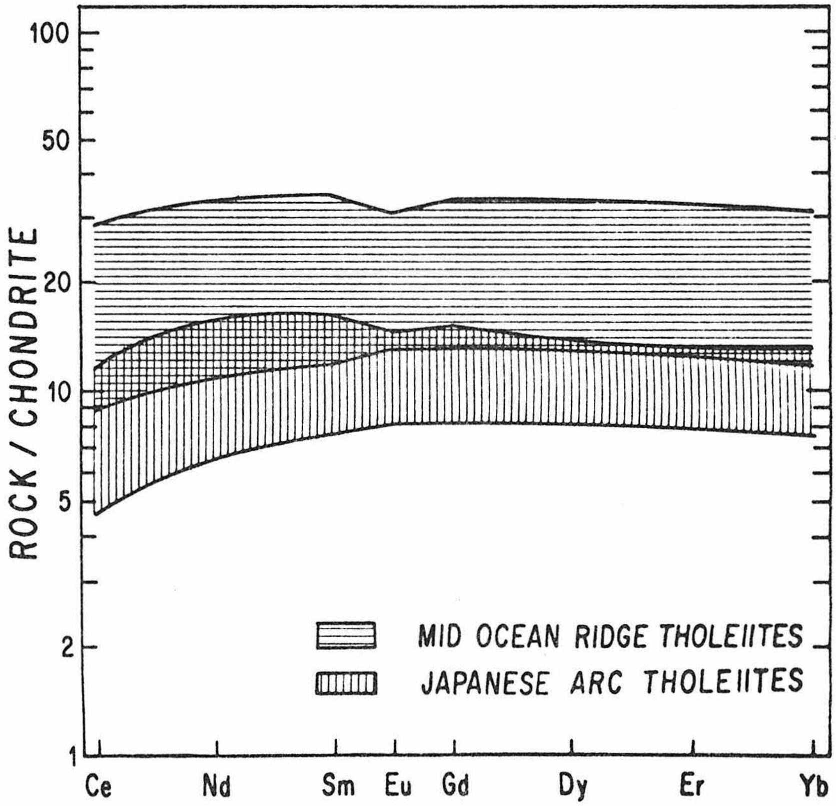
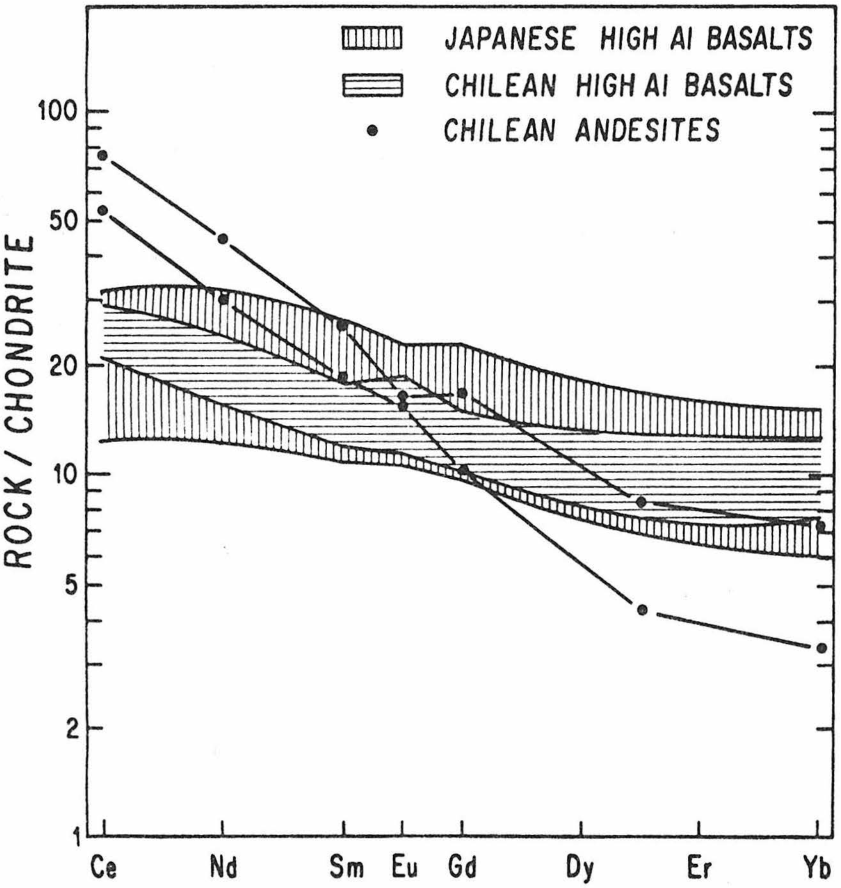

100
50
20
10
5
2
1
ROCK / CHONDRITE
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
MID OCEAN RIDGE THOLEIITES
JAPANESE ARC THOLEIITES

FIGURE 11. REE patterns of some calc-alkaline volcanics which appear characteristic of the circum-Pacific region.

Horizontal pattern: Field for 6 Chilean high alumina basalts (Lopez-Escobar et al., 1977).

Vertical pattern: Field for many high alumina basalts from Japan (Masuda, 1968; Philpotts et al., 1971).

Additionally, 2 Chilean andesites from High Andean volcanoes (Lopez-Escobar et al., 1977).

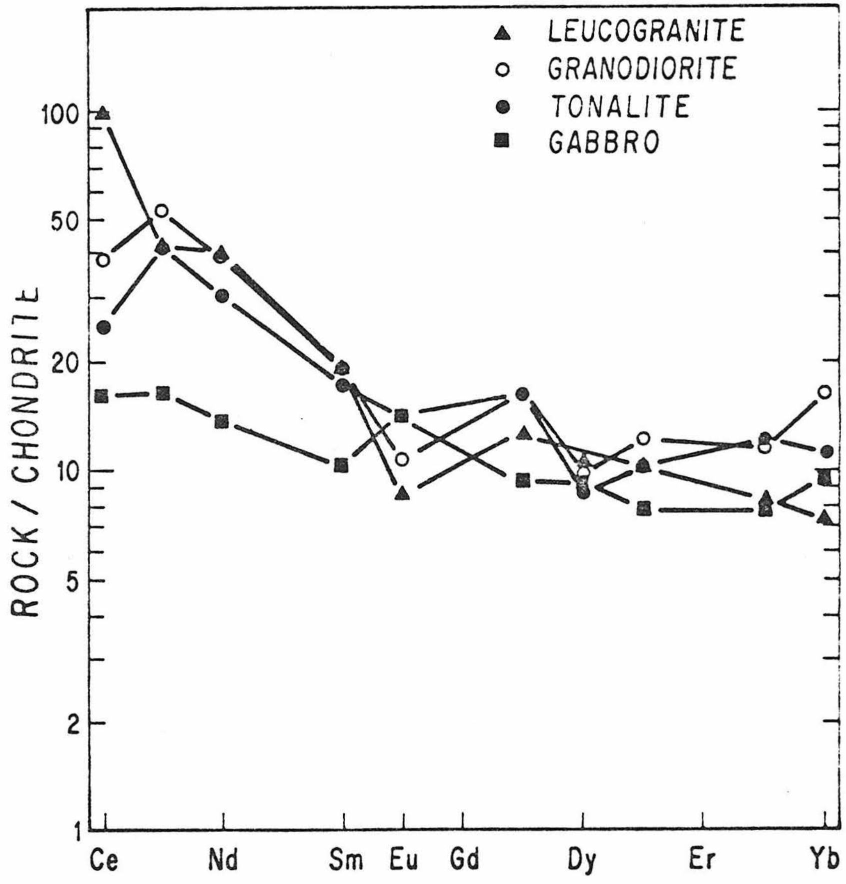

JAPANESE HIGH Al BASALTS
CHILEAN HIGH Al BASALTS
CHILEAN ANDESITES
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

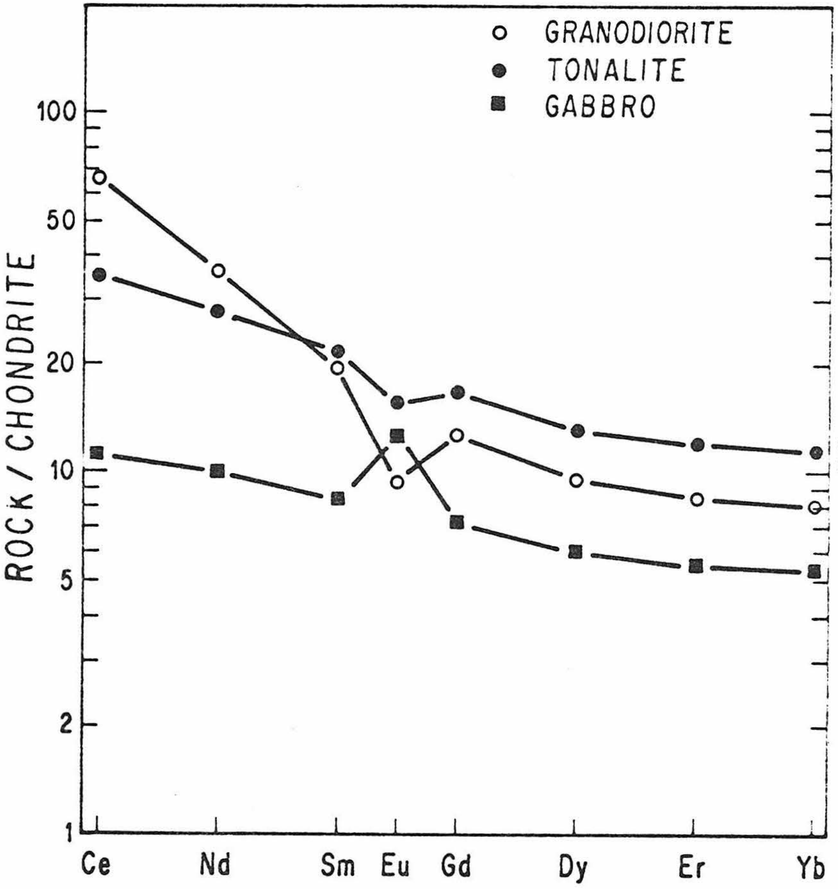
a comparable relative position in a west-east sense as many of the other samples. Searching further for geographic trends in continental margin arcs, limited studies in the Cascade Range of the northwestern U.S. (Condie and Swenson, 1973; Helmke and Haskin, 1973; Condie and Hayslip, 1975) have not revealed systematic variations with geographic position and in general surprisingly little variation with composition (some light REE enrichment in the more SiO_2-rich lithologies).

Studies of several granitic terrains have been reported. In a pioneering study, Towell, Winchester, and Spirn (1965) analyzed four samples representative of the western Peninsular Ranges: a gabbro, tonalite, granodiorite, and leucogranite. These data are presented in Figure 12a, and are accompanied in Figure 12b by analyses from the present study of three of the same samples. Although the analytical procedures for determining REE concentrations were in their infancy in the early 1960's, and detail is lost to the scatter of points, the general features of the REE patterns of these rocks as reported by Towell et al. have been confirmed. All of the rocks have a fairly flat heavy REE pattern at roughly 10 times chondrites, and a progressive enrichment in light REE and a relative depletion of Eu through the suite.

Haskin et al. (1968) reported REE contents of several composites of intermediate rocks and granites grouped by SiO_2 content, without details of location or sampling. These compositions parallel the patterns for shales, with negative

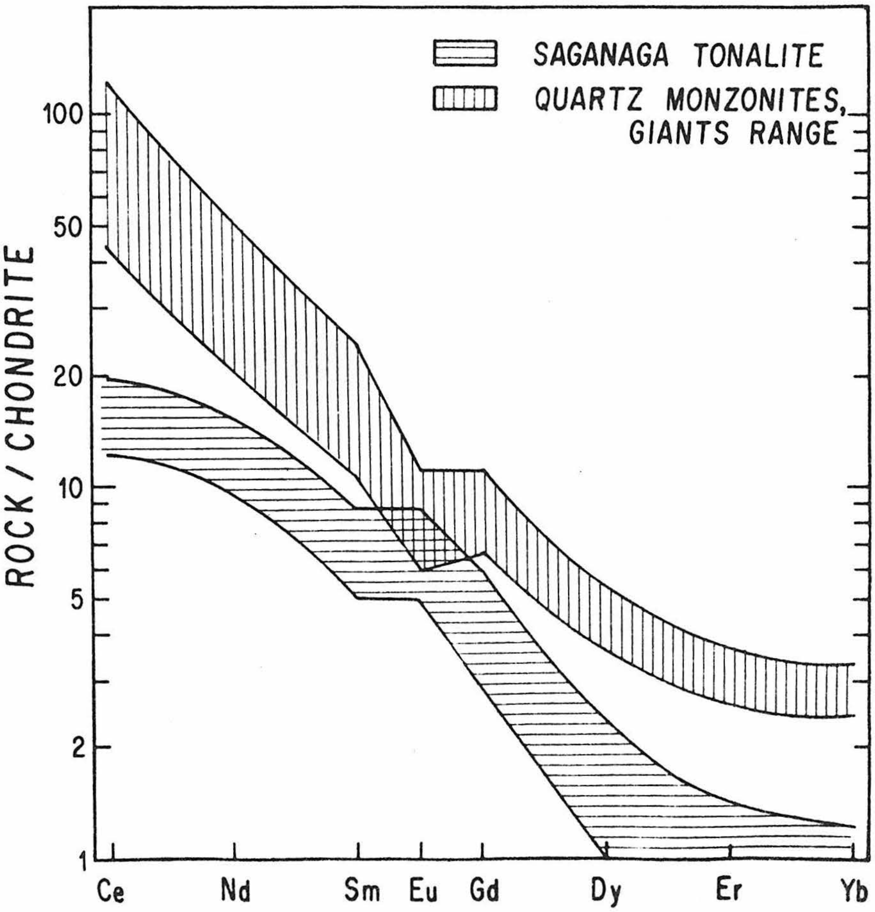
Figure 12. a) REE patterns of four representative lithologies from the northwestern Peninsular Ranges batholith, southern California (Towell et al., 1965). These are a gabbro from near Pala, a tonalite from near Val Verde, a leucogranodiorite from near Temecula, and a coarse leucogranite from Mt. Rubidoux. b) REE patterns of three of the four sample in Figure 12a, redetermined in this study. Sample numbers are 8 (gabbro), 1 (tonalite), and 3 (leucogranodiorite).


▲ LEUCOGRANITE
○ GRANODIORITE
● TONALITE
■ GABBRO
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb


○ GRANODIORITE
● TONALITE
■ GABBRO
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

Eu anomalies increasing with SiO_2 content. A detailed study of a granitic suite of Archean age from northeastern Minnesota-northwestern Ontario (Arth and Hanson, 1972, 1975) demonstrated that most of the ancient granitic rocks of this area possess fractionated REE patterns (Fig. 13). The total REE content increases in the more alkaline lithologies. Large negative Eu anomalies are present only in some quartz monzonites. These authors also reported a REE analysis of a trondhjemitic plutonic rock from the Klamath Mountains, northern California, which has a fractionated pattern very similar to Archean rocks they studied of comparable composition. Subsequent studies of the Mesozoic Sierra Nevada batholith (Frey, Chappell, and Roy, 1978) have indicated at least some other portions of this batholith are characterized by highly fractionated REE patterns. The limited sampling of this batholith leaves open to question the possible existence of regional variations. Both flat and fractionated heavy REE patterns have been observed in some granitic rocks from New England (Buma, Frey, and Wones, 1971), but without any apparent spatial relationship. The only sampling of a granitic terrain which reports suggestions of systematic regional variation in REE patterns is that of Barker, Arth, Peterman, and Friedman (1976) for four middle Proterozoic siliceous igneous rocks in the southern Rocky Mountains. From south to north, REE patterns vary from flat to highly fractionated and depleted in the heavy REE. These authors have also noted

FIGURE 13. REE patterns of some granitic rocks from northeastern Minnesota-northwestern Ontario. Fields are shown for the Saganaga tonalite and some quartz monzonites from the Giants' Range (Arth and Hanson, 1975).


SAGANAGA TONALITE
QUARTZ MONZONITES, GIANTS RANGE
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

complementary variations in other aspects of the chemistry of these rocks.

It may be noted from this summary of selected studies that igneous rocks with heavy REE depleted and fractionated patterns are not generally observed in island arcs, and only occasionally encountered in modern continental margin arcs, but are fairly abundant in granitic terrains. This contrast is poorly understood. The variability in REE patterns observed among and within granitic terrains is likely to be due in part to the complex historical evolution of most of these orogenic belts. How much variation occurs within simple batholithic belts and its significance have not been established. In formulating this study, it was anticipated that a detailed investigation of the REE characteristics of a comparatively simple magmatic belt already known to display marked asymmetries in geochemical and petrologic properties, as is observed in the Peninsular Ranges batholith, would squarely address these problems.

CHAPTER 2

OBSERVATIONS ON REE VARIATIONS

REE IN MINERALS

INTRODUCTION

One of the most important considerations in a study of trace element variations in igneous rocks is an understanding of the distribution of these elements in the mineral assemblages observed in the rocks. With such information it is possible to analyze the behavior of trace elements during crystallization and related differentiation processes near the level of emplacement. Integration of the distribution of trace components, mineral and bulk compositional relationships, paragenetic sequences deduced from petrographic observations, and information from experimental phase equilibria studies can lead to reasonable constraints on high level differentiation processes.

Towards this end, the distribution of the REE among the minerals of a representative granodiorite from the eastern part of the batholith has been studied in detail. Supplementing these data are additional REE determinations on two plagioclases and one hornblende from tonalites located on the west side. This work follows that of Towell et al. (1965) who determined the REE contents of some minerals in the rocks they studied from the northwestern Peninsular Ranges batholith.

REE DISTRIBUTION AMONG MINERALS IN A GRANODIORITE

Description of the Granodiorite

The granodiorite chosen for this study (sample 19) is located at the eastern margin of the Peninsular Ranges, 25 km SE of Palm Springs, near the town of Indio, California. The rock is a hornblende-biotite granodiorite which in hand specimen displays conspicuous large subhedral hornblende grains and well distributed green sphene grains. The rock is medium grained, foliated, with an overall hypidiomorphic-granular texture typical of the major plutons of the batholith.

Petrographically, subhedral plagioclase of up to 5 mm in dimension are not strongly zoned. Broad areas of the interior of grains are rather uniform at about An_{35-40}, with rims usually no less than An_{30}. A few plagioclase grains have a small amount of antiperthitic exsolution of alkali feldspar, localized as a small group of checkered patches in part of a grain. Quartz is anhedral, often as polygonal aggregates up to 5 mm in diameter. Alkali feldspar anhedra up to 3 mm in dimension commonly have a fine grained microperthite exsolution.

Biotite occurs as subhedral to anhedral plates, only occasionally grouped into small aggregates. Subhedral hornblende is pleochroic in light greenish brown to bluish green and has a 2V of $35-40^{\circ}(-)$. Sphene appears as independent, well formed euhedral to subhedral crystals sometimes exceeding 2 mm, and is often twinned; a few grains are anhedral and associated with hornblende and biotite. Allanite grains are euhedral, pleochroic in brownish and yellowish orange, somewhat irregularly zoned with brownish colors in the interior and yellowish colors toward the rim. An anhedral overgrowth of clear epidote commonly discontinuously or sometimes continuously surrounds allanite. This epidote overgrowth is always well developed against biotite. The boundary between epidote and allanite is sharp. Epidote also occurs as a secondary mineral in the limited alteration of plagioclase and mafic minerals, associated with trace amounts of chlorite and clay minerals. Apatite and zircon in euhedral habits are common, well distributed accessory minerals. Small quantities of myrmekite are often but not always developed in plagioclase where adjacent to alkali feldspar. Opaque minerals are very low in abundance, and the little there is occurs as inclusions in sphene.

A chemical analysis, norm, and a modal analysis based upon point counts of 10 thin sections are presented in Table 2. The chemistry of this rock is characterized by an intermediate silica content, and sizable amounts of all feldspar components. The calcic nature of this rock, typical of the Peninsular Ranges batholith, is reflected in the normative plagioclase composition of An_{40}. Sphene, epidote, apatite, allanite, and zircon are all significant accessories. Note, however, the large uncertainties in their abundances, estimated simply from counting statistics. The abundance of these minerals appears to vary greatly among thin sections.

Electron microprobe analyses of typical grains of

TABLE 2

CHEMICAL ANALYSIS		NORM (CIPW)		MODAL ANALYSIS (WGT. %)	
SiO_2	66.45	Qz	23.1	Quartz	26.1
TiO_2	.65	Or	18.8	Plagioclase	45.6
Al_2O_3	15.89	Ab	28.5	Alkali feld.	11.2
Fe_2O_3	1.08	An	19.2	Biotite	10.9
FeO	2.67	Cpx	.44	Hornblende	4.82
MgO	1.56	Opx	6.78	Sphene	.65±.06
MnO	.06	Mt	1.58	Epidote	.52±.06
CaO	4.15	Ilm	1.24	Apatite	.11±.03
Na_2O	3.34	Ap	.37	Allanite	.077±.02
K_2O	3.15			Zircon	.010±.010
P_2O_5	.17				
H_2O^+	.64				
TOTAL	99.81				

B. Chappell, ANU, Analyst

10 thin sections
18,041 pts.

most of the minerals in this rock, determined on two thin sections, are given in Appendix II. The most important features of these analyses are the pargasitic nature of the hornblende, the K-rich composition of the alkali feldspar, and the high REE and Th contents of the allanite.

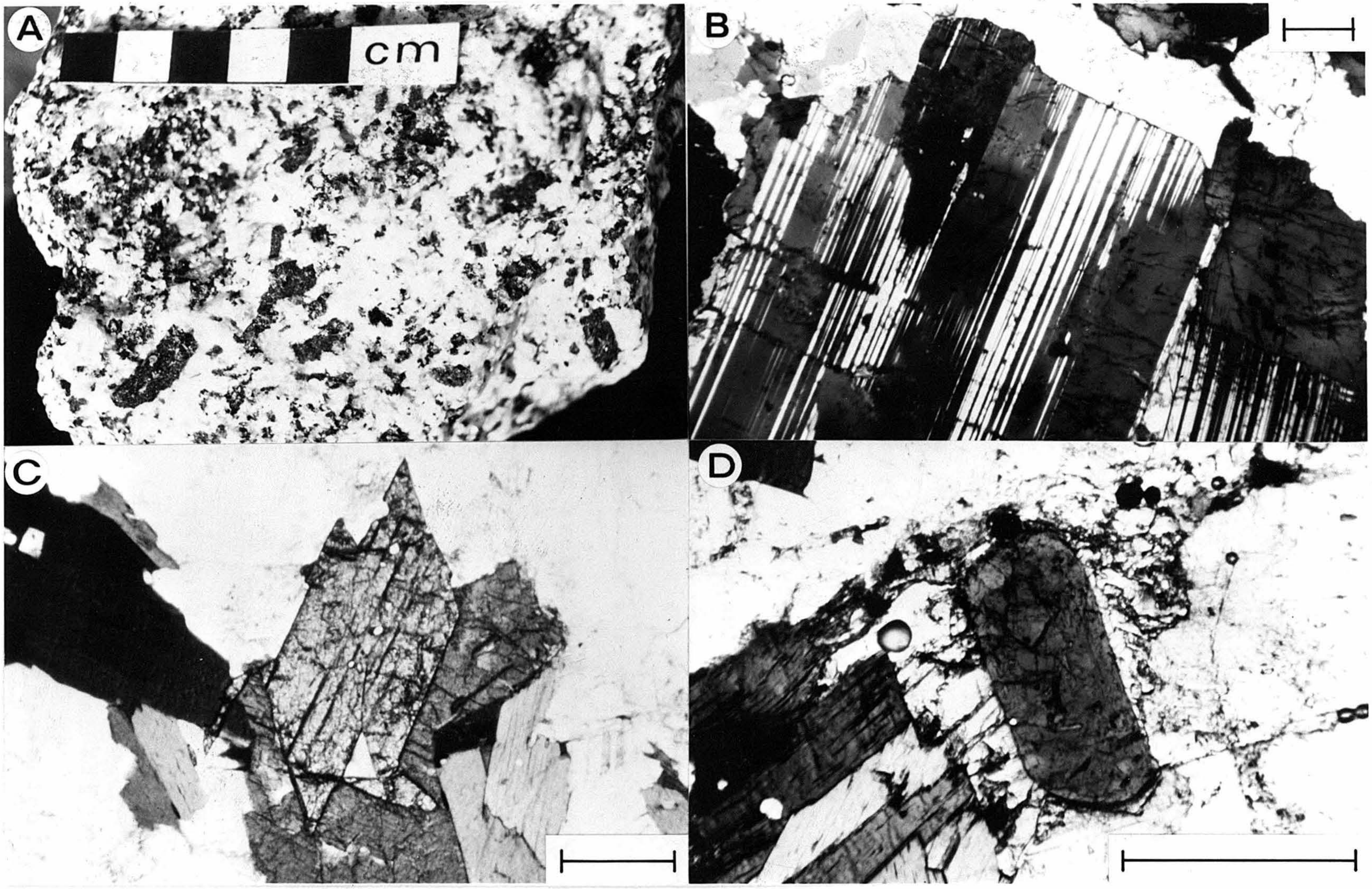
On the basis of textural relationships, plagioclase, allanite, sphene, and opaques are the earliest minerals to crystallize. Plagioclase forms large tabular grains with very few inclusions. There is no evidence for early formed crystals of plagioclase from either cumulate or restite origin. Plagioclase does preserve euhedral faces against some minerals (commonly alkali feldspar) but appears to compete for space during late growth with many others (Fig. 14b). In nearly all observed instances allanite (Fig. 14d) and larger sphene grains (Fig. 14c) have preserved their euhedral faces against all other minerals. This suggests that these minerals began crystallization early and underwent only limited late growth. Opaque grains are essentially restricted to the interiors of occasional sphene grains, indicating possible nucleation of sphene on epaque minerals early in the crystallization sequence.

Some of the largest hornblende grains often have a high proportion of small, randomly oriented inclusions of quartz (anhedral), apatite, zircon, and sphene. Minute rods of apatite in particular are abundant and occur mostly in the quartz inclusions. Most of the hornblende has clear inner-

FIGURE 14. Photograph and photomicrographs of a granodiorite near Indio, California (sample 19). Bar scales in all the photomicrographs are 0.5 mm.

a) Hand specimen of granodiorite, displaying large subhedral hornblende grains. b) Twinned plagioclase, with weak zoning. Euhedral crystal face against alkali feldspar (top center), and anhedral face against quartz (top left. Crossed polarizers. c) Three euhedral sphene grains, the two largest of which form a twin. Hornblende has grown around these grains. Plane light.

d) Euhedral allanite, discontinuously rimmed by clear epidote. Note core to rim zoning in color of allanite. Plane light.


A
cm
B
C
D

most interiors. Toward the rims, inclusions of small quartz and subhedral biotite become common. Hornblende is seen to mold around euhedral to subhedral sphene (Fig. 14c) and plagioclase, but also appears to compete for space with plagioclase and biotite. These relationships argue for an extended crystallization interval for hornblende, overlapping with the intervals for most of the other minerals. Alkali feldspar is often interstitial, as is some quartz and biotite.

REE Abundances in Minerals

REE analyses of nine separated minerals (plagioclase, alkali feldspar, biotite, hornblende, sphene, epidote, apatite, allanite, and zircon) are presented in Table 3, and their chondrite normalized REE patterns are displayed in Figure 15. The whole rock pattern is also plotted for comparison. Among these minerals there is a greater than 10^5 range in concentration of all REE except Eu. These enormous variations in abundances and fractionations are an indication of the strong crystal chemical control on the behavior of these elements, and a secondary measure of the purity of the prepared mineral separates.

Before discussing these data, several comments about contaminants in the mineral separates and the representativeness of these separates is warranted. Grain counts of grain mounts of most of the separates are presented in Appendix III. Most are better than 99% pure. The integrity of REE analyses of minerals with low REE contents such as feldspar and bio-

TABLE 3

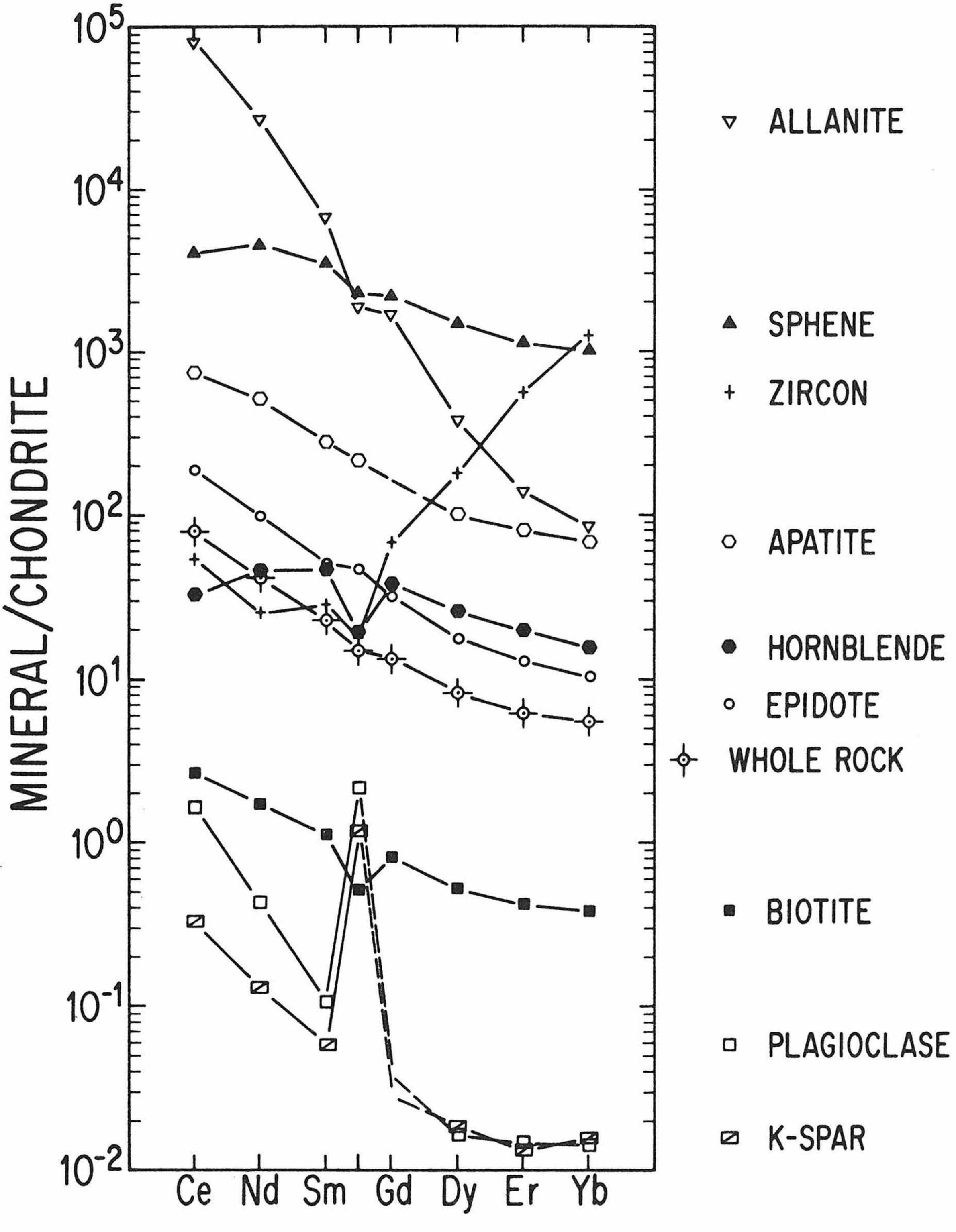
RARE EARTH ELEMENT CONCENTRATIONS OF MINERALS IN A GRANODIORITE IN PPM

	Plagioclase*	Alkali Feldspar	Biotite		Hornblende
			Finely Ground	Coarse	
Ce	1.36	.264	2.20	9.04	22.5
Nd	.252	.0769	1.03	4.17	27.5
Sm	.0200	.0112	.221	.698	8.67
Eu	.155	.0821	.0377	.208	1.375
Gd	---	---	.213	.63	9.74
Dy	.00552	.0060	.170	.412	8.29
Er	.00308	.0029	.0913	.225	4.18
Yb	.00301	.0033	.0792	.195	3.18

	Sphene	Epidote	Apatite	Allanite	Zircon
Ce	3305	152	596	66,560	42.3
Nd	2680	58.5	302	16,060	14.9
Sm	655	9.45	52.9	1,260	5.40
Eu	165	3.38	15.2	133.4	1.27
Gd	564	8.15	---	460	17.4
Dy	470	5.67	31.7	118.4	56.9
Er	237	2.69	17.1	28.5	116
Yb	207	2.10	13.9	17.4	253

*Corrected for the presence of 27% quartz in plagioclase concentrate.

FIGURE 15. Chondrite normalized REE patterns of minerals in the granodiorite near Indio. The whole rock REE pattern is also plotted for comparison. The biotite pattern is that of the finely ground separate. Note the large range in REE abundances among these minerals, and the high concentrations in allanite and sphene.


MINERAL/CHONDRITE
10^5
10^4
10^3
10^2
10^1
10^0
10^{-1}
10^{-2}
Ce
Nd
Sm
Gd
Dy
Er
Yb
▽ ALLANITE
▲ SPHENE
+ ZIRCON
⬡ APATITE
⬢ HORNBLENDE
○ EPIDOTE
WHOLE ROCK
■ BIOTITE
□ PLAGIOCLASE
▨ K-SPAR

tite could be seriously compromised by even minute amounts of a contaminant such as allanite, sphene, or apatite. Great care was taken in the preparation of mineral separates of all phases, insuring very pure separates in most cases. Mineral grains containing minute inclusions, however, might not be eliminated. In the case of the feldspars, additional steps were taken to rid the separates of small apatite inclusions. The feldspars were very finely ground by hand in an agate mortar, then acid washed and thoroughly rinsed to dissolve any apatite present.

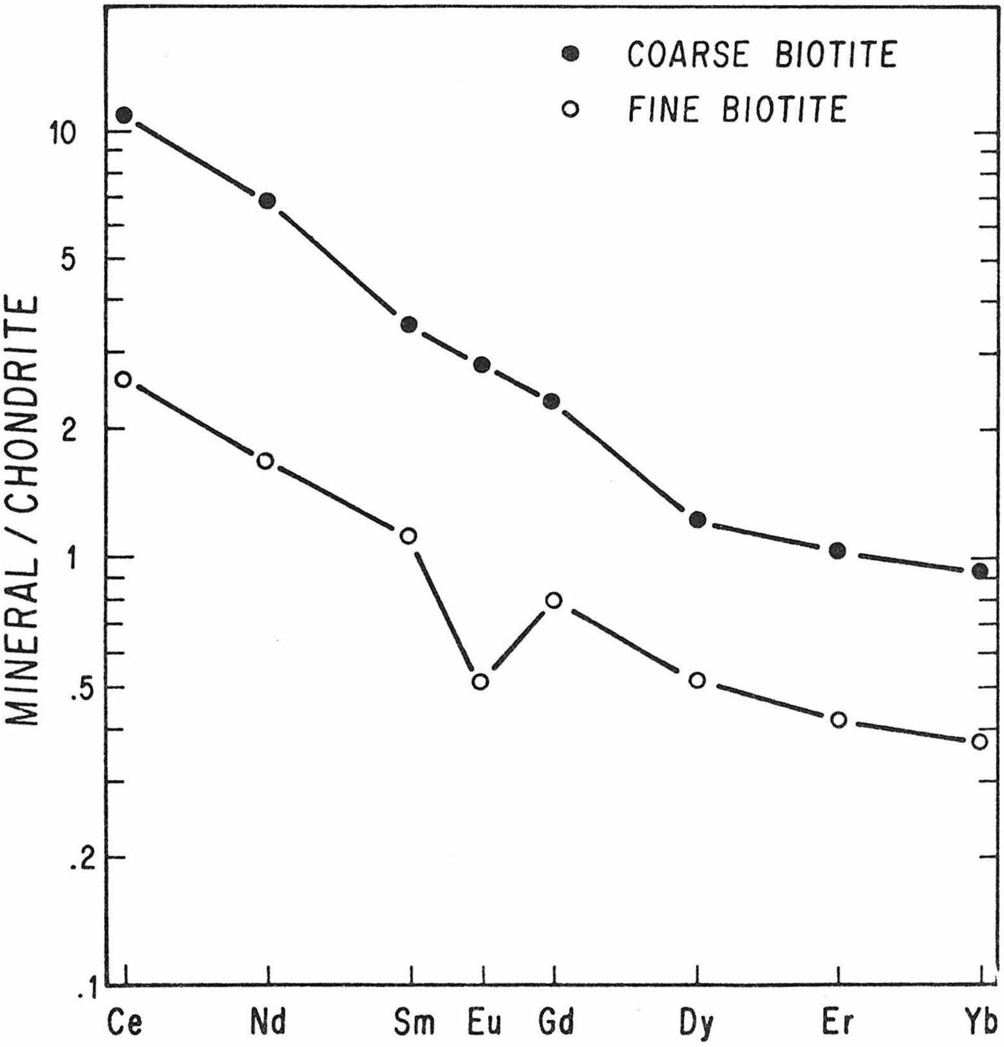
Plagioclase, K-feldspar, and biotite have concentrations of REE much lower than the whole rock, even for Eu in the feldspars. Both plagioclase and K-feldspar have very similar patterns with steep slopes in the light REE, large positive Eu anomalies, and very low concentrations of heavy REE. Biotite, epidote, and apatite have patterns which parallel the whole rock, but with abundances varying from less than .1 to greater than 10 times these values. Hornblende is the only major phase containing higher concentrations of REE (except Ce) than the whole rock. Zircon is sharply heavy REE enriched. The two remaining accessory phases, allanite and sphene, are very REE-rich. The allanite displays extreme light REE enrichment, as would be expected. This allanite separate contains 6.6% Ce and 1.6% Nd. It is somewhat surprising that the allanite and epidote do not have similarly sloped patterns. The sphene is more uniformly enriched, at levels 100 to 200 times whole rock values for most REE.

No mineral other than feldspar could be responsible for the large positive Eu anomalies, reducing the question to how much of the trivalent REE could be due to contamination. The concave-up shape of the feldspar patterns are unlike any other mineral analyzed. The contaminant would have to be a mixture of minerals in roughly the same proportions for both feldspars, or a single mineral but only in quantities sufficient to contribute to part of the pattern. In the latter case, the values measured for some of the REE would be maximum values of the amounts actually present in the feldspars. Evidence arguing against contamination by other mineral phases controlling the abundance and fractionation of the REE in the feldspar separates centers about the similarity of the plagioclase and K-feldspar patterns. These two feldspars have different positions in the paragenetic sequence, and also a contrast in density. The nature and amount of potential contaminants would be expected to vary, yet a strong similarity in the patterns exists.

As biotite contains numerous inclusions, this separate was ground as fine as practical (pass 200 mesh, retain 325 mesh) during preparation so that the density and magnetic techniques could be most effective. Since most but not all inclusions could be eliminated, a coarse biotite separate (retain 30 mesh) was prepared in which no effort was made to eliminate inclusions. The coarse biotite separate is composed of the larger biotite grains in the rock, but the fine biotite is not representative of the finer biotite in the rock. It is composed of all biotite components which have been finely ground during preparation. The coarse biotite separate has a generally similar REE pattern, but more than twice the abundances of all REE (Fig. 16). The distinct negative Eu anomaly displayed for the fine grained biotite cannot be correlated with the patterns of other minerals in the rock. The disappearance of this Eu feature in the coarse biotite probably reflects a contaminant with a pattern similar in slope to the fine biotite, a positive Eu anomaly, and much higher concentrations of REE due to its optically limited abundance. Based on these criteria, apatite appears to be the most reasonable contaminant, and its common presence as an inclusion in biotite is supportive. A simple calculation indicates that less than 1% apatite added to fine biotite would be sufficient to produce most of the characteristics of the coarse biotite pattern. A small amount of epidote and/or plagioclase may also contribute. Due to the great differences in the amount of inclusions present in these two separates, it is concluded that the fine biotite pattern is a close approximation to the actual REE character of the biotite.

One source of uncertainty in the use of these mineral data is the possibility that each prepared mineral separate is not representative of the mean of the compositional

FIGURE 16. Chondrite normalized REE patterns for coarse and finely ground biotite.


● COARSE BIOTITE
○ FINE BIOTITE
MINERAL / CHONDRITE
10
5
2
1
.5
.2
.1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

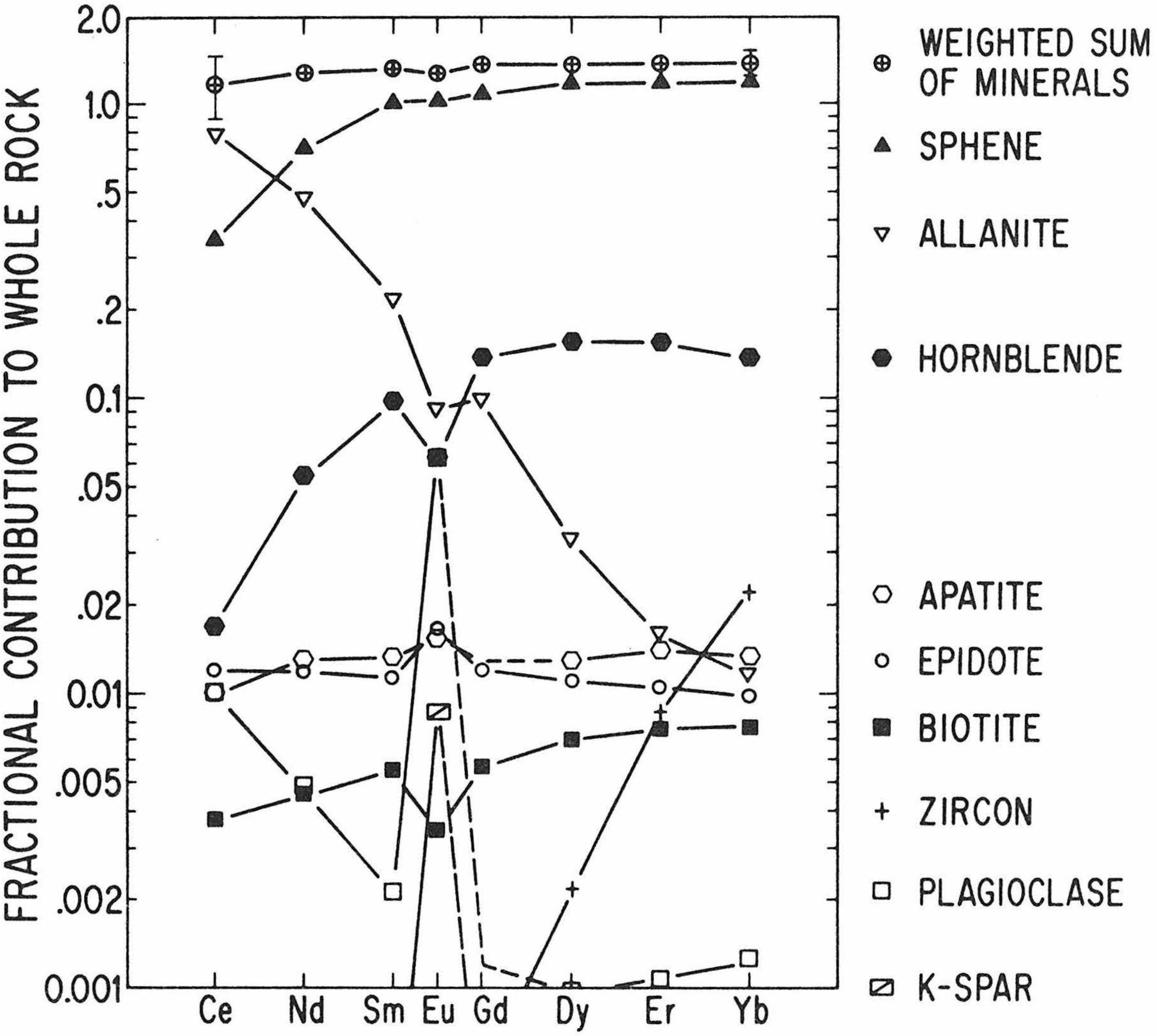
variations of each mineral species. The preparation of a mineral separate presents many opportunities to introduce biases. In many if not most cases, a compromise has been made between maintaining a representative nature in the separate and its purity.

Mineral Contributions to the Whole Rock REE

The fraction of each REE contributed to the measured whole rock abundances are obtained by dividing the REE abundances of the mineral by the whole rock value, which is then weighted by its abundance in the rock. These values, along with the composite reconstructed rock calculated from the sum of the minerals, are plotted in Figure 17. This diagram clearly indicates that accessory allanite and sphene are the major sites for REE in this rock. Before further discussion, the deviation of the sum of the minerals from the whole rock value must be addressed.

The sum of the minerals plots as a line deviating only slightly from flatness, indicating that this sum follows the overall REE fractionation pattern measured in the whole rock. The sum of the minerals, however, exceeds the whole rock abundances by some 20% to 40%, far exceeding the analytical uncertainties in the measurement of the REE. There are several possible causes for this discrepancy: inaccurate modal abundances, nonrepresentative mineral separates, and a nonrepresentative whole rock powder. The whole rock powder is considered to be representative despite the fact that 200 mg of

FIGURE 17. Fractional contribution of each mineral to the whole rock REE abundances in the granodiorite near Indio. A composite reconstructed rock, calculated from the sum of the minerals, is also plotted. Error bars on Ce and Yb for this sum indicate the uncertainty arising simply from the uncertainty in the modal abundances of allanite and sphene. Allanite contributes most of the uncertainty in Ce, whereas sphene does for Yb. Note that most of the REE in this rock are present in sphene and allanite. The shapes of the patterns on this diagram are equivalent to whole-rock normalized values.


FRACTIONAL CONTRIBUTION TO WHOLE ROCK
2.0
1.0
.5
.2
0.1
.05
.02
0.01
.005
.002
0.001
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
⊕ WEIGHTED SUM OF MINERALS
▲ SPHENE
▽ ALLANITE
● HORNBLENDE
⬡ APATITE
○ EPIDOTE
■ BIOTITE
+ ZIRCON
□ PLAGIOCLASE
▨ K-SPAR

powder (a few thin section volumes) was analyzed as the total material processed for the whole rock powder was greater than 500 g. Both nonrepresentative mineral separates and inaccurate modal analyses can be important in explaining the discrepancy, but only for the few phases making major contributions to the whole rock REE (allanite, sphene, and perhaps hornblende). As the other phases contribute the order of 1% or less, even large uncertainties in their modal abundances or REE concentrations will not have a significant effect on the sum. Sphene must be the major source of the discrepancy in the middle and heavy REE as it alone appears to contribute more than the measured whole rock. The conservatively estimated uncertainty in the modal abundances of sphene and allanite can account for much of the discrepancy, as indicated in Figure 17.

Plagioclase, biotite, and K-feldspar together account for roughly 1% of each REE except Eu, to which plagioclase alone contributes about 7%. Apatite and epidote each provide about 1% of each REE, and zircon can contribute up to a few percent of the heaviest REE. Hornblende is the only major mineral containing significant quantities of REE, up to 15% of the heavier REE. The remainder, from 80% to more than 90% of each REE, are in sphene and allanite. Sphene is of dominating importance in the middle to heavy REE, and allanite is increasingly important in light REE.

The strong localization of REE in sphene and allanite prompted further electron probe determinations on these minerals, with particular emphasis on zoning. Three allanites and one sphene from one thin section were probed from rim to rim. Representative results are presented in Appendix II, and summarized in Figures 18 and 19. The allanites have unusually low totals. Wavelength scans did not reveal other unanalyzed elements. The low totals are apparently due to either inaccurate probe standards for REE and Th, imprecise analysis of these standards due to low abundance, or incorrect absorption factors related to REE and Th (both major element absorption of REE and Th radiation and REE and Th absorption of major element radiations).

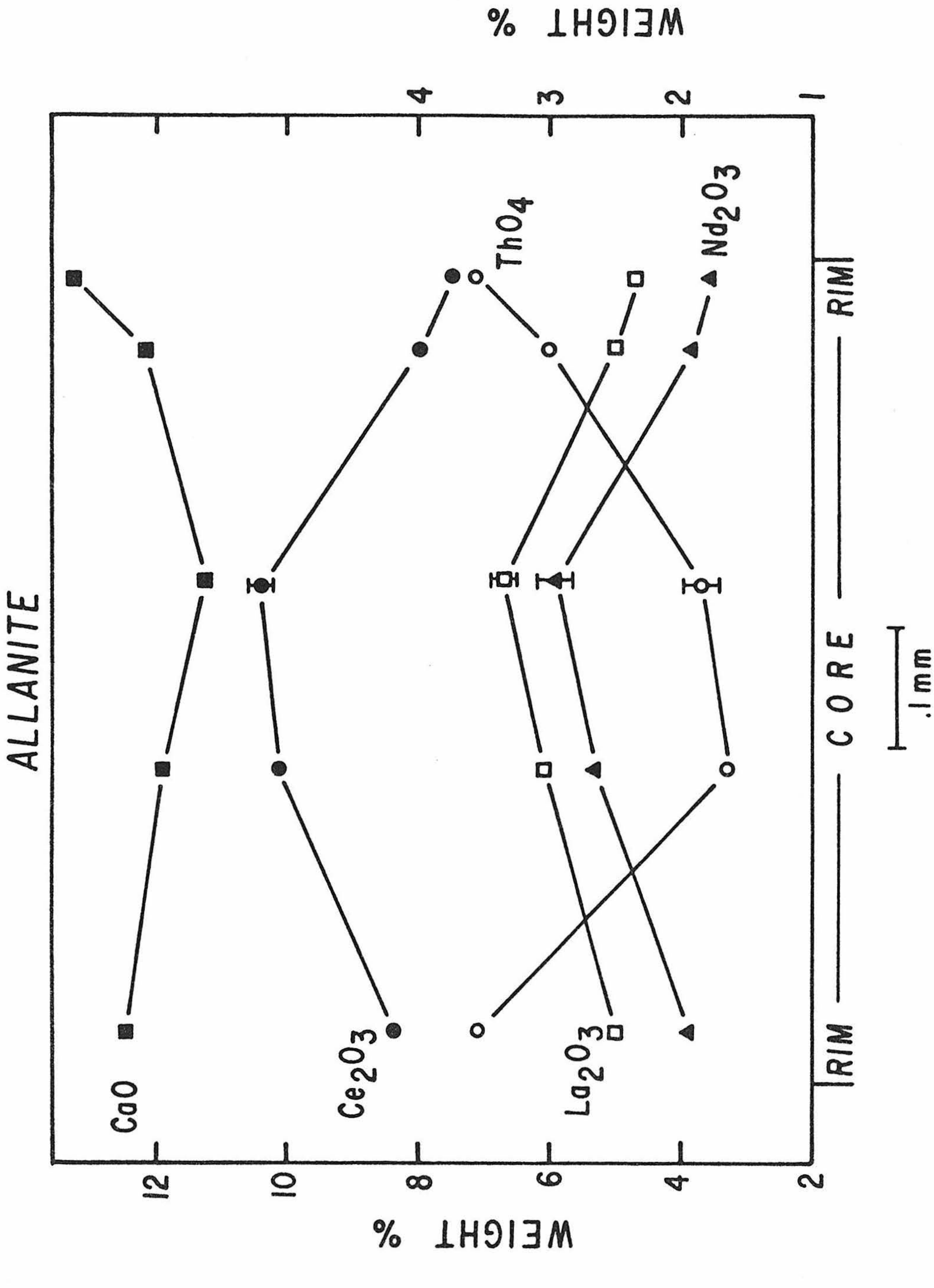
Both allanite and sphene display core to rim decreases in Ce contents of 25% to 50%, respectively. In allanite, this is accompanied by core to rim increases in Ca and Th contents. The anticorrelation of Ce and Th is consistent among the three allanite grains analyzed, suggesting a charge-balanced substitution in allanite:

$$\mathrm{Ca}^{2+} + \mathrm{Th}^{4+} = 2\ \mathrm{REE}^{3+}$$

This relationship is shown rigorously with a plot of the sum La+Ce+Nd (the most significant REE components in allanite) against Ca+Th (Fig. 20). These data clearly indicate this zoning within allanite is not a substitution toward an epidote endmember (REE and Th-poor, Ca-rich). This is consistent with the petrographic observation that epidote is a distinct phase which has formed as an overgrowth on allanite.

FIGURE 18. Core to rim zoning of La_2O_3, Ce_2O_3, Nd_2O_3, ThO_2, and CaO in an allanite grain in the granodiorite near Indio. Error bars indicate uncertainty arising simply from counting statistics on the electron microprobe. This zoning pattern has also been recognized in the two other allanite grains analyzed in the same thin section.

FIGURE 19. Core to rim zoning of Ce_2O_3 in sphene in the granodiorite near Indio. Error bars indicate uncertainty arising simply from counting statistics on the electron microprobe.


ALLANITE
WEIGHT %
12
10
8
6
4
2
WEIGHT %
4
3
2
1
CaO
Ce_2O_3
La_2O_3
ThO_4
Nd_2O_3
RIM
CORE
RIM
.1 mm

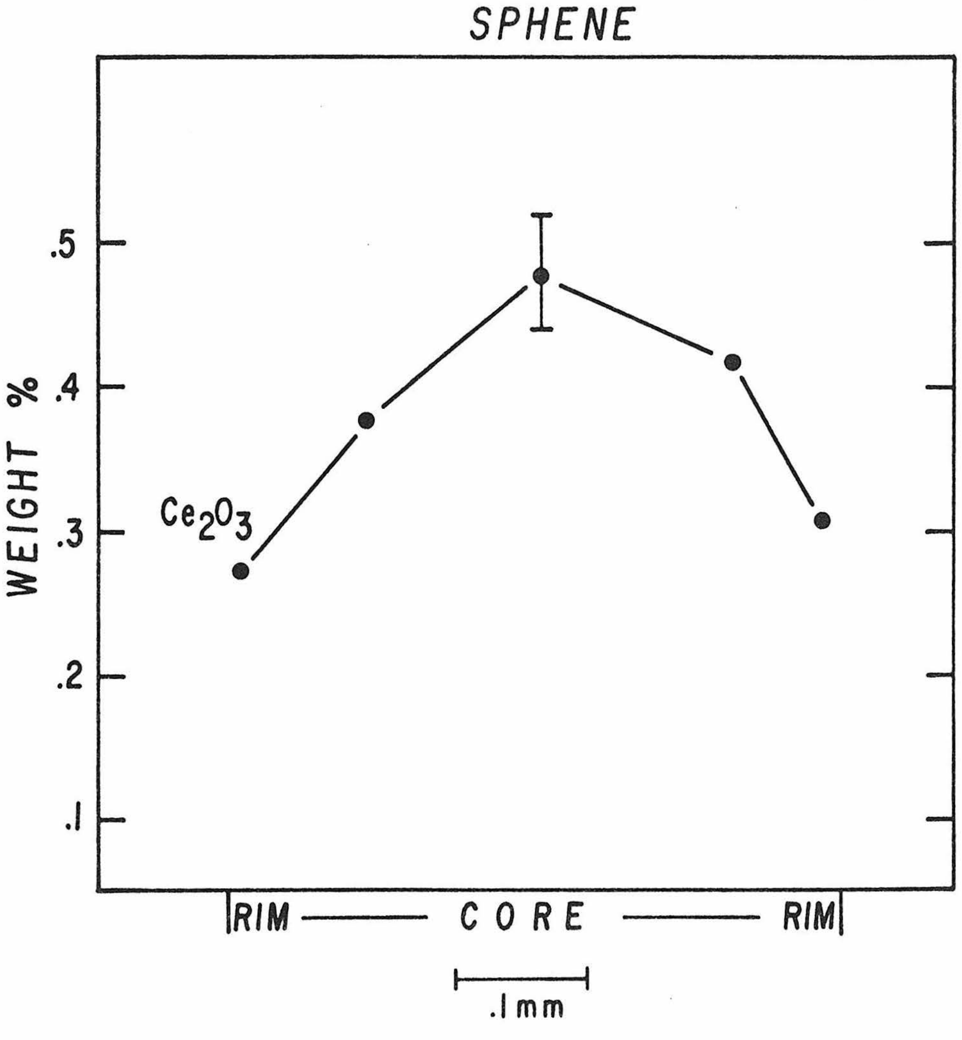
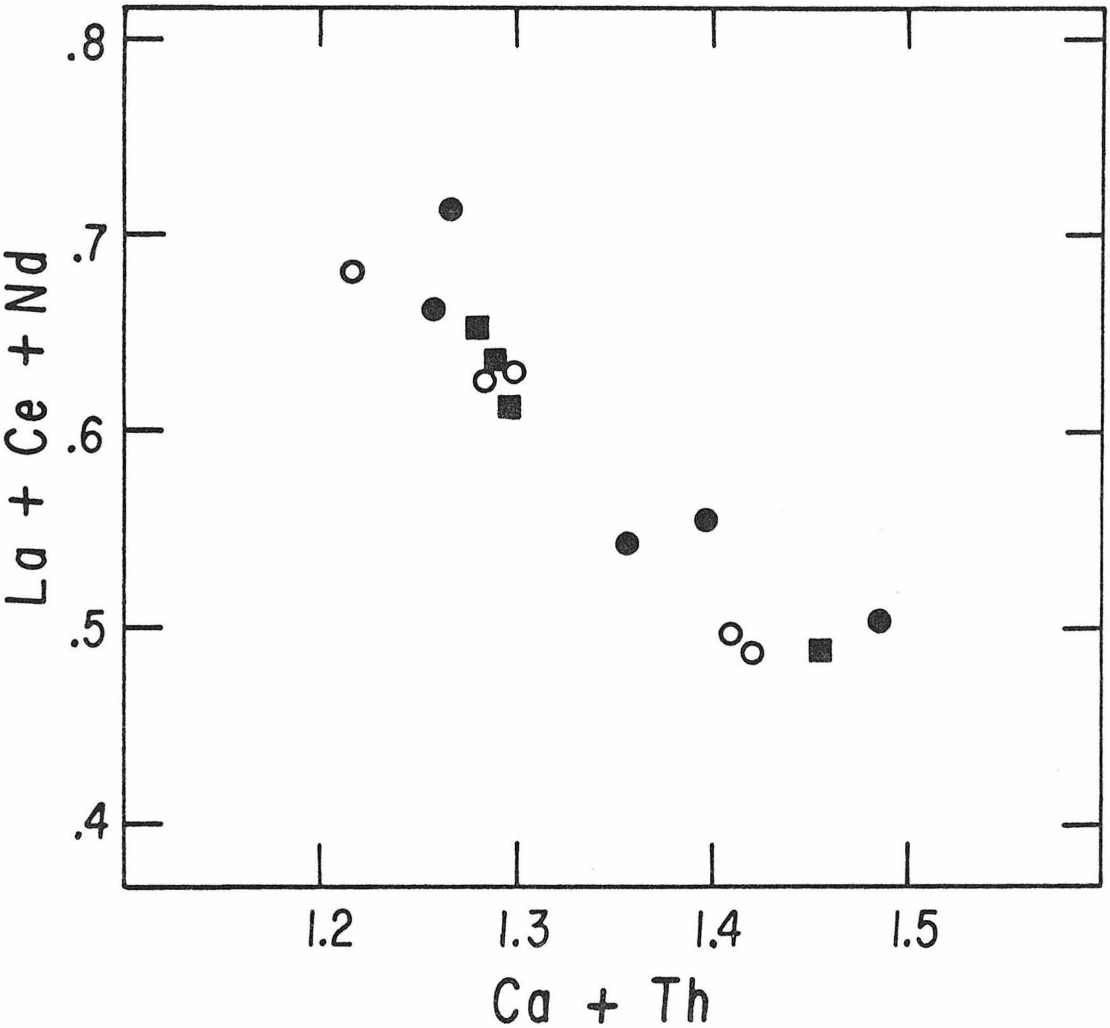

SPHENE
WEIGHT %
.5
.4
.3
.2
.1
Ce_2O_3
RIM
CORE
RIM
.1mm

FIGURE 20. La+Ce+Nd versus Ca+Th, in formula proportions, for 3 zoned allanite grains in the granodiorite near Indio. The three different symbols indicate different grains in the same thin section. Note the correlation within and among these grains, suggesting a coupled substitution.


.8
.7
.6
.5
.4
La + Ce + Nd
1.2
1.3
1.4
1.5
Ca + Th

REE IN MINERALS IN OTHER ROCKS

In addition to this detailed study of the distribution of REE among minerals in one rock, several other mineral analyses have been made: a plagioclase and hornblende from a typical Bonsall-type tonalite near Val Verde (sample 1), and a plagioclase from the seriate porphyritic unit of the tonalitic San Jose pluton (sample 50), Baja California (Murray, 1978).

The tonalite near Val Verde is typical of this major rock type of Larsen (1948) and has been described in detail by Osborn (1939). Subhedral andesine plagioclase is strongly zoned and hornblende, also subhedral, coexists without reaction with biotite. Quartz is anhedral, and K-feldspar is interstitial. The San Jose tonalite is distinguished by abundant subhedral andesine plagioclase which is very strongly zoned. Hornblende and biotite appear interstitial to plagioclase, as is quartz to a greater degree. A petrographic description, mode, and chemical analysis for each of these rocks is provided in Appendix I.

The REE analyses of these minerals are presented in Table 4, and their chondrite-normalized patterns in Figures 21 and 22. Both feldspars have low concentrations of REE with fairly steep slopes in the light REE and large positive Eu anomalies. The hornblende is enriched in REE with a substantial negative Eu anomaly. For most of the middle to heavy REE, this hornblende has concentrations more than five

TABLE 4

RARE EARTH ELEMENT CONCENTRATIONS OF SELECTED MINERALS IN TWO TONALITES IN PPM

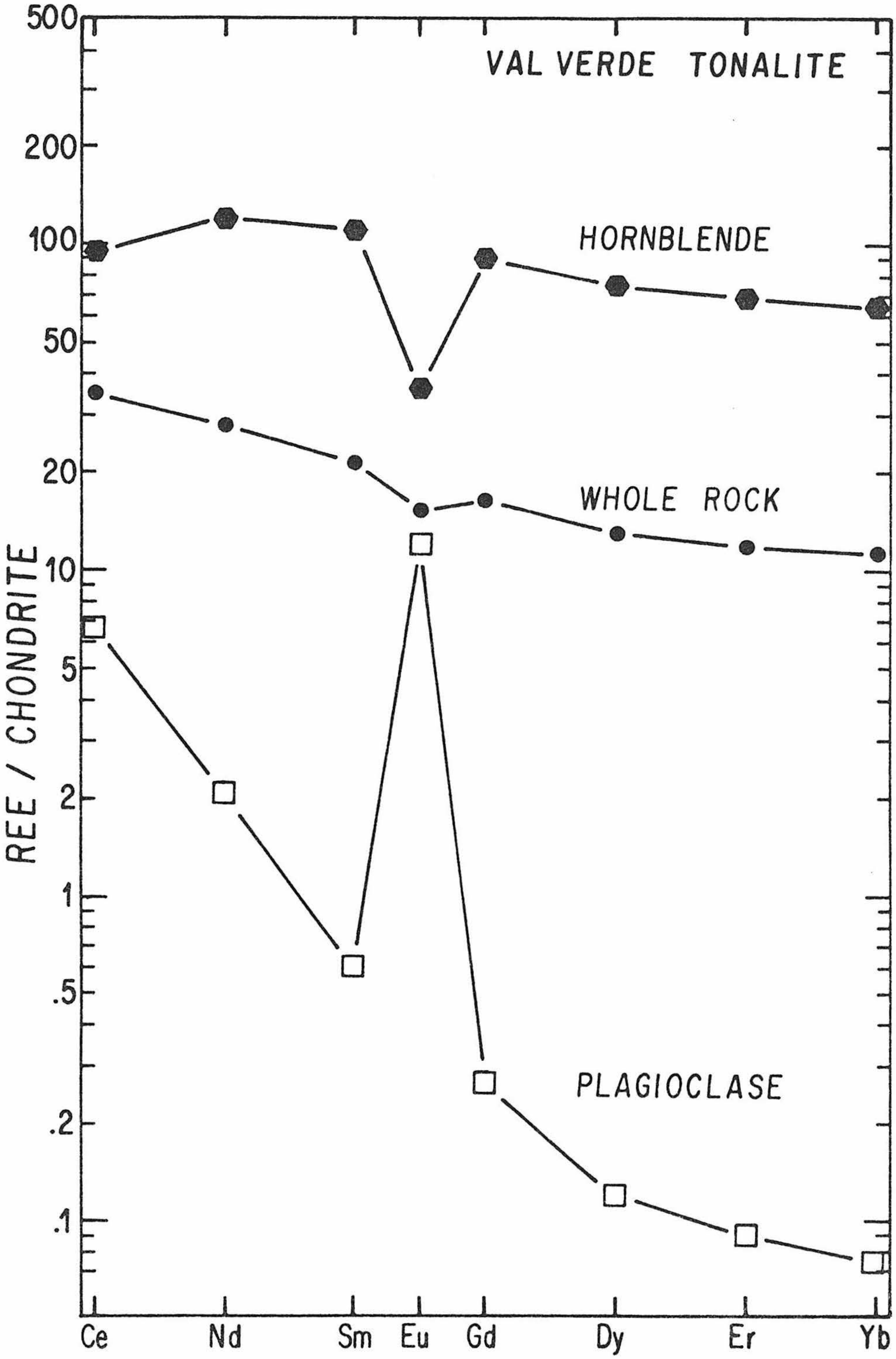
	SAMPLE 1		SAMPLE 50
	Plagioclase*	Hornblende	Plagioclase**
Ce	5.51	78.0	4.43
Nd	1.27	73.8	1.04
Sm	.118	21.6	.0857
Eu	.882	2.65	.416
Gd	.071	23.9	.0400
Dy	.040	24.5	.0176
Er	.020	14.5	.0077
Yb	.017	13.3	.0066

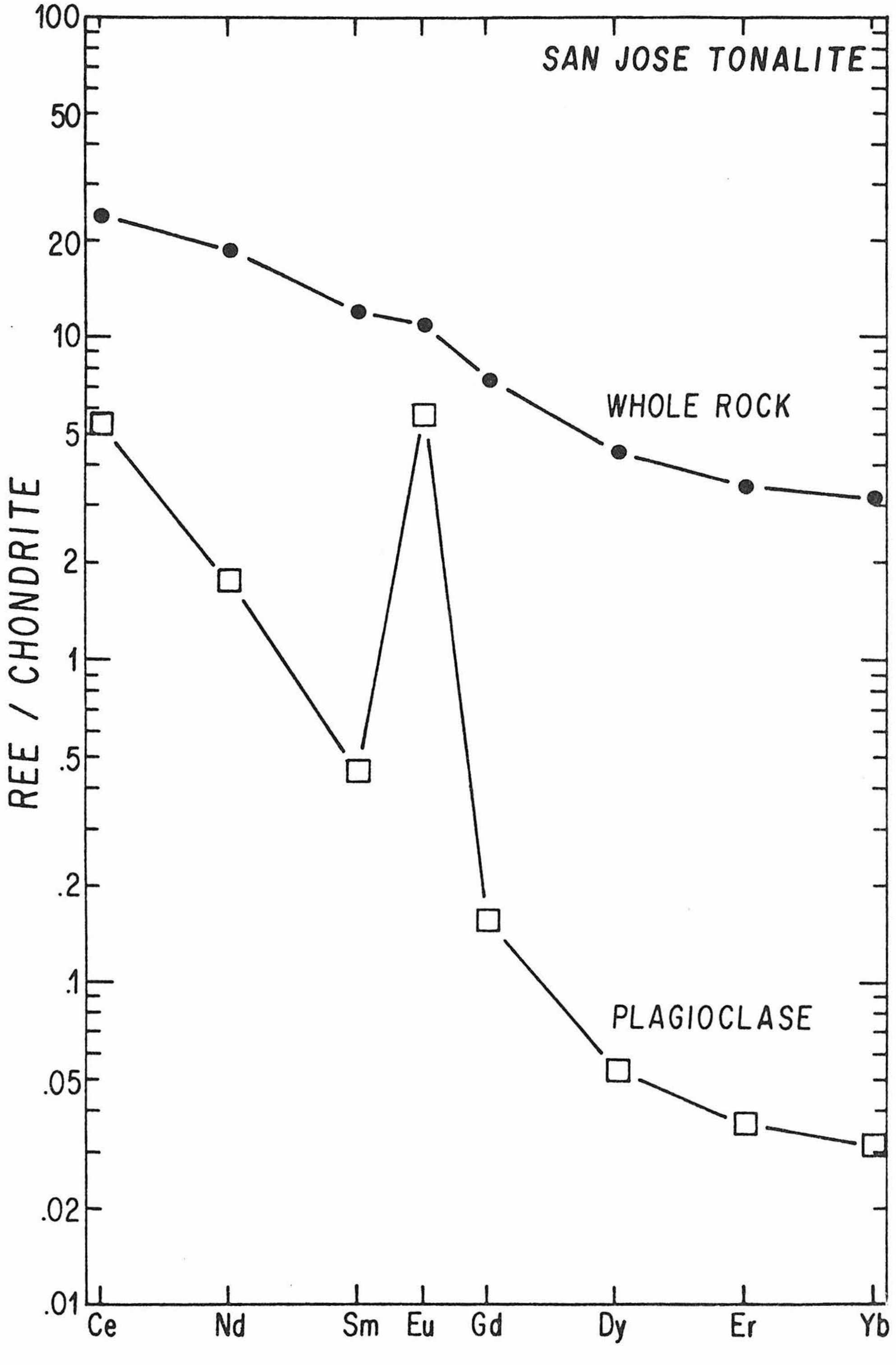
*Corrected for 10% quartz present in plagioclase concentrate

**Corrected for 30% quartz present in plagioclase concentrate

FIGURE 21. Chondrite normalized REE patterns for plagioclase, hornblende, and whole rock in the tonalite near Val Verde, Ca. (sample 1).

FIGURE 22. Chondrite normalized REE patterns for plagioclase and whole rock in a tonalite of the San Jose pluton, Baja California (sample 50).


VAL VERDE TONALITE
HORNBLENDE
WHOLE ROCK
PLAGIOCLASE
REE / CHONDRITE
500
200
100
50
20
10
5
2
1
.5
.2
.1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb


SAN JOSE TONALITE
WHOLE ROCK
PLAGIOCLASE
REE / CHONDRITE
100
50
20
10
5
2
1
.5
.2
.1
.05
.02
.01
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

times whole rock levels.

These data on the distribution of REE among minerals in rocks of the batholith place important constraints on high level crystal fractionation processes. Most significant in this regard is the dominating contributions of REE-rich accessories in the granodiorite, indicating that the crystallization and removal of these phases can profoundly modify the REE characteristics of the melt from which they grew. A detailed discussion of the significance of this and other implications of these data follows the presentation of the whole rock data.

REE IN WHOLE ROCKS

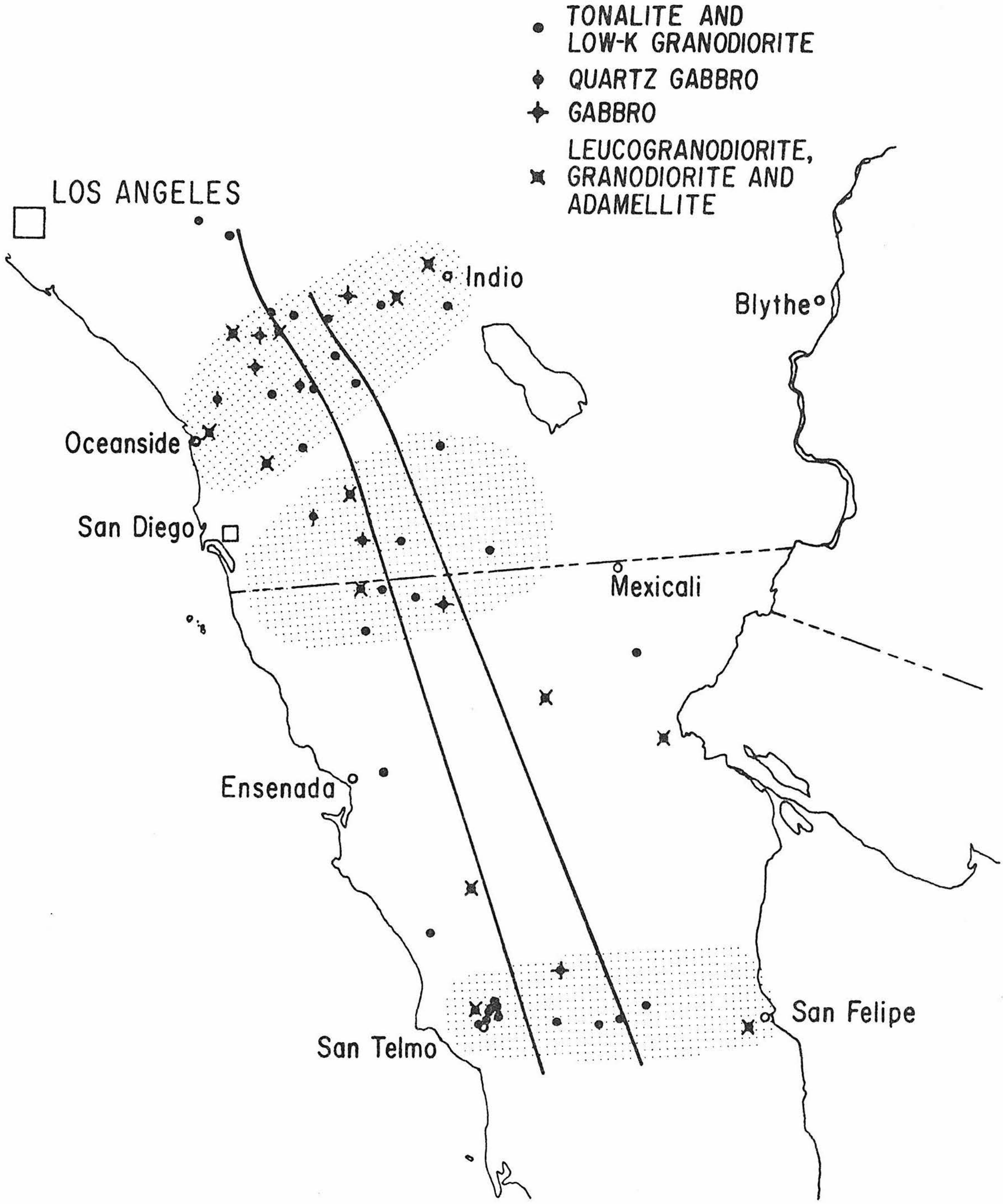
SAMPLING DESIGN

Selection of Samples

The selection of samples for this study has been strongly influenced by previous work. The transverse asymmetries in geochronologic and geochemical features of the batholith established by Silver and his coworkers have made the determination of variations in REE characteristics of intrusive rocks across the batholith an important objective of this work. Another major objective was to sample the spectrum of major lithologies within the batholith so that the processes responsible for its petrologic diversity could be studied. Most of the samples analyzed for REE are also those used in preceding and parallel investigations, and were collected and classified by Prof. Silver in the course of his petrologic studies of the batholith.

For the purposes of examining transverse variations, sampling was concentrated in three major profiles across the batholith (Fig. 23). The best studied profile is within southern California, trending ENE between the cities of Oceanside and Indio. A second profile extends east from San Diego and Tijuana on both sides of the border. A third profile well south of Ensenada reaches from San Telmo, east across the Sierra San Pedro Martir, to San Filipe on the gulf. The first and second profiles have largely merged in southern California, providing almost continuous coverage

Figure 23. The location of samples analyzed for REE in the Peninsular Ranges. The shaded regions indicate the location of three traverses of concentrated sampling across the batholith. The boundaries dividing the batholith into western, central, and eastern regions are shown as solid lines paralleling the length of the peninsula. The different lithologies sampled are indicated by symbols.


● TONALITE AND LOW-K GRANODIORITE
QUARTZ GABBRO
GABBRO
LEUCOGRANODIORITE, GRANODIORITE AND ADAMELLITE
LOS ANGELES
Indio
Blythe
Oceanside
San Diego
Mexicali
Ensenada
San Telmo
San Felipe

between them. A few other samples are widely distributed between and peripheral to these profiles and provide additional geographic coverage.

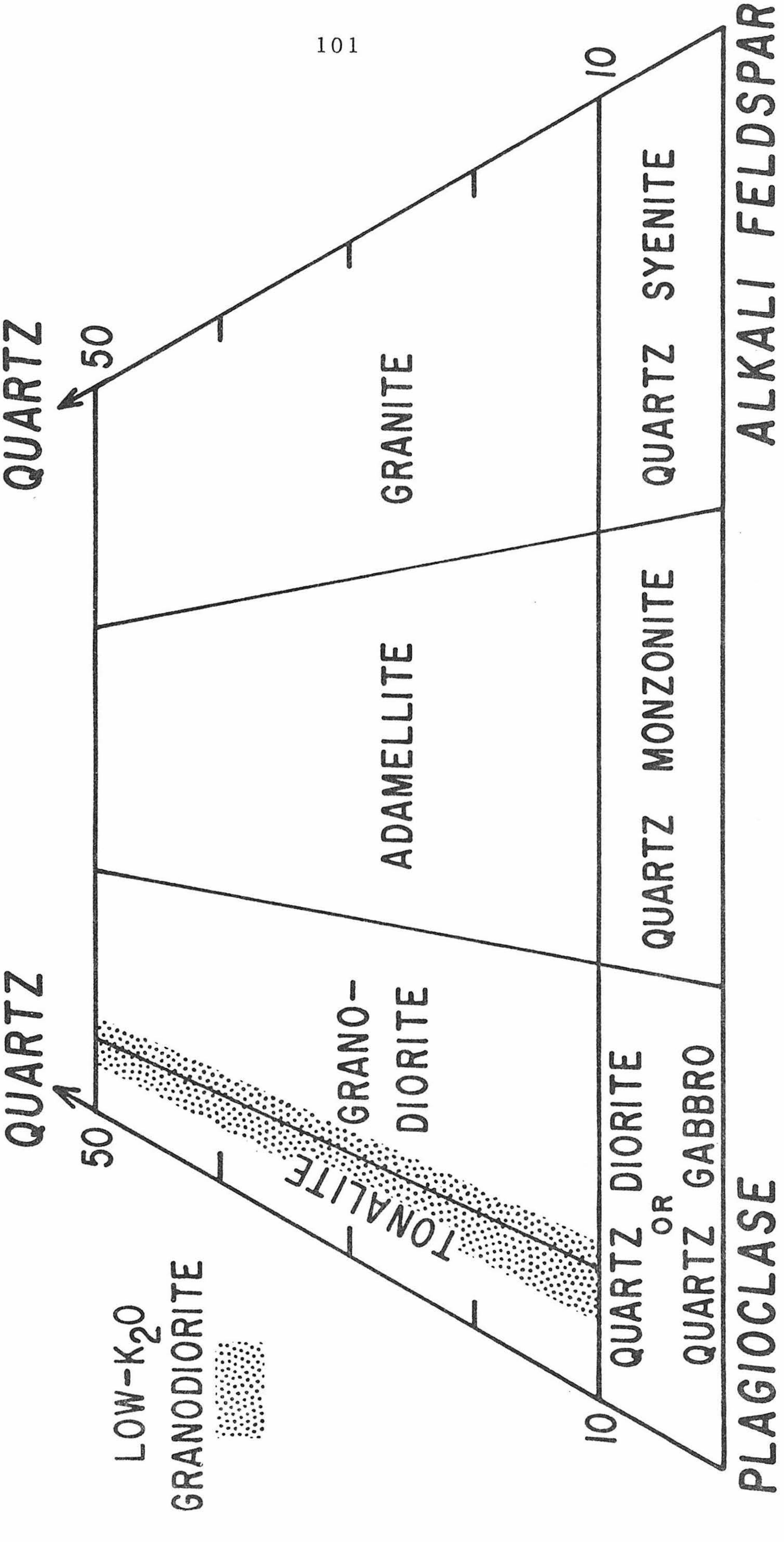
During the early course of this study, it became apparent that the REE data were naturally falling into three distinct elongate pattern groups related to geographic position transverse to the batholith: western, central, and eastern. The characteristics of these groups are presented in the following section. The areal distribution of these groups is displayed in Figure 23. The placement of the boundaries among these groups is based solely on the REE data presented here. These boundaries are not controlled by the much younger right lateral faults known within the Peninsular Ranges, and probably do not correspond to the precise plutonic boundaries.

For simplification in the presentation of these data, the sampled areas are arbitrarily divided at the international border, one in southern California and the other in Baja California.

Rock Classification

The rock classification used in this study follows the working classification of Prof. Silver, and is based on petrographic and chemical criteria. The petrographic framework of this classification is illustrated in Figure 24. The lithologies are distinguished by the relative abundance of plagioclase, alkali feldspar, and quartz. The distinction

Figure 24. A modal quartz-plagioclase-alkali feldspar ternary illustrating the rock classification used for quartz-bearing lithologies. An approximate shaded field for low-K_2O granodiorites is indicated, which overlaps the tonalite and granodiorite fields.


QUARTZ
50
QUARTZ
50
10
10
LOW-K_2O GRANODIORITE
TONALITE
GRANO-DIORITE
ADAMELLITE
GRANITE
QUARTZ DIORITE OR QUARTZ GABBRO
QUARTZ MONZONITE
QUARTZ SYENITE
PLAGIOCLASE
ALKALI FELDSPAR

between quartz diorite (rare) and quartz gabbro is made at a modal plagioclase composition of An_{50}. The name gabbro is applied to those rocks which are saturated or undersaturated in quartz. An inherent problem in this petrographic classification is that in many of these rocks, a large proportion of the K_2O is present in biotite, not alkali feldspar. This is usually the case in tonalites and low-K_2O granodiorites, which are granodiorites with roughly 5-12% alkali feldspar and K_2O contents in the approximate range of 2.00% to 2.75%. An approximate field for low-K_2O granodiorites is shown in Figure 24, and there is a continuous petrographic variation between them and the tonalites. Granodiorites have up to approximately 3.80% K_2O, and adamellites roughly up to 4.50% K_2O.

This classification provides a petrologically more useful framework for the description of the lithologic diversity of these rocks than is found in the more general classification of Streckeisen (1973).

A breakdown of the samples analyzed as to lithologic type and geographic region is presented in Table 5. Additionally, a listing of the samples and their rock type is given in Appendix V.

Lithologies Sampled

Tonalite is the most common lithology present in substantial amounts everywhere in the batholith. Accordingly, emphasis was placed on sampling tonalitic intrusives in or-

TABLE 5

LISTING OF SAMPLES BY GEOGRAPHIC REGION AND LITHOLOGY

WESTERN REGION					CENTRAL REGION			EASTERN REGION				
GAB	QTZ GAB	TON	LOW-K GRD	LEUCOGRD	GAB	TON	LOW-K GRD	GAB	TON	LOW-K GRD	GRD	AD
(Southern California)												
7	9	1	10	3		4	28	16	15	17	18	
8	12	2	13	6		5			26	20	19	
27	24	11		23		14			29			
		22		25		21						
(Baja California)												
	47	31	34	30	33	50	32		37		36	38
		35	48	39	49	51			53		54	
		40		41		52						
		42										
		43										
		44										
		45										
		46										

GAB = Gabbro
QTZ GAB = Quartz Gabbro
TON = Tonalite

LOW-K GRD = Low-K_2O Granodiorite
LEUCOGRD = Leucogranodiotite
AD = Adamellite

der that regional variations could be studied among rocks of similar bulk composition. Tonalites do display a range in modal and chemical composition, but have several unifying characteristics. Compare petrographic descriptions in Appendix I. These are feldspathic rocks containing roughly 50% to 60% zoned andesine feldspar. Hornblende and biotite in variable proportions are the dominant mafic components, together yielding a color index of 15±3 for most. Quartz makes up 20% to 25% of most tonalites, and alkali feldspar up to a few percent.

Consideration of various tonalites analyzed illustrates some of the differences among them. Sample 1 is representative of the Bonsall type of tonalite of Larsen (1948), a characteristic tonalite of the western region of the batholith (a petrographic description of this rock is provided in Appendix I). This rock was distinguished from other tonalites of the region (sample 2 from the Lakeview Mountains, for example) by its abundance of mafic inclusions. An unusual suite of more leucocratic tonalites from the San Telmo pluton (samples 42-46; 43 in Appendix I) are much higher in quartz and lower in mafic minerals than most tonalites but still contain an andesine feldspar. In the central region of the batholith, some tonalites have been sampled from very large apparent intrusive units such as the Cahuilla Valley pluton (Sharp, 1965)(sample 5; in Appendix I). This and the San Jose tonalite (sample 50; Appendix I) are examples of some of the more feldspathic tonalites of the batholith. In contrast to the western and central regions, the tonalites of the eastern regions usually display prominent euhedral to subhedral sphene, sample 29 (Appendix I) being particularly notable in this regard. Despite its deeper level of emplacement, this sample retains fine oscillatory zoning in plagioclase. In Baja California, sample 37 is gneissic and some of its original igneous texture is obscured.

Quartz gabbro as an intrusive unit is essentially restricted to the western region of the batholith. Some bodies are found to grade into tonalites in the field (Silver, per. comm.), and are distinguished by their more calcic feldspar (sodic labradorite), higher color index, and lower quartz content. A quartz gabbro from the San Marcos Mountains (sample 9; Appendix I) is a typical example of this rock type.

The term granodiorite covers a rather wide variety of, and probably not very closely related, lithologies. In the western region, particularly in southern California, most granodiorites are of the Woodson Mountain type of Larsen (1948). These are quite leucocratic granodiorites and commonly form large masses. Sample 23 (Appendix I) is from the type locality, and is characterized by a calcic oligoclase and abundant quartz. Samples 3,6,25,30 are similar rocks. Other western granodiorites are low-K_2O granodiorites, which resemble tonalites in the field and have andesine plagioclase (samples 10,13,34,48), and granophyric granodiorites (samples

10,41). Sample 10 is a low-K_2O granodiorite, but also leucocratic and granophyric.

In the central region, low-K_2O granodiorites grade into tonalites (Larsen, 1948). Samples 28 and 32 (32 in Appendix I) are two samples of one of the largest bodies in the Peninsular Ranges, the La Posta pluton of Miller (1935). As in most low-K_2O granodiorites, they have an andesine plagioclase and both biotite and hornblende. The granodiorites of the eastern region include both low-K_2O (samples 17,20,36,54) and normal granodiorites (samples 18,19; 19 in Appendix I). As has been noted, in hand specimen many of these resemble tonalites in all aspects except alkali feldspar content. Sample 19 is an example of such rocks, and is characterized by an andesine plagioclase and a color index of 14. An important textural feature of this and many other rocks of the central and eastern regions is the prominence of independent, euhedral sphene and to a lesser extent euhedral allanite grains.

Adamellites are generally uncommon in the western and central regions of the batholith, and none have been analyzed in the study. One sample of adamellite from the eastern region has been studied (sample 38).

The gabbros are among the most heterogeneous rock types of the batholith, and individual bodies are often composed of many related lithologies (e.g. Miller, 1937; Larsen, 1948). The various gabbro samples analyzed are discussed in more detail in the following section when the REE data are presented.

These samples appear to represent all of the major lithologies of the batholith, but not necessarily in proportion to their actual abundances. The location of samples by number is indicated on Figure 25.

ANALYTICAL RESULTS

All of the analytical determinations for REE in this study, including those on minerals already reported, were by isotope dilution mass spectrometry. A description of procedures, including sample preparation, dissolution and chemical separations, and mass spectrometry are provided in Appendix IV. The results are presented in tabular form (Table 6), and are discussed graphically here according to geographic region. The gabbros are discussed separately.

Southern California

Western Region

REE analyses of 3 quartz gabbros and 3 tonalites (Fig. 26) covering most of the western region include rocks mapped as the San Marcos gabbro, and the Bonsall and Lakeview tonalites (Larsen, 1948). These samples have a significant range in bulk composition, are from separate, distinct plutons over a wide geographic area, and have emplacement ages spanning 15 my (Silver, per. comm.). Yet, they display a remarkable similarity in REE patterns. All are near flat in the heavy REE with 11 to 15 times chondritic levels, and have a moderate relative enrichment in the light REE. Every sample has a modest (15-20%) negative Eu anomaly. The slight curvatures

Figure 25. The location of samples analyzed for REE with sample numbers. Closely spaced samples within the San Telmo pluton, Baja California are shown in an expanded array in the lower left of the diagram.

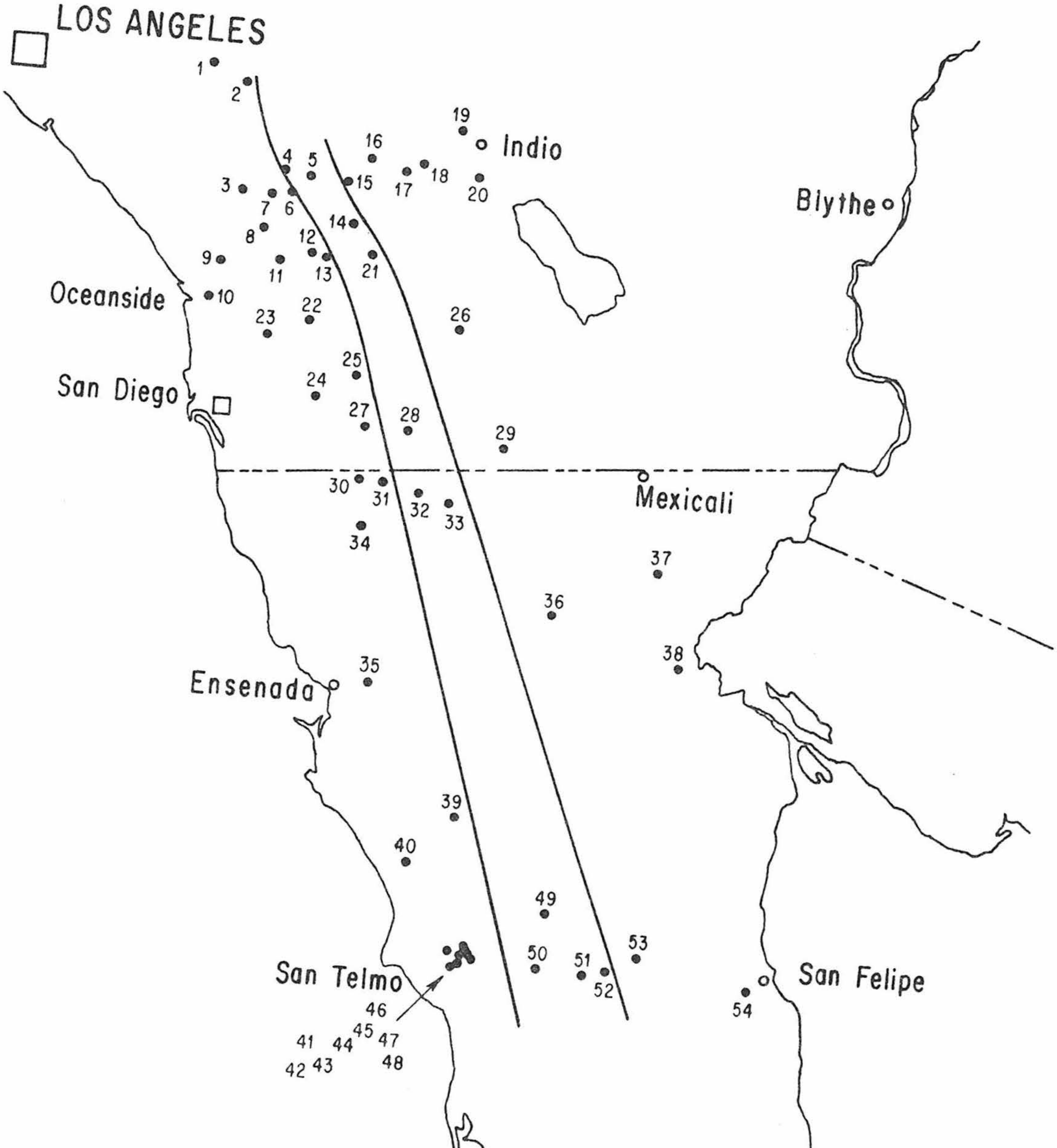

LOS ANGELES
1
2
19
Indio
16
4
5
18
3
15
17
20
Blythe
7
6
8
14
12
9
11
13
21
Oceanside
10
23
22
26
San Diego
25
24
27
28
29
30
31
32
33
Mexicali
34
37
36
38
Ensenada
35
39
40
49
53
50
51
52
San Telmo
San Felipe
54
46
45
41
44
47
42
43
48

TABLE 6

RARE EARTH ELEMENT CONCENTRATIONS OF WHOLE ROCKS IN PPM

Sample #	1	2	3	4	5	6	7	8
Ce	28.5	25.3	53.6	18.27	32.0	46.9	10.3	8.91
Nd	16.8	16.5	21.5	9.91	18.3	22.8	6.92	5.95
Sm	4.13	4.20	3.71	2.11	3.97	4.84	1.87	1.59
Eu	1.12	1.05	.675	.947	1.19	.749	.731	.918
Gd	4.32	4.49	3.28	1.92	3.42	4.35	2.26	1.87
Dy	4.27	4.52	3.08	1.44	2.44	3.91	2.37	1.95
Er	2.56	2.69	1.82	.689	1.11	2.02	1.44	1.17
Yb	2.37	2.52	1.70	.594	.899	1.92	1.32	1.10

Sample #	9	10	11	12	13	14	15	16
Ce	26.4	33.6	23.3	29.16	34.2	32.3	62.8	2.30
Nd	16.2	21.4	15.0	15.85	21.5	15.8	24.5	1.48
Sm	4.14	5.14	4.04	3.80	5.25	3.30	4.02	.408
Eu	1.07	.995	1.06	.973	1.15	1.06	1.07	.295
Gd	4.60	5.50	4.63	4.04	5.30	2.97	3.03	.480
Dy	4.79	5.64	5.05	4.12	5.08	2.38	1.99	.482
Er	2.90	3.24	3.13	2.49	2.83	1.20	.843	.264
Yb	2.74	2.95	3.05	2.33	2.53	1.11	.633	.222

TABLE 6, continued

Sample #	17	18	19	20	21	22	23	24
Ce	61.8	103	64.5	105.2	38.0	28.0	28.0	28.1
Nd	33.6	41.8	25.0	42.2	18.3	19.4	14.4	16.14
Sm	7.04	7.42	4.32	7.05	3.66	4.70	3.44	3.98
Eu	1.51	1.55	1.07	1.64	1.065	1.24	.730	1.012
Gd	5.99	6.13	3.46	5.42	3.00	4.72	3.99	4.45
Dy	4.85	4.68	2.63	3.77	2.14	4.22	4.41	4.57
Er	2.35	2.37	1.32	1.75	.956	2.27	2.96	2.79
Yb	1.93	2.04	1.12	1.44	.835	1.96	3.13	2.70

Sample #	25	26	27	28	29	30	31	32
Ce	31.4	51.0	4.78	37.3	102.6	43.0	22.4	36.8
Nd	9.68	25.2	6.27	16.0	45.4	22.3	11.2	15.8
Sm	2.04	4.69	2.02	2.81	6.16	4.81	2.53	2.71
Eu	.590	1.205	.863	.910	1.65	.974	.923	.873
Gd	1.99	3.55	.225	2.25	4.08	4.99	2.69	2.15
Dy	2.13	2.21	1.80	1.54	2.50	5.10	2.73	1.50
Er	1.26	.917	.854	.718	1.01	2.93	1.67	.709
Yb	1.36	.730	.639	.629	.761	2.47	1.68	.630

TABLE 6, continued

Sample #	33	34	35	36	37	38	39	40
Ce	5.15	48.8	32.0	67.7	67.8	75.2	64.1	20.0
Nd	5.62	21.7	18.4	27.2	38.7	26.9	40.1	12.4
Sm	2.02	4.69	4.41	5.15	7.67	4.25	9.83	3.29
Eu	.897	1.02	1.07	1.25	1.75	1.10	1.63	.829
Gd	2.64	4.99	4.77	4.37	6.40	2.83	10.6	3.79
Dy	2.60	5.35	4.53	3.35	5.07	1.57	12.8	4.20
Er	1.33	3.41	2.67	1.59	2.51	.529	8.35	2.66
Yb	1.02	3.43	2.49	1.34	2.14	.368	8.61	2.69

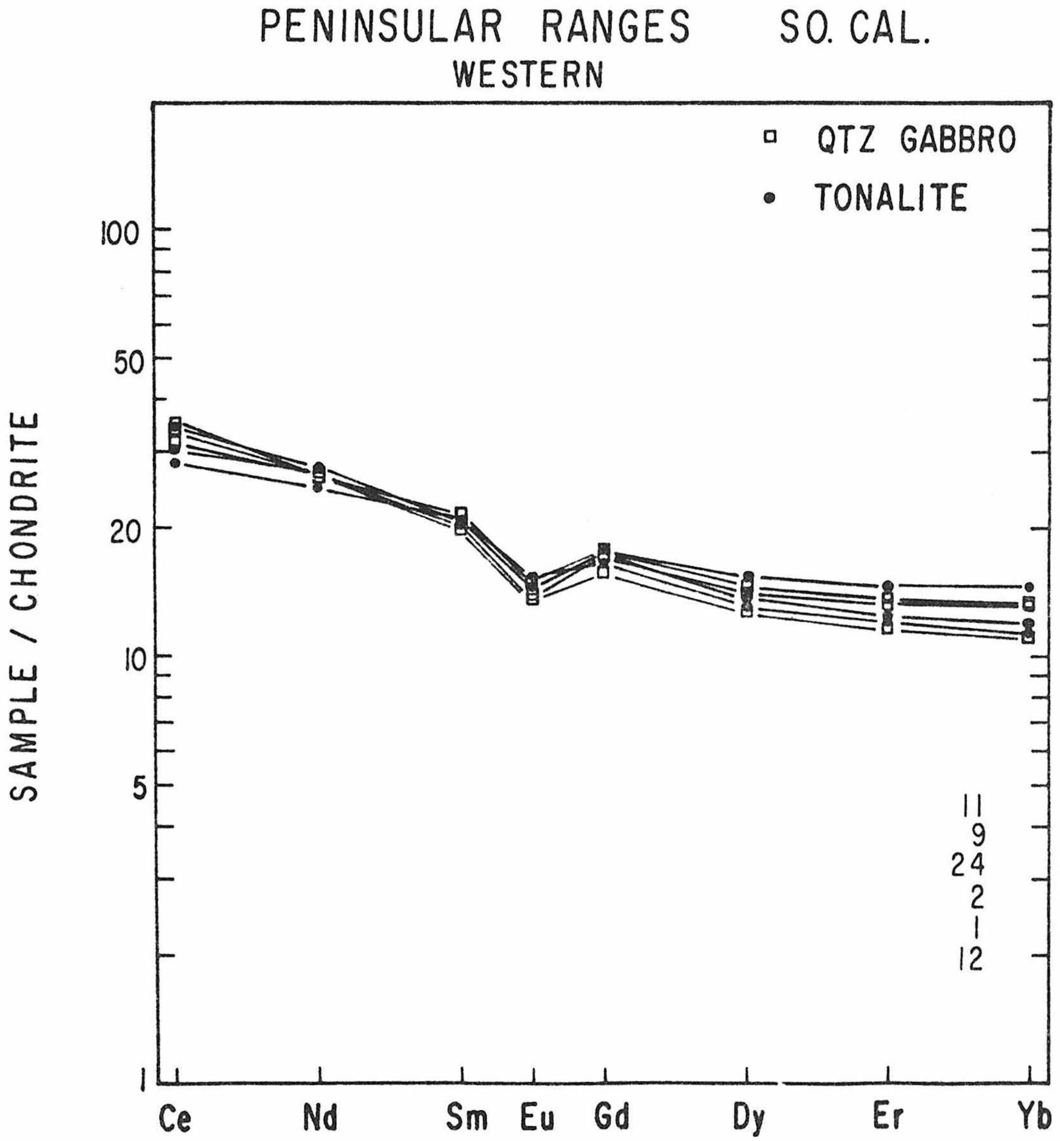
Sample #	41	42	43	44	45	46	47	48
Ce	13.9	17.9	29.7	11.9	10.2	17.0	8.60	22.2
Nd	12.0	14.3	28.5	9.12	7.83	12.85	7.47	19.1
Sm	3.81	4.34	7.23	2.71	2.34	3.72	2.57	5.25
Eu	.737	1.18	1.12	.955	.841	1.53	7.20	1.65
Gd	5.09	5.55	8.24	3.45	2.88	4.61	3.44	6.22
Dy	6.56	6.55	8.79	4.07	3.26	5.02	4.31	6.56
Er	4.57	4.30	5.60	2.78	2.10	3.15	2.89	4.24
Yb	4.50	4.25	5.34	2.70	2.05	3.01	2.89	4.17

TABLE 6, continued

Sample #	49	50	51	52	53	54
Ce	16.0	19.5	37.6	32.3	57.3	40.9
Nd	10.1	11.3	19.8	14.6	24.3	15.3
Sm	2.31	2.32	3.86	2.59	3.95	2.57
Eu	.954	.792	1.12	.874	1.09	.789
Gd	2.24	1.92	3.17	1.97	2.80	2.07
Dy	1.87	1.43	2.29	1.27	1.66	1.48
Er	1.00	.737	1.11	.586	.674	.714
Yb	.911	.666	.959	.470	.532	.602

	Argillitic metasediment, French Valley Formation	Volcanic sandstone, Baja California
Ce	71.8	48.0
Nd	38.3	35.1
Sm	7.24	9.93
Eu	1.20	2.51
Gd	5.72	11.8
Dy	3.14	12.1
Er	1.38	6.44
Yb	1.38	4.93

FIGURE 26. REE patterns of characteristic tonalites and quartz gabbros from the western region, southern California. Sample numbers are listed in the order of occurrence of the patterns immediately above them, as will be true for all of the following diagrams.


PENINSULAR RANGES SO. CAL.
WESTERN
□ QTZ GABBRO
● TONALITE
SAMPLE / CHONDRITE
100
50
20
10
5
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
11
9
24
2
1
12

of the patterns are concave-down for the light REE and concave-up for the heavy REE.

A tonalite and a low-K_2O granodiorite from near the eastern margin of the western area (Fig. 27) differ slightly in their REE characteristics. They have comparable abundances of heavy REE, but their patterns have slightly steeper slopes which transgress the parallel slopes of the other tonalites. Both of these tonalites have somewhat higher light REE contents, and one also has a larger Eu anomaly.

Five samples from leucogranodioritic plutons have been analyzed (Fig. 28). These samples include two leucogranodiorites (samples 3,23) from plutons mapped as Woodson Mountain granodiorite and one from the Escondido Creek leucogranodiorite (sample 10)(Larsen, 1948). The two remaining are from bodies outside of Larsen's study area but strongly resembling Woodson Mountain granodiorite in lithology. The most striking difference in REE patterns between these rocks and the quartz gabbros and tonalites is the large negative Eu anomalies which all but one of these leucogranodiorites possess. The slopes for the light REE range to steeper values and are nearer linearity than the tonalites. Although there is a much larger range in abundances of heavy REE, the slopes are close to those of the more mafic rocks. Note that two of the samples (23,25), however, flatten and reverse their slopes becoming positive in the heaviest REE. In particular, sample 25 has a very pronounced concave-up curva-

FIGURE 27. REE patterns of a tonalite and a low-K_2O granodiorite from near the eastern margin of the western region, southern California. Field for characteristic western tonalites is indicated.

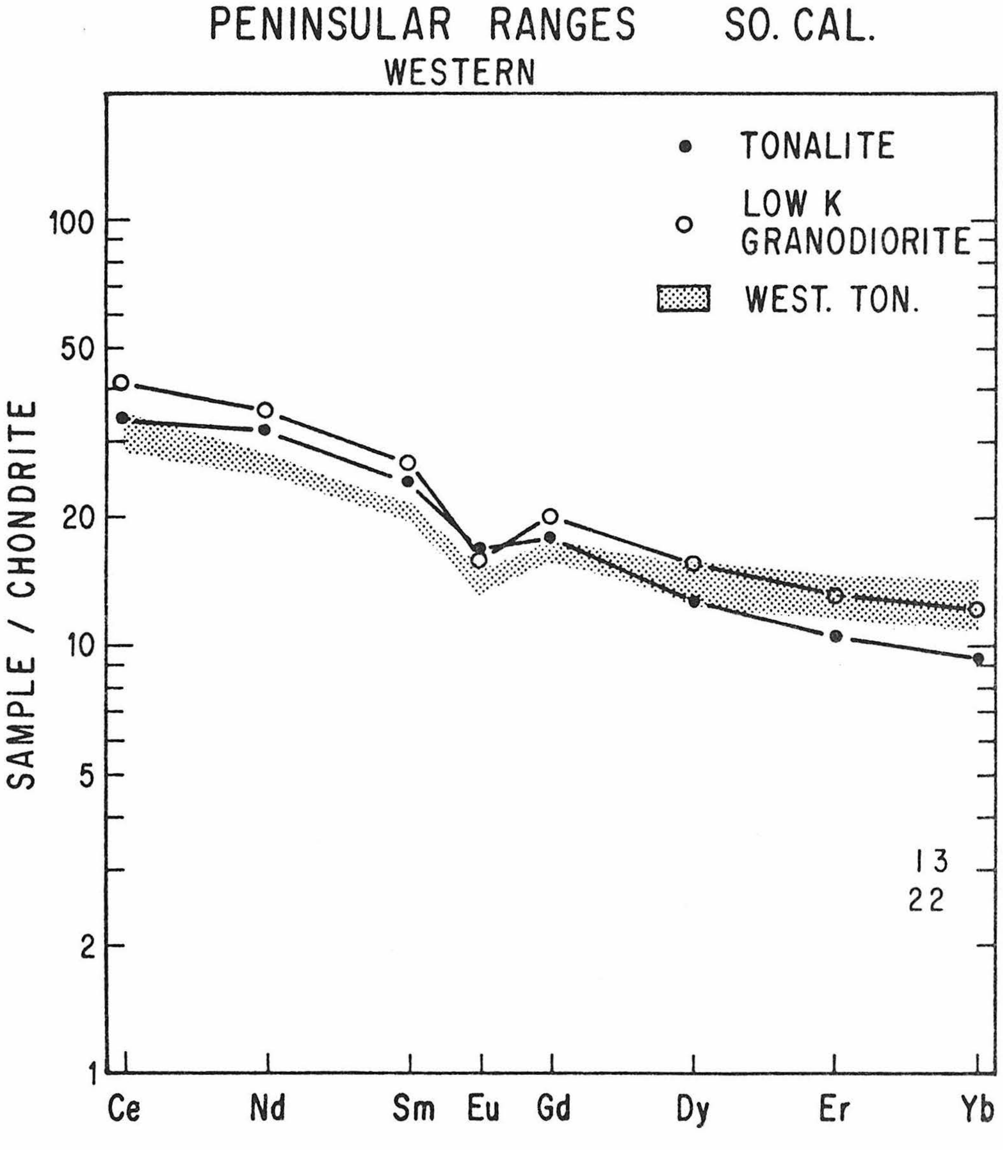
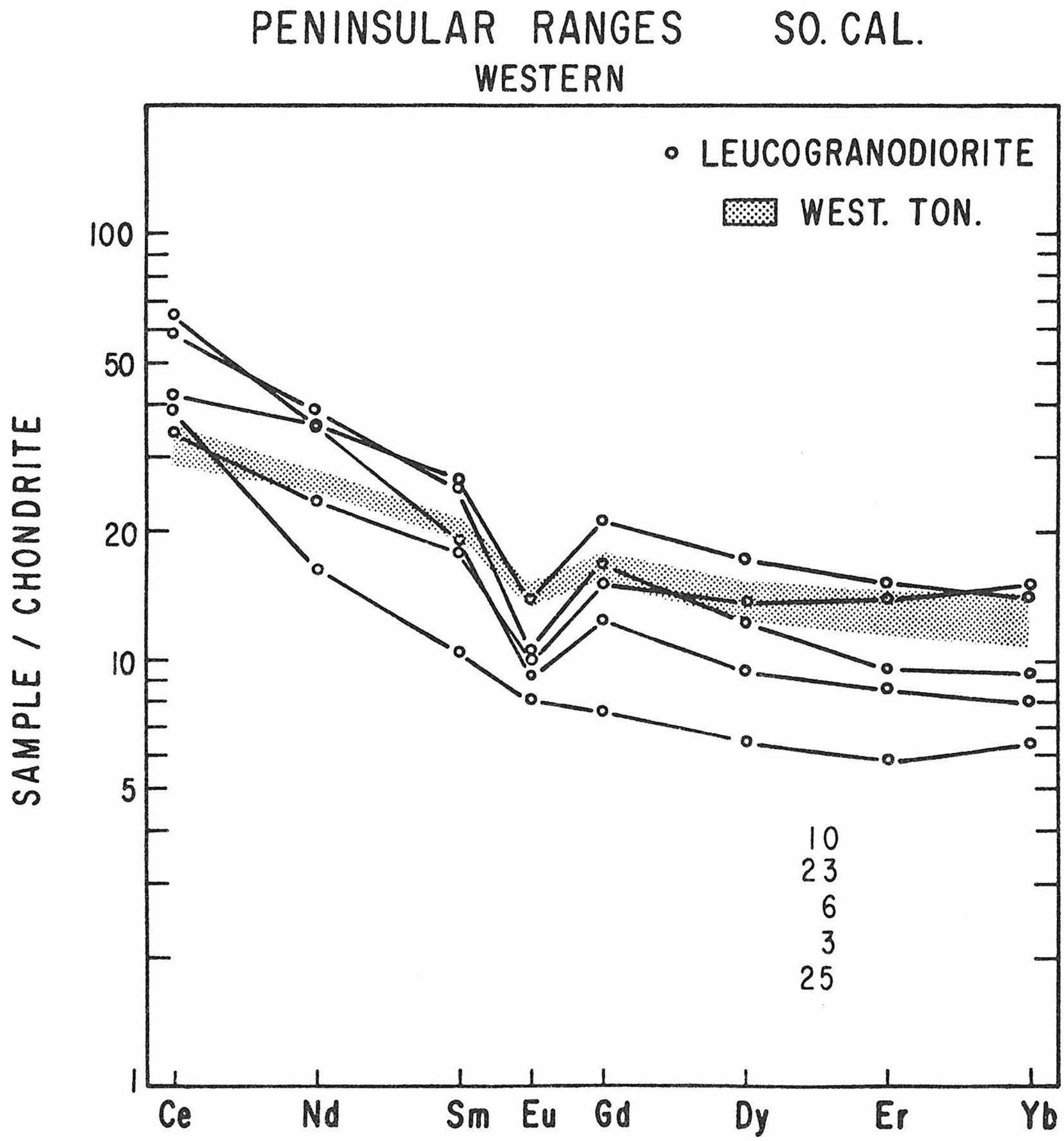

PENINSULAR RANGES SO. CAL.
WESTERN
● TONALITE
○ LOW K GRANODIORITE
▨ WEST. TON.
SAMPLE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
13
22

FIGURE 28. REE patterns of 5 leucogranodiorites from the western region, southern California. Field for characteristic western tonalites is indicated.

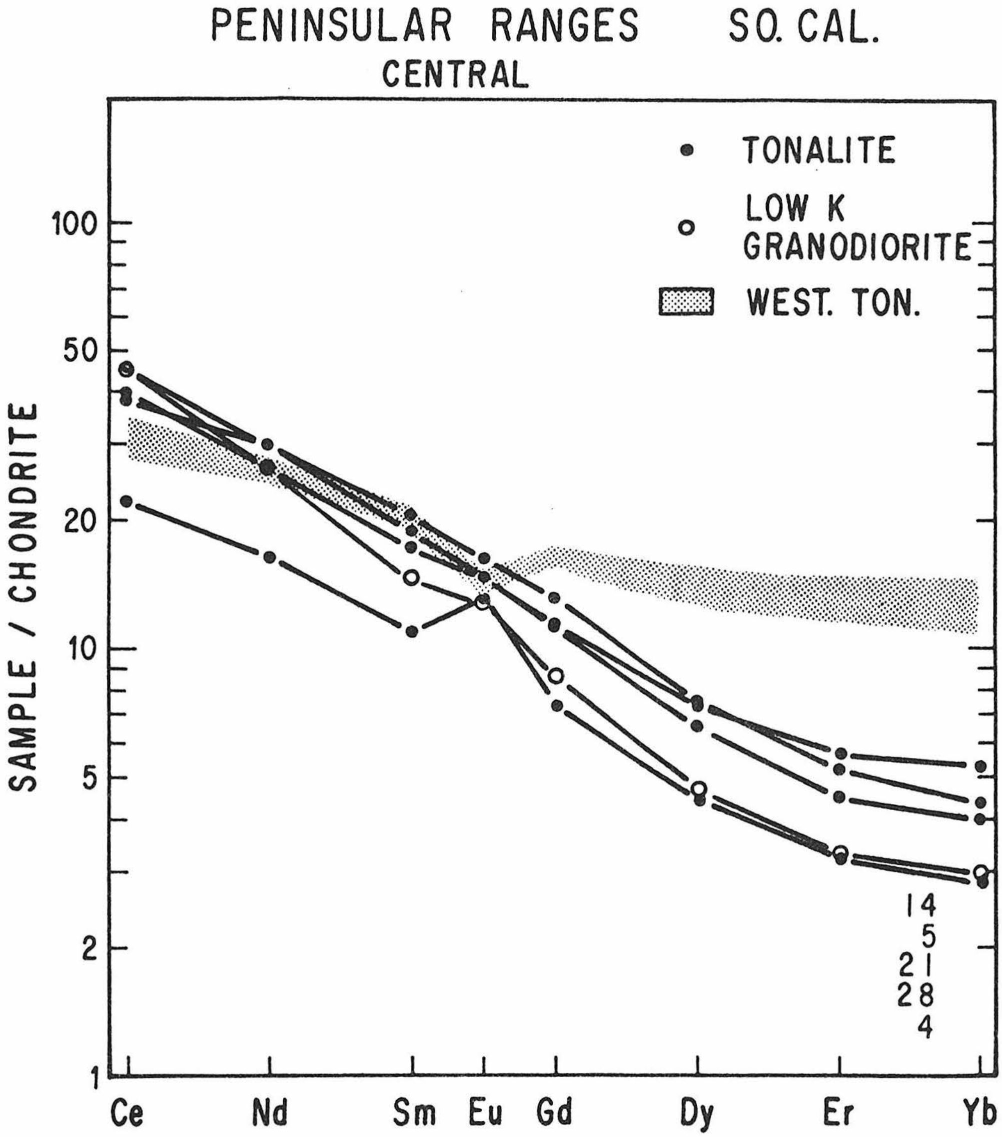

PENINSULAR RANGES SO. CAL.
WESTERN
○ LEUCOGRANODIORITE
▨ WEST. TON.
SAMPLE / CHONDRITE
100
50
20
10
5
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
10
23
6
3
25

ture over its entire pattern, and is further distinguished from the other leucogranodiorites by its overall lower abundances of REE and very small Eu anomaly.

Central Region

The rocks of the central region are made up almost entirely of tonalite. REE analyses of four tonalites and one low-K_2O granodiorite representative of this region are presented in Figure 29. All of these samples show a notable depletion of heavy REE (up to half an order of magnitude) compared to the western tonalites. This is accompanied by a steeper slope in the middle and heavy REE. The light REE contents are roughly similar to the western tonalites, and slopes are but slightly steeper. These features are the criteria used to distinguish the western and central regions. The negative Eu anomaly that is characteristic of the western tonalites is not present in these samples. Three of the tonalites and the low-K_2O granodiorite have no or a very small positive anomaly, and one tonalite (sample 4) has a distinct positive one. This tonalite has Eu contents similar to the other rocks, but lower abundances of trivalent REE. As in the western region, the patterns of these rocks are all essentially parallel (excepting Eu, of course). The curvatures are slightly concave-down in the light REE and concave-up in the heavy REE.

FIGURE 29. REE patterns of 4 tonalites and a low-K_2O granodiorite from the central region, southern California. Field for characteristic western tonalites is indicated.


PENINSULAR RANGES SO. CAL.
CENTRAL
● TONALITE
○ LOW K GRANODIORITE
▨ WEST. TON.
SAMPLE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
14
5
21
28
4

Eastern Region

The REE patterns for all of the observed granitic rocks are highly fractionated. Three tonalite samples (Fig. 30) display a strong parallelism in their patterns. Slopes are steep and near linear in the light and middle REE, flattening somewhat in the heavy REE. The highly fractionated patterns and high light REE contents (Ce > 50X chond.) are the distinguishing characteristics of eastern tonalites. Eu anomalies are negligible.

Four granodiorites (Fig. 31) have patterns quite similar in form to the tonalites. The REE contents of these rocks range to higher values, especially in the heavy REE, and small to moderate negative Eu anomalies are present. One low-K_2O granodiorite (sample 17) differs somewhat from the other samples in its distinct concave-down curvature in the light REE.

Baja California

Western Region

A concentrated effort has been made investigating the San Telmo ring complex, but other rocks studied south of the border are discussed first. Two of the three tonalites analyzed (samples 35,40; Fig. 32) have the same REE characteristics as the tonalites of southern California. The heavy REE are near flat, the light REE have a moderate relative enrichment, and both tonalites have a modest negative Eu anomaly. Another tonalite (sample 31) exhibits several dif-

FIGURE 30. REE patterns of 3 tonalites from the eastern region, southern California. Fields for characteristic western tonalites and central tonalites are indicated.

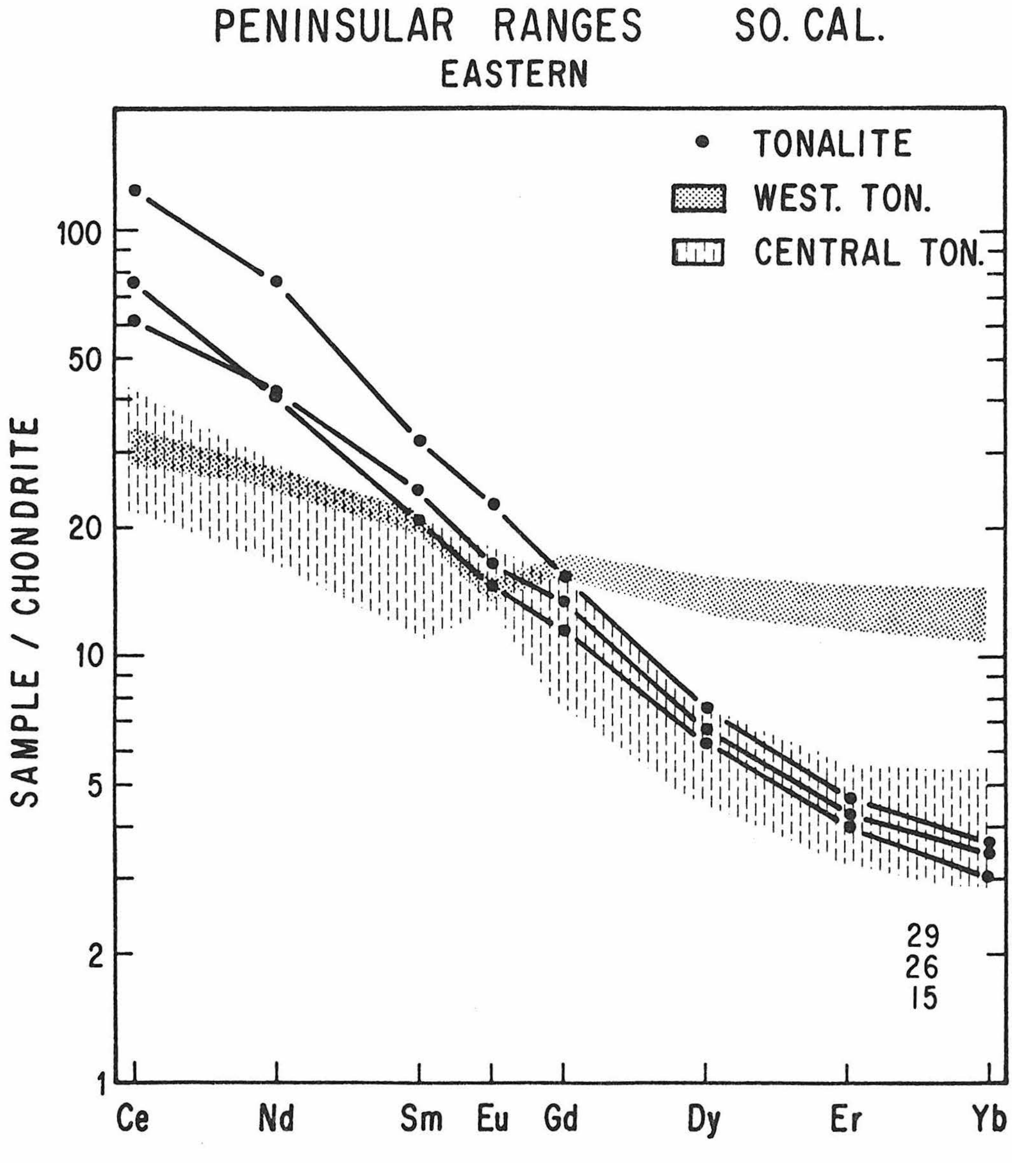

PENINSULAR RANGES SO. CAL.
EASTERN
● TONALITE
▨ WEST. TON.
▥ CENTRAL TON.
SAMPLE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
29
26
15

FIGURE 31. REE patterns of granodiorites from the eastern region, southern California. Fields for characteristic western tonalites and eastern tonalites are indicated.

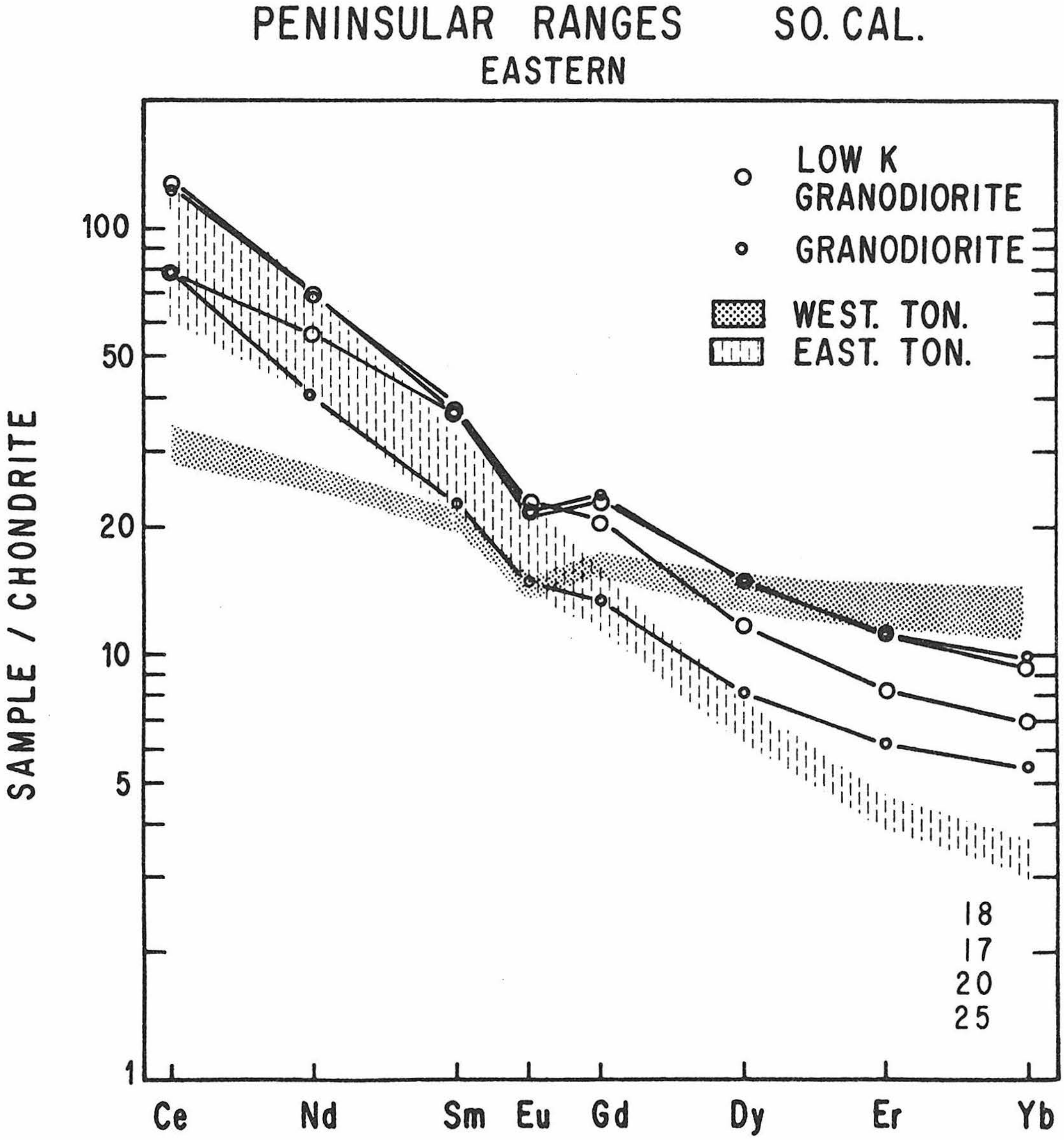
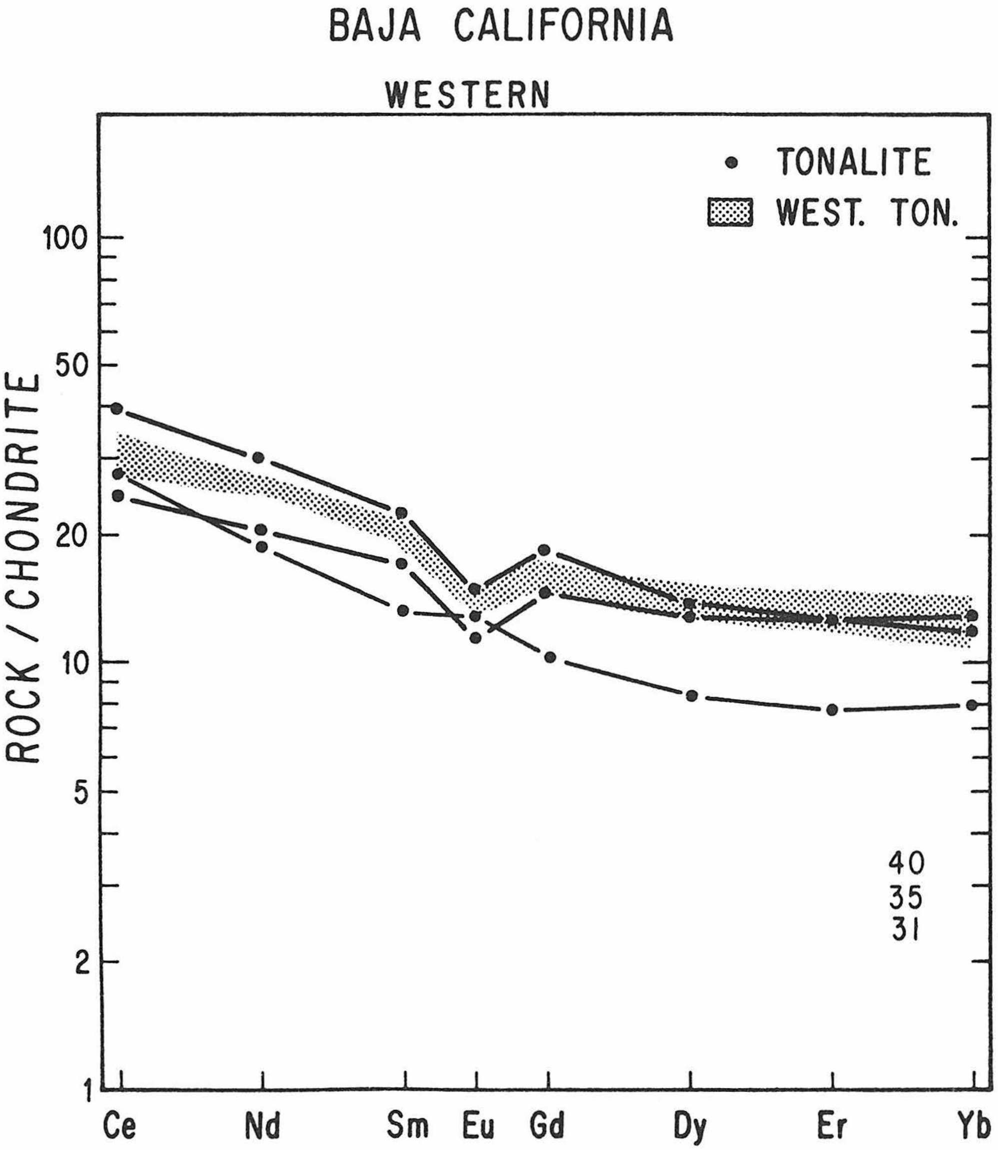

PENINSULAR RANGES SO. CAL.
EASTERN
LOW K GRANODIORITE
GRANODIORITE
WEST. TON.
EAST. TON.
SAMPLE / CHONDRITE
100
50
20
10
5
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
18
17
20
25

FIGURE 32. REE patterns of 3 tonalites from the western region, Baja California. Field for characteristic western tonalites in southern California is indicated.


BAJA CALIFORNIA
WESTERN
● TONALITE
▨ WEST. TON.
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
40
35
31

ferences. This sample has lower REE concentrations and a small positive Eu anomaly. Its pattern has a reversal in slope in the heaviest REE, and a slightly steeper and more linear slope in the light REE. Note, however, the similarity in slope of the intermediate REE between this rock and all the other tonalites from the western Peninsular Ranges.

One low-K_2O granodiorite (sample 34) and two leucogranodiorites (samples 30,39) plotted in Figure 33 have REE patterns roughly parallel to the tonalites except for their more pronounced negative Eu anomalies. Two of these rocks have abundances at the high end of the range observed for leucogranodiorites in southern California. The remaining sample (39) has concentrations of middle and heavy REE far exceeding all other rocks analyzed in this study, and has a flatter fractionation pattern than any other granodiorite except for the San Telmo suite.

The dominantly tonalitic rocks of the San Telmo ring complex, 120 km SSE of Ensenada, have REE patterns distinguishing them from all other observed patterns in the batholith. Four samples of leucocratic tonalite (samples 43-46) and a closely related low-K_2O granodiorite (sample 48) from the central and eastern parts of the complex have exceptionally flat, unfractionated patterns (Fig. 34b). All these samples show a depletion of Ce relative to Nd, a feature that has not been observed in any other quartz-bearing rocks in the peninsula. The two most leucocratic and alkali feldspar-

FIGURE 33. REE patterns of 1 low-K_2O granodiorite and 2 leucogranodiorites from the western region, Baja California. Field for characteristic western tonalites in southern California is indicated.

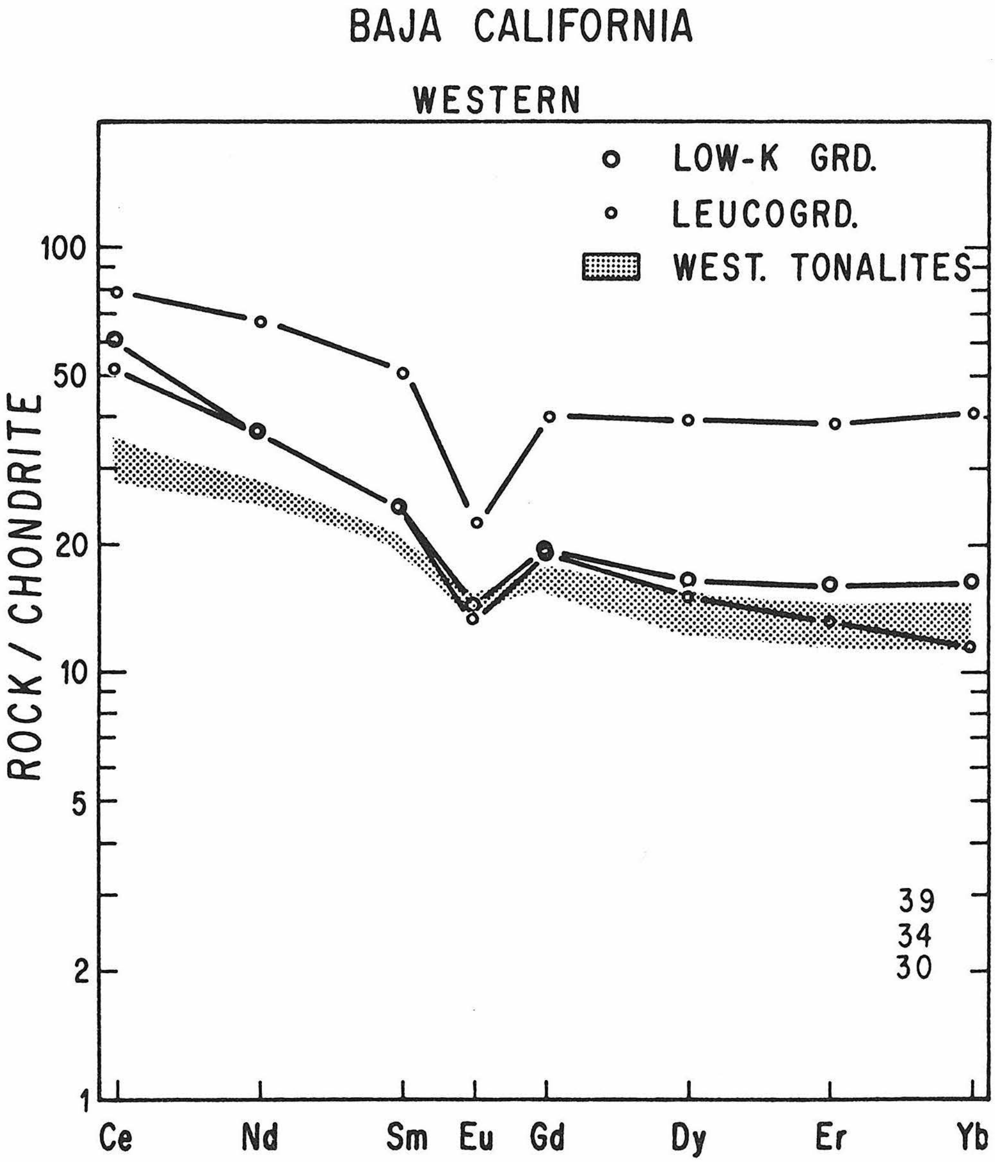
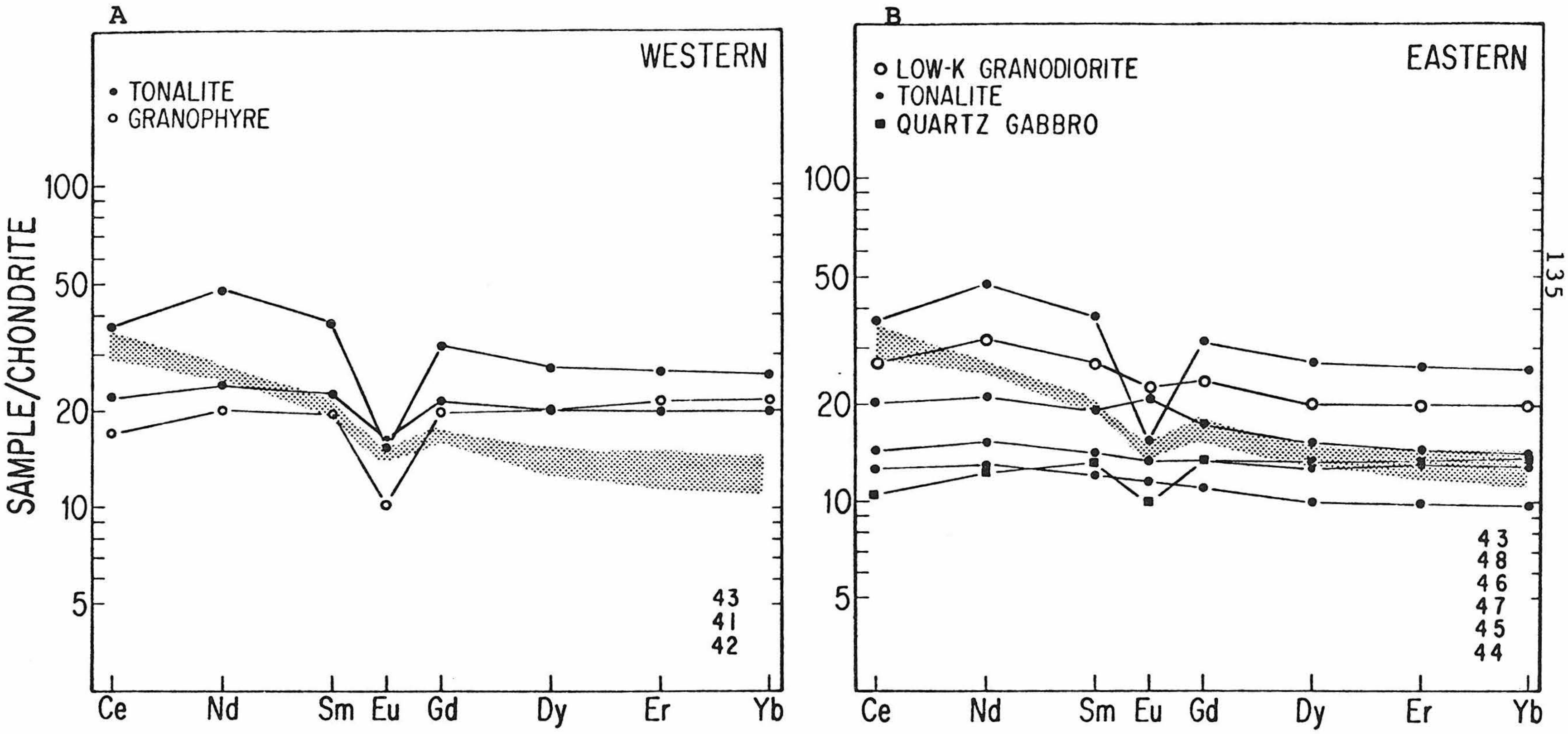

BAJA CALIFORNIA
WESTERN
LOW-K GRD.
LEUCOGRD.
WEST. TONALITES
ROCK / CHONDRITE
100
50
20
10
5
2
1
39
34
30
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

FIGURE 34. REE patterns of rocks of the San Telmo pluton, Baja California. Field for characteristic western tonalites in southern California is indicated.

a) 2 leucotonalites and 1 leucogranodioritic granophyre from the western and central parts of the pluton.

b) 4 leucotonalites (sample 43 repeated), 1 low-K_2O granodiorite, and 1 quartz gabbro from the eastern and central parts of the pluton.

SAN TELMO RING COMPLEX



poor tonalites (samples 44,45) have the lowest REE abundances and no Eu anomalies. The most mafic rocks (samples 46,48) have intermediate REE abundances with a small negative and positive Eu anomaly, respectively, while a leucocratic but alkali feldspar-rich leucocratic tonalite (sample 43) has the highest REE contents and a very pronounced negative Eu anomaly. A quartz gabbro (sample 47) displays a more pronounced depletion in the light REE.

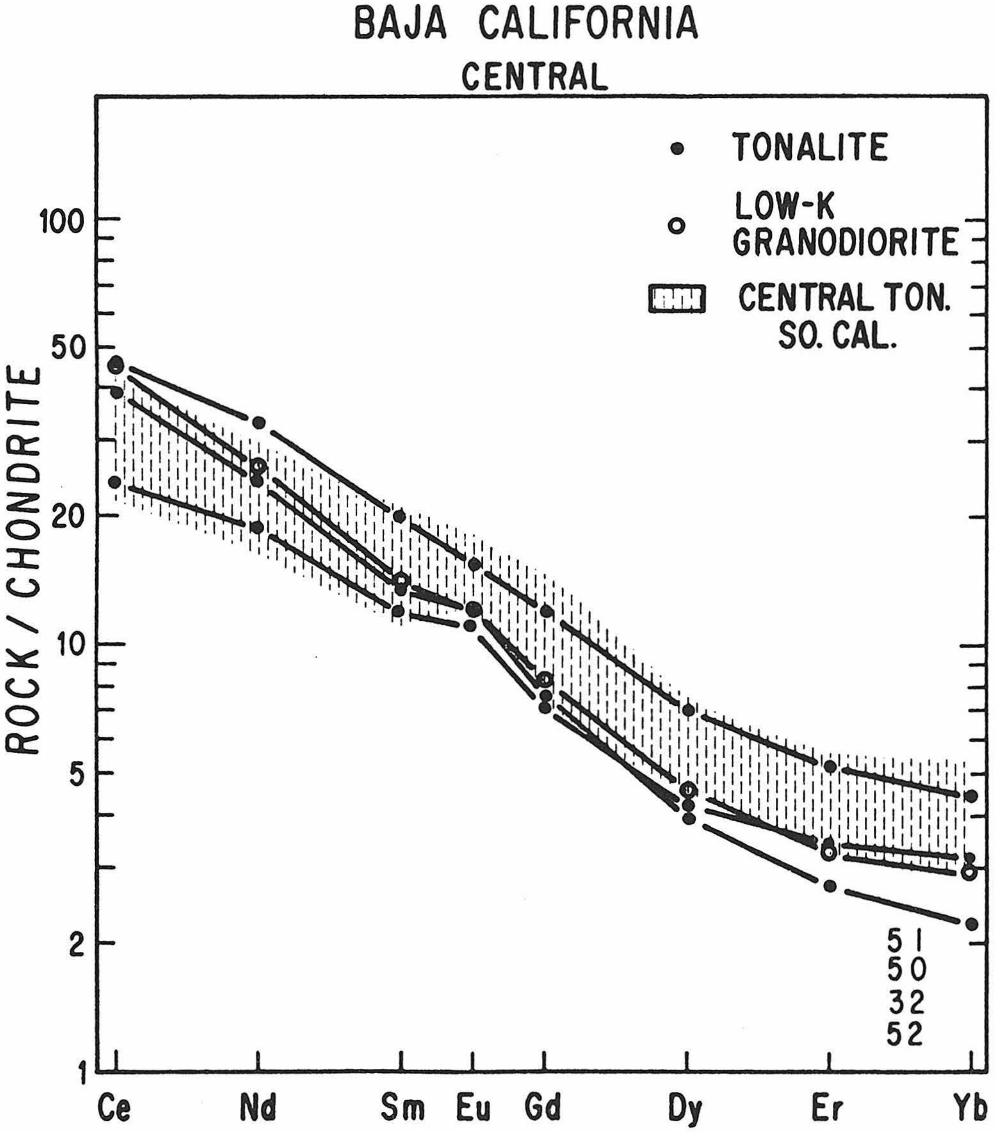
On the western side of the complex, a tonalite (sample 42) and a granophyre (sample 41) (Fig. 34a) also have flat, unfractionated patterns with a small relative depletion in the light REE. The granophyre has a large negative Eu anomaly and an overall positive slope, dipping toward the light REE.

Central Region

Three samples of tonalite from the foothills and western flank of the Sierra San Pedro Martir, and an additional sample from the southern portion of the La Posta pluton have been analyzed (Fig. 35). One of the tonalites (sample 50) is from the seriate porphyritic unit of the distinctive San Jose pluton (Murray, 1978).

All of these rocks display the same heavy REE depleted pattern found in the central region in southern California. Two of the tonalites (samples 50,52) and the low-K_2O granodiorite (sample 32) have slight positive Eu anomalies. The pattern of the low-K_2O granodiorite, from the La Posta pluton,

FIGURE 35. REE patterns of 3 tonalites and 1 low-K_2O granodiorite from the central region, Baja California. Field for similar rocks of the central region in southern California is indicated for comparison.


BAJA CALIFORNIA
CENTRAL
● TONALITE
○ LOW-K GRANODIORITE
▥ CENTRAL TON. SO. CAL.
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
51
50
32
52

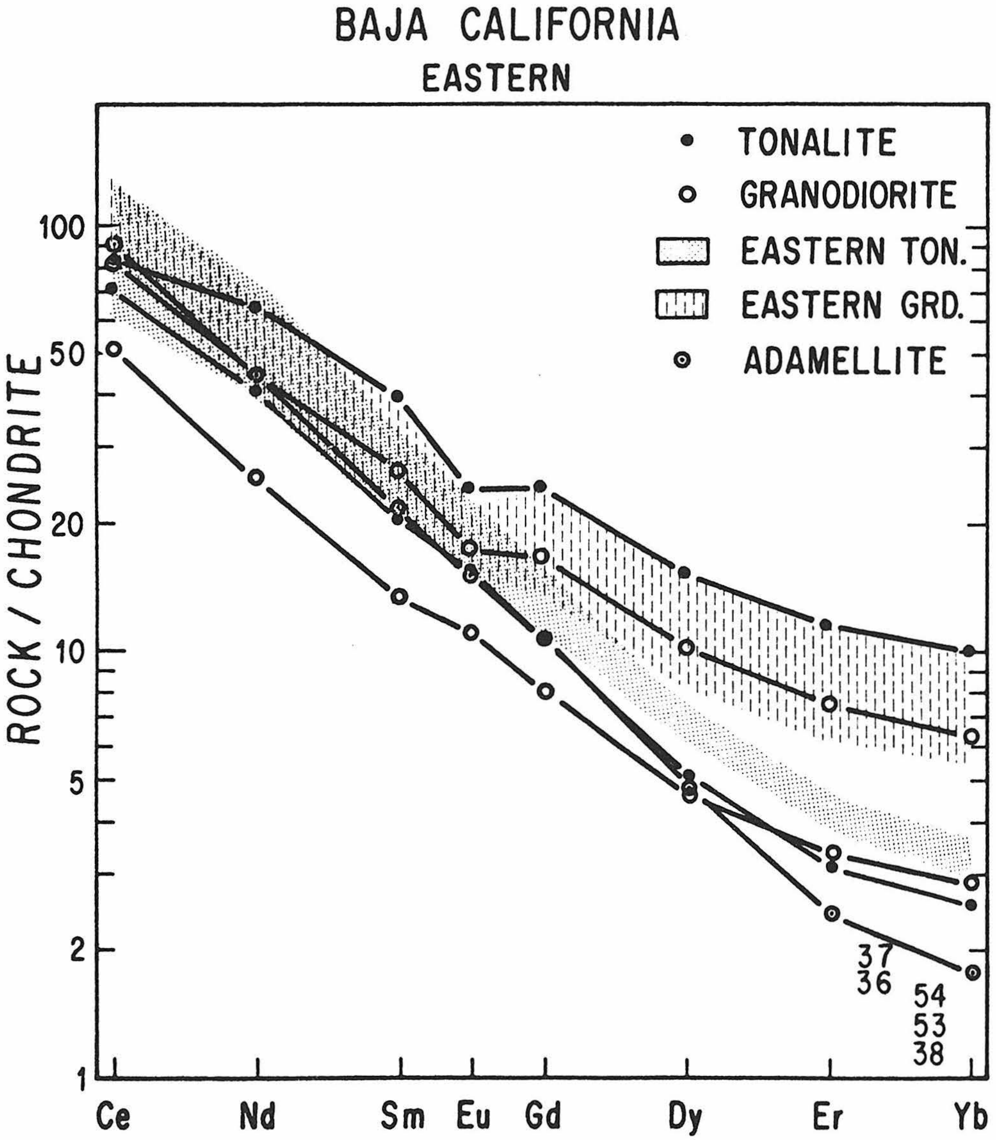
is essentially identical to the pattern obtained from the sample of this pluton in southern California, 40 km to the north. The pattern of the remaining tonalite (sample 51) is indistinguishable from similar tonalites in southern California (Fig. 29).

Eastern Region

A tonalite from the base of the eastern scarp of the Sierra San Pedro Martir (sample 53), a tonalite gneiss from the Sierra de la Cucopah (sample 37), granodiorites from the Sierra Juarez (sample 36) and the desert ranges east of San Felipe (sample 54), and an adamellite from the Sierra del Mayor (sample 38) all have fractionated REE patterns (Fig. 36). In all these samples except the tonalite gneiss, the near linear patterns in the light and middle REE grades into a concave-up curvature in the heavy REE, and Eu anomalies are very small or absent. A strong similarity exists between the shape of these patterns and those obtained for eastern tonalites and some granodiorites in southern California, although the light REE contents of these rocks do not range to as high values.

The tonalite gneiss is distinct from the other Baja California samples, but it is quite similar to a low-K_2O granodiorite in southern California (sample 17). Its REE pattern is concave-down in the light REE, has a moderate negative Eu anomaly, and has much higher concentrations of heavy REE. Slopes in the middle to heavy REE are still subparallel

FIGURE 36. REE patterns of 2 tonalites, 2 granodiorites, and an adamellite from the eastern region, Baja California. Fields for eastern tonalites and eastern granodiorites in southern California are indicated.


BAJA CALIFORNIA
EASTERN
● TONALITE
○ GRANODIORITE
EASTERN TON.
EASTERN GRD.
ADAMELLITE
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
37
36
54
53
38

to other samples in this region.

Gabbros

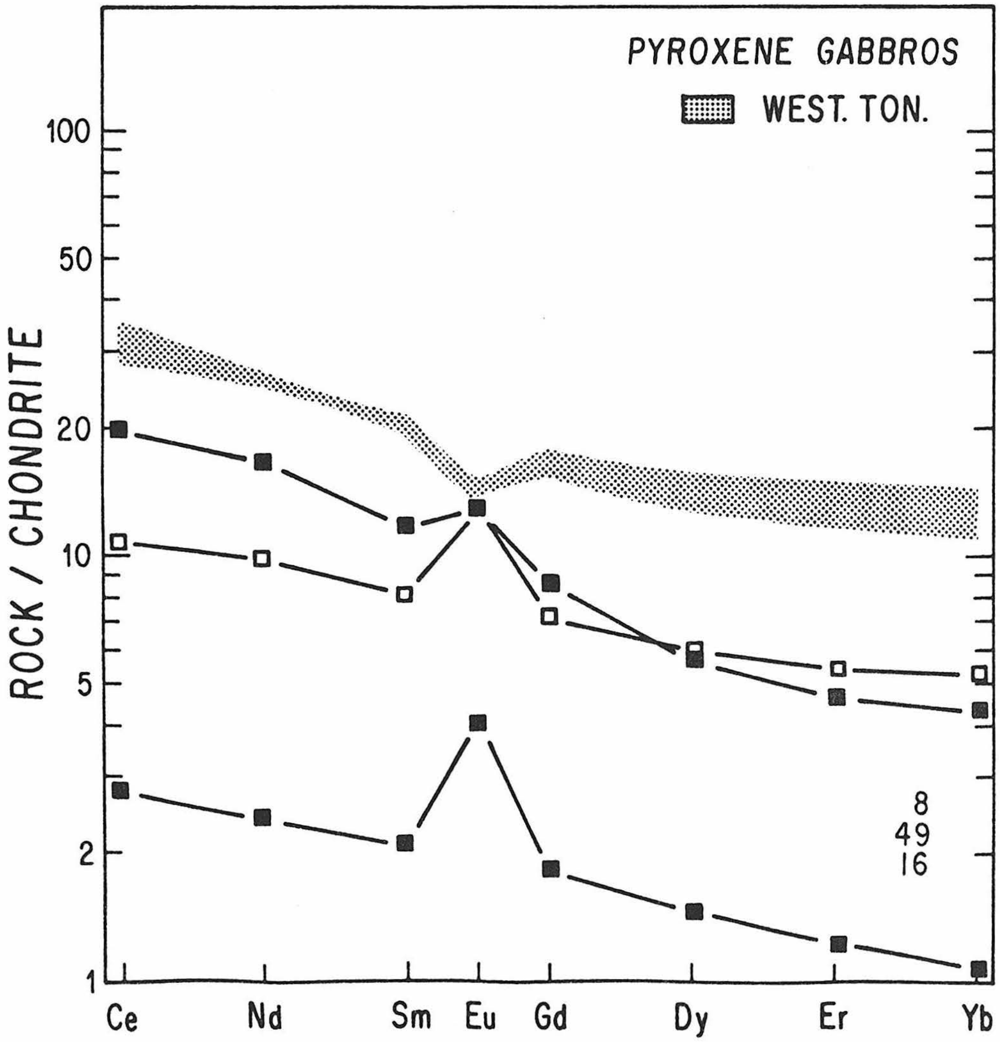
The original use of the name gabbro in the Peninsular Ranges batholith included a wide variety of rock types from peridotite to quartz-biotite gabbros (Miller, 1937; Larsen, 1948). In this work, however, a distinction is drawn among the gabbros, dividing them into two groups: those containing at least several percent free quartz and biotite, and a sodic labradorite plagioclase, termed quartz gabbros; and those with only traces or no free quartz and biotite, and a more calcic plagioclase, hereafter referred to as gabbros with various other mineralogic or chemical qualifying terms.

A set of significant petrologic and geochemical distinctions can be drawn among these two gabbro groups. In the field as well as petrographically, quartz gabbros are transitional to tonalite. They share the same mineral assemblages, differing only in proportions and plagioclase composition. The geochemical characteristics of the quartz gabbros are closely related to the tonalites, which is particularly well displayed by the indistinguishability of quartz gabbro and tonalite REE patterns (Fig. 26). In contrast, gabbros do not appear to grade into more silica-rich lithologies. Some gabbroic bodies in central San Diego County have been shown to be differentiated complexes (Walawender, 1976; Nishimori, 1976); others are not. Their compositions tend to scatter on otherwise smooth chemical variation diagrams and trends (Larsen, 1948; Baird, Baird, and Welday, 1974), and they have somewhat displaced regional variations in initial $^{87}Sr/^{86}Sr$ ratios (Early and Silver, 1973). As is seen below, these gabbros usually display exceptional features in their REE patterns.

The six gabbro samples which have been analyzed are presented grouped according to their dominant mafic mineral. Two of three pyroxene gabbros (Fig. 37) are in southern California: a noritic hornblende gabbro (sample 8) from the Johnson quarry in Pala, which is located in the western region; an olivine-rich, anorthitic gabbro (sample 16; in Appendix I) located within the eastern region. The REE pattern of the gabbro from Pala is parallel to the western quartz gabbros and tonalites, but with much lower concentrations of REE and a large positive Eu anomaly. The REE abundances of sample 16 are exceptionally low, the lowest of any rock measured in this study. In contrast to the gabbro from Pala, the slope of the pattern of sample 16 is highly divergent from those of other rocks of the eastern region.

A pyroxene gabbro (sample 49) containing some free quartz, from the central region in Baja California, has abundances and fractionations of REE comparable to nearby tonalites, with the qualification that its positive Eu anomaly is slightly larger. In some contrast to the two other pyroxene gabbros, this gabbro follows the trends and abundances displayed by rocks of the same geographic region.

FIGURE 37. REE patterns of 3 pyroxene gabbros. Field for characteristic western tonalites in southern California is indicated.


PYROXENE GABBROS
▨ WEST. TON.
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
8
49
16

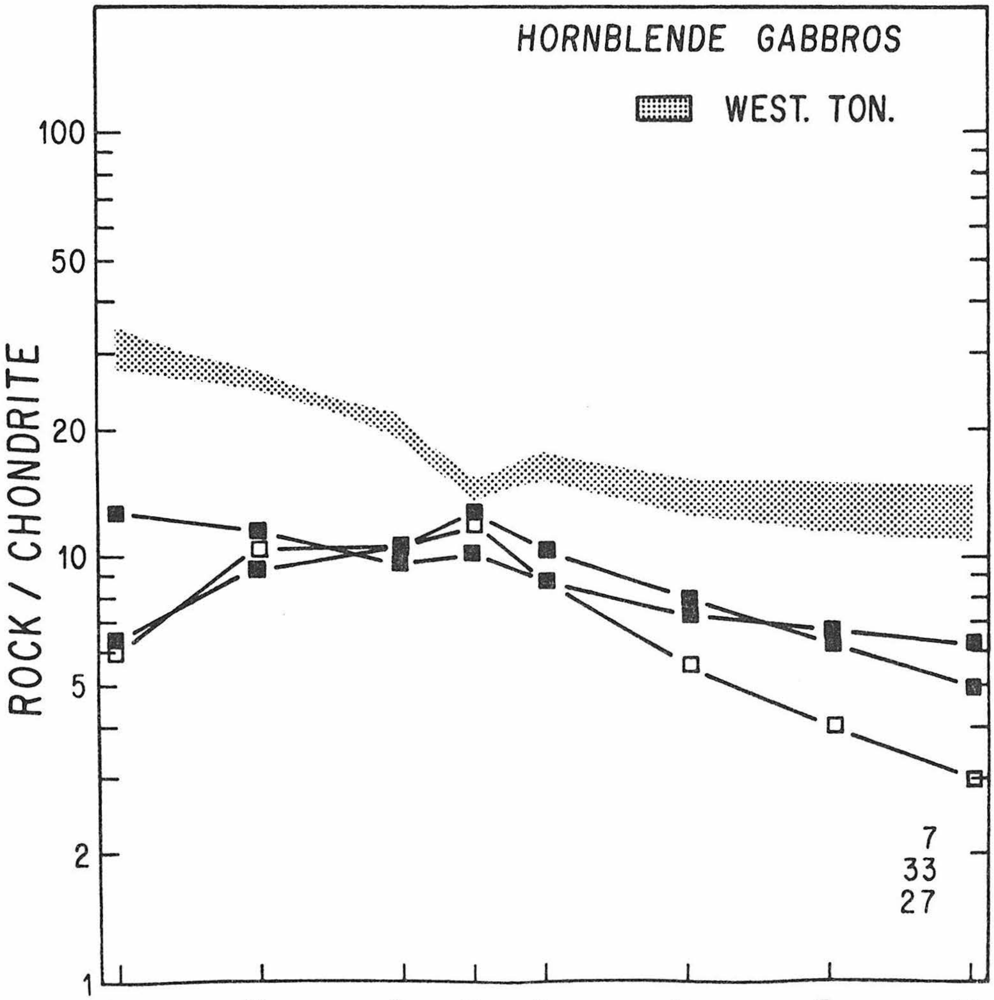
The three hornblende gabbros (Fig. 38) have additional variations. A gabbro (sample 7) 15 km NE of the Pala gabbro has similar REE abundances and a parallel slope to this sample, but a distinctly smaller positive Eu anomaly. The two other hornblende gabbros have strikingly different characteristics. A gabbro from Los Pinos Mountain (sample 27) (Walawender, 1976) located in the western region, and a gabbro (sample 33) from the central region just south of the border both possess an unusual light REE depletion with a very pronounced dip from Nd to Ce. This is accompanied by an unfamiliar linear sloping characteristic in the heavy REE. Both gabbros also have modest positive Eu anomalies.

The diversity in the field, petrographic, and chemical features of the gabbroic rocks of the Peninsular Ranges are matched by the range of the REE variations observed here. A more comprehensive study of the gabbros is desirable.

REGIONAL PATTERNS OF REE VARIATIONS IN ROCKS

The foremost feature of these data is that the REE fractionations of granitic rocks vary systematically and drastically across the Peninsular Ranges batholith, from near flat in the west to highly fractionated in the east. This variation is transverse to the long axis and structural trends, and parallel to known asymmetries in petrologic, geochronologic, and Sr isotopic properties. While there are differences in the REE characteristics associated with variations in bulk composition within each region, these differences are

FIGURE 38. REE patterns of 3 hornblende gabbros. Field for characteristic western tonalites in southern California is indicated.


HORNBLENDE GABBROS
WEST. TON.
ROCK / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
7
33
27

for the most part secondary compared to the magnitude of the geographically correlated trends. In contrast, there is a strong tendency towards persistent REE characteristics parallel to the axis of the batholith in each of the longitudinal regions.

The abundant tonalites, which have comparable bulk and mineral compositions whether from the western, central, or eastern regions, illustrate the regional trends. The uniformity in the abundances and fractionations of the REE, including the magnitude of the Eu anomaly, in almost all the tonalites and quartz gabbros of the western region is striking. Tonalites of the central region also display much internal similarity in their REE characteristics, but contrast with western rocks of similar composition. They are strongly fractionated in the middle to heavy REE, with contents of the heaviest REE a factor of three to five less than the western tonalites. Also, unlike their western analogues, these tonalites do not have negative Eu anomalies. The tonalites of the eastern region, with one exception, display a consistent steep-sloped fractionation pattern. The concentrations of heavy REE range to lower values and have a somewhat steeper slope than found in most of the central tonalites. The largest difference is in the abundance and fractionation in the light REE. Whereas the Ce contents of the central tonalites do not exceed 50 times chondrites, they range from 75 to over 120 times chondritic levels in these rocks. The overall fractionation patterns become nearly linear.

The tonalites thus display two separate and distinct major changes in their REE patterns from west to east. The first transition is a fractionation and depletion of the progressive middle to heavy REE and disappearance of negative Eu anomalies. The second transition is light REE enrichment with a steep, near linear slope. These variations were summarized in Figure 30.

In southern California, a special effort was made to obtain samples close to the geographical boundary between the western and central regions. There were two purposes: to more precisely locate this boundary, and to determine whether the transition was sharp and discontinuous, or a smooth variation. This effort yielded a tonalite and a low-K_2O granodiorite which appear to have some slightly transitional properties in their REE patterns (Fig. 27). These rocks are distinguished from the rest of the western tonalites by their very slightly steeper slopes in the middle to heavy REE and somewhat higher light REE contents. They resemble most closely the tonalites of the western region, their differences being small compared to the other regional changes. The distribution of data points restrict these rocks to only a narrow strip compared to the rest of the western and central tonalites, which underly broad areas over which they have uniform characteristics.

The boundary between the western and central areas is placed at or just to the east of these rocks, and the transition is inferred to be abrupt. This boundary is drawn as a smooth curve in Figure 23, although in a few places it must be bent. The actual boundary may be somewhat more irregular, perhaps following contacts between intrusive units. This would be true if large intrusive units were internally uniform with respect to regional variations, a fact which has not been generally demonstrated.

The boundaries do appear to be modified by the young NW-trending right lateral faults of the Peninsular Ranges, and appear to cross them. The density of REE data points does not allow any independent quantitative estimate of offset on the Elsinore and San Jacinto faults; offsets in excess of several tens of kilometers appear unlikely.

In addition to systematic west to east REE variations, some variations along the axis of the batholith are apparent. The flat to light REE depleted patterns of the San Telmo pluton in Baja California have not yet been recognized to the north. This type of pattern has only been commonly observed in near-trench volcanics of island arcs and in mid-ocean ridge basalts. Their absence to the north does not seem to represent lack of exposures. These observations complement previous studies suggesting the batholith was part of a continental margin magmatic arc in southern California, extending to an oceanic island arc along strike into Baja California (Silver et al., 1963; Silver, 1974).

Two other minor features discriminate between REE data from southern and Baja California. The rocks of the eastern region in Baja California do not attain as high absolute concentrations of light REE as rocks in comparable positions to the north, although fractionations are similar. Also, fewer of the rocks from the central and eastern regions in Baja California have negative Eu anomalies, although sampling may be too sparse to firmly demonstrate this.

REE FRACTIONATIONS AND CONCENTRATIONS AS A FUNCTION OF ROCK BULK COMPOSITION

One of the most surprising features of these data is a lack of correlation of variations in REE characteristics with major variations in bulk composition. The striking uniformity observed in the REE patterns of the western tonalites and quartz gabbros in southern California persists despite the considerable range in the modal and chemical compositions of these rocks. This range is reflected in the more calcic feldspar and higher color indices of the quartz gabbros, and in SiO_2 contents from 53% to 65% (Silver, per. comm.). The magnitude of the Eu anomalies is essentially constant among these rocks, and the small variations in this quantity as well as in the abundances and fractionation of the REE do not correlate with rock type. However, two low-K_2O granodiorites have larger negative Eu anomalies.

This discussion can be extended to the highly siliceous leucogranodiorites of the western region. Some of these rocks may have steeper slopes in the light REE, a reversal in slope in the heavy REE, and most have larger negative Eu anomalies. But not all these rocks have each of these features: some may have one but not others. With the exception of the Eu anomaly, some of these siliceous rocks of Woodson Mountain type have patterns very much like the tonalites. The overall similarity in slope has already been noted. The unusual REE pattern of granodiorite sample 25, with low contents of REE, negligible Eu anomaly, and turned up ends is closely paralleled by the equally unusual pattern for a tonalite (sample 31) just south of the border. Compared to the other rocks with which they are classified, these rocks do not possess unusual petrographic characteristics. This indicates that these features are not the result of extensive high level mineral fractionation processes.

Two low-K_2O granodiorites from the central regions, both from the large La Posta pluton, have REE abundances within the range observed for the tonalites of the same region, and very similar fractionations. In the eastern region in southern California, the granodiorites are distinguished from the tonalites by their higher heavy REE contents and some development of negative Eu anomalies. In Baja California, the eastern granodiorites differ little from the tonalites in REE abundances, and they do not have negative Eu anomalies. Oddly, one of the tonalites does have a negative Eu anomaly.

The only compositional group of rocks which tend to have distinctive REE patterns are the gabbros. If the few analyses of these rocks may be generalized, they always have low concentrations of REE and positive Eu anomalies. The considerable variations in the shape of the patterns may relate to the differences in mineral assemblages in these rocks, which will be discussed in a later chapter.

Thus it appears that with the possible exception of gabbros, few elements of consistency can generally be drawn between REE concentrations and fractionations and bulk compositional variations. A more detailed and quantitative treatment of this topic, addressed within the context of differentiation processes and utilizing previously presented data on the distribution of REE among minerals, is given in the next chapter.

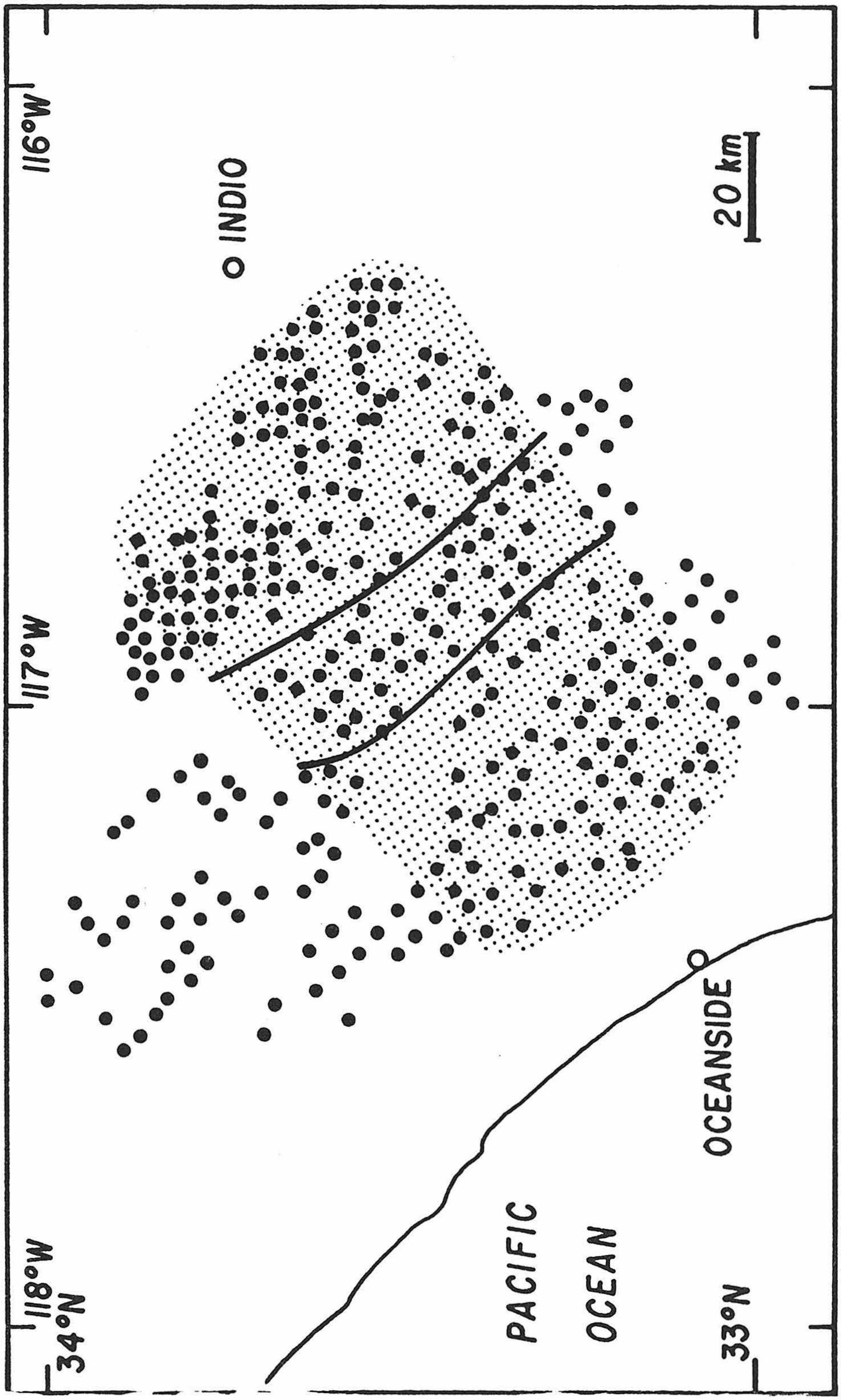
REGIONAL VARIATIONS IN PETROLOGIC PATTERNS

It was noted earlier that petrologic asymmetries are present across the batholith. Among these is the limitation of abundant gabbro (including gabbro and quartz gabbro) and siliceous leucogranodiorite to the western side of the batholith. Larsen (1948) noted in southern California that the boundary between that part of the batholith in which these lithologies were abundant and that in which they are not is about parallel to the long axis and structural trends in the batholith. This boundary appears to be fairly sharp, although not necessarily areally smooth, at least in the parts of the batholith with which I have familiarity.

During the course of this study, it became apparent that the boundary between the western and central regions - a boundary defined solely by a discontinuity in the REE characteristics of the rocks - occurred at about the same place. To pursue this possible correlation, the extensive study of the chemical composition variations of the batholithic rocks of part of southern California by Prof. A. Baird and colleagues, Pomona College, (unpublished, but summarized in Baird et al., 1974a,b) was consulted. This work involved statistical sampling and X-ray fluorescence analysis of plutonic rocks of the northern part of the Peninsular Ranges batholith and adjacent regions in southern California. By recasting chemical data made available by Prof. Baird into differentiation indices or norms, it is possible to obtain an estimation of the regional distribution of the various lithologies.

The areal divisions obtained from the REE study have been superimposed upon Baird's array of sample locations, as seen in Figure 39. The area in which the density of REE analyses is highest is a 70 km wide traverse across southern California, from Oceanside to Indio, as indicated by the shaded pattern in the figure. As a means of indexing the composition of the rocks with a single parameter, the Larsen differentiation index ($1/3\ SiO_2+K_2O+CaO-MgO-FeO_{total}$) was chosen. Frequency histograms of Larsen indices were compiled for rocks of the areas of the three principal REE regions

FIGURE 39. Array of sample locations within the northernmost Peninsular Ranges for the chemical studies of Baird and others. Shaded region is the area of the greatest density of REE analyses, and only those samples occurring within this region are considered in Figure 40. Curved lines indicate areal division into western, central, and eastern REE groups.

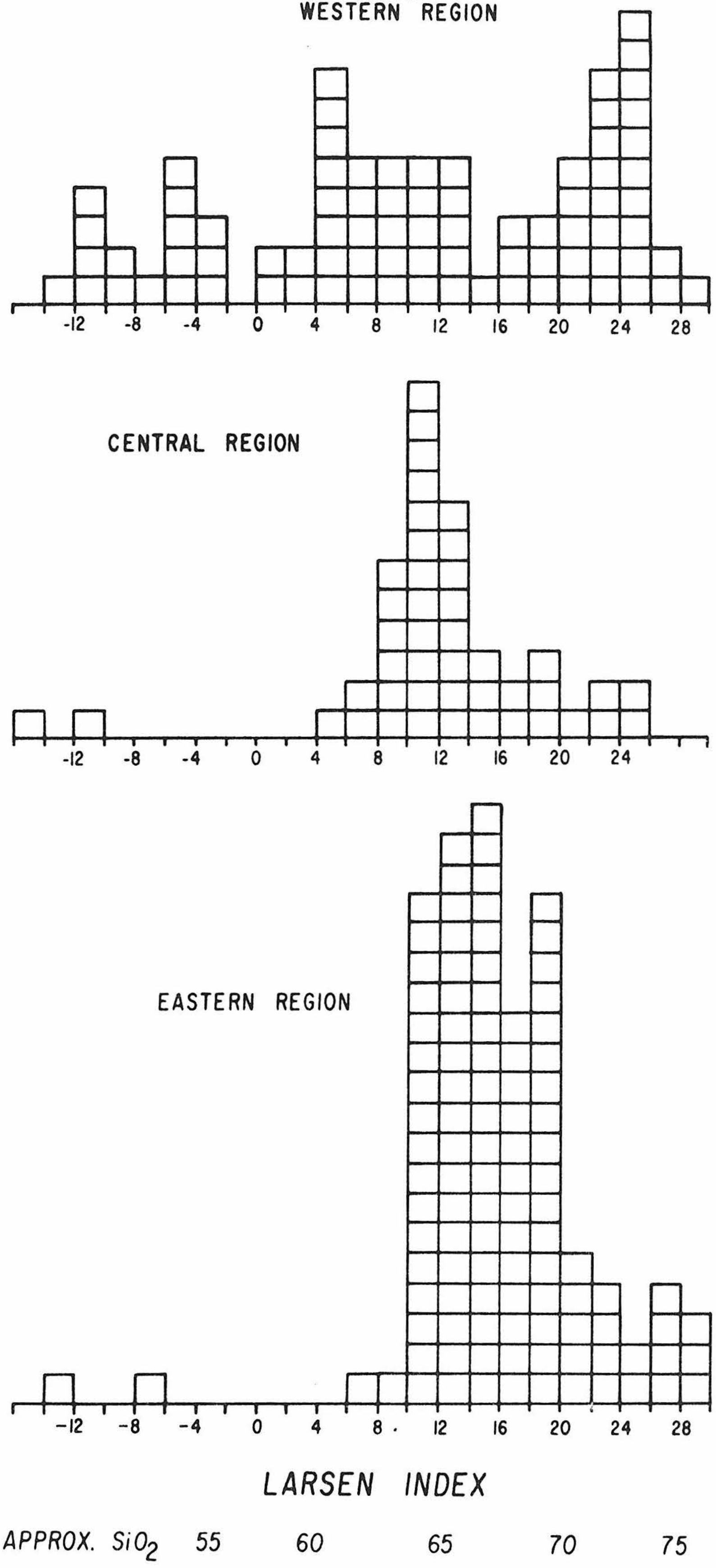

116°W
INDIO
20 km
117°W
118°W
34°N
33°N
PACIFIC
OCEAN
OCEANSIDE

traversed in this profile (Fig. 40).

The histogram of Larsen indices for the plutonic rocks of the western region displays the broad range in bulk composition from calcic gabbros to leucocratic adamellites, representing SiO_2 contents from less than 50% to greater than 75%. The frequency distribution of Larsen indices is not uniform, possessing distinct modalities for gabbros, tonalites, and granodiorites and adamellites. A distinct minimum is present between the tonalites (Larsen indices up to 14) and the siliceous leucogranodiorites of the Woodson Mountain type and leucocratic adamellites (most with Larsen indices between 20 and 26). This diagram emphasizes the local abundance of these quartz-rick rocks and complements Larsen's map of the northwestern corner of the batholith. Although plutons do display variations in composition and the distinction among rock types is in places arbitrary, Larsen's estimates of the areal abundances of the various lithologies also indicate a significant areal underrepresentation of rocks with compositions intermediate between tonalite and siliceous leucogranodiorite.

This compositional distribution is contrasted against the histogram of Larsen indices of rocks from the central region. Most rocks of the central region are represented by a much more restricted range of indices, with very few gabbros present and a much reduced number of quartz-rich rocks. The distribution is unimodal with a strong maximum corresponding

Figure 40. Frequency histograms of Larsen indices (1/3 SiO_2 + K_2O -CaO -MgO - FeO_{total}) for the REE-defined western, central, and eastern regions in the northernmost Peninsular Ranges. The Larsen indices are calculated from unpublished chemical analyses provided by Prof. A. Baird, Pomona College.


WESTERN REGION
-12
-8
-4
0
4
8
12
16
20
24
28
CENTRAL REGION
-12
-8
-4
0
4
8
12
16
20
24
EASTERN REGION
-12
-8
-4
0
4
8
12
16
20
24
28
LARSEN INDEX
APPROX. SiO_2
55
60
65
70
75

to the ubiquitous tonalites and lesser low-K_2O granodiorites. Quartz gabbros are essentially absent. The tonalites of this region tend to have lower color indices, which is reflected in their somewhat higher Larsen indices. The form of the frequency distribution for the rocks of the eastern region is similar to that of the central region, but shifted to higher Larsen indices. Tonalites and low-K_2O granodiorites are subequal in abundance, and a lesser but significant number of more siliceous lithologies are present. As in the central region, gabbros are rare.

The histograms of compositional abundance support the significant correlation between the changes in REE character between the western and central regions and the compositional variations indicated by Larsen's map. Information about rock distribution in San Diego County and northern Baja California suggest the correlation persists, although somewhat less rigorously to the south (Silver, per. comm.). It appears that the combination of source region characteristics and petrogenetic processes which produced the petrographic diversity of the west side of the batholith also controlled its REE character. The lithologic distinction between the central and eastern regions is less pronounced.

REE CHARACTERISTICS COMPARED TO GEOCHEMICAL, ISOTOPIC, AND AGE PATTERNS

General Statement

The Sr concentrations, $^{87}Sr/^{86}Sr$ initial ratios, oxygen isotopic compositions and ages of the granitic rocks have been shown by others to vary systematically across the batholith. Sr concentrations, initial $^{87}Sr/^{86}Sr$ ratios, and $\delta\,^{18}O$ values generally increase to the east and ages decrease. In as much as the REE patterns of the rocks vary across the batholith, some significant correlations between the REE patterns and each of these parameters are indicated. However, each correlation has its own qualities, and ultimately its own significance.

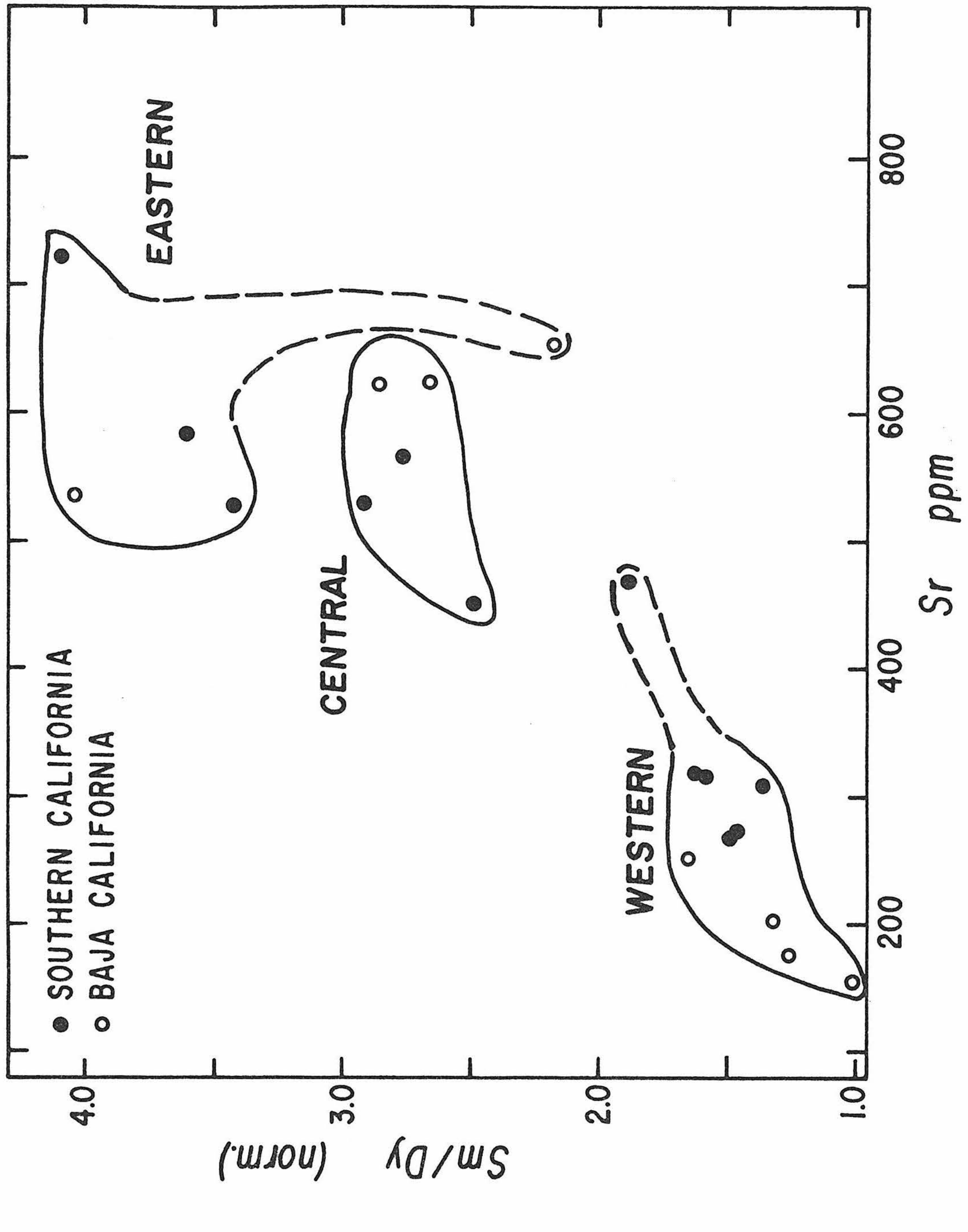
REE Fractionation Versus Sr Concentration

In the comparisons of the variations of REE fractionation patterns with compositional abundances it was observed that the pattern slopes in the middle to heavy REE region displayed little dependence on lithology. A measure of the slope in this region will therefore provide an index of REE fractionation largely free of the influence of bulk composition. For this purpose, a chondrite normalized Sm/Dy ratio is selected as it covers the most important part of this region.

In Figure 41, the normalized Sm/Dy ratio is plotted against Sr concentration for all tonalite and quartz gabbro samples for which both REE and Sr have been determined. Consideration has been limited to these lithologies as Sr concentration is somewhat dependent on lithology. REE fractionation and Sr concentration exhibit a good correlation on this plot, and a grouping by regions is observed. The quartz

Figure 41. Chondrite-normalized Sm/Dy ratio versus Sr concentration for quartz gabbros and tonalites in the Peninsular Ranges. The samples are grouped by geographic region. The symbols differentiate between samples from southern and Baja California. Sr concentrations are from unpublished data of Early and Silver.

The western and central groups show limited overlap in Sr concentration, whereas the central and eastern groups have considerable overlap. An eastern tonalite (sample 37) appears anomalous in Sm/Dy ratio. Note that within the western and central groups, this limited sampling suggests a north-south distinction in Sr content.


● SOUTHERN CALIFORNIA
○ BAJA CALIFORNIA
EASTERN
CENTRAL
WESTERN
Sm/Dy (norm.)
4.0
3.0
2.0
1.0
Sr ppm
200
400
600
800

gabbros and tonalites of the western regions have low Sm/Dy ratios and Sr contents and are tightly grouped with the exception of a transitional tonalite (sample 22). Tonalites of the central and eastern regions tend to have progressively higher values of both quantities.

There appears to be a gap in the distribution of points between the western and central regions, even though there is not a geographic gap among the sample locations. In fact, in southern California there is a higher density of samples near the boundary as a result of efforts to study this transition. This gap is the expression of not only an abrupt change in REE characteristics between the western and central regions, but also in Sr concentration for all but one of the tonalites. This appears particularly pronounced for the samples from Baja California, but more extensive sampling is required to adequately document this.

The tonalites from the eastern region in southern California plot along trend, but one of the eastern tonalites from Baja California is anomalous. There is substantial overlap in Sr concentration among the tonalites from the central and eastern regions.

Thus it appears the western and central regions have distinct REE fractionation and Sr concentration characteristics which persist along the length of the batholith segment studied. The differences between the central and eastern groups, in terms of these parameters, are less pronounced.

REE Fractionation Versus $^{87}Sr/^{86}Sr$ Initial Ratio, $\delta^{18}O$, and Age

A plot of Sm/Dy ratio against $^{87}Sr/^{86}Sr$ initial ratio (Fig. 42) indicates a general correlation between these parameters, but with some very different characteristics. In contrast to the gaps in both REE fractionation and Sr concentration data between the western and central regions, there is a very considerable overlap in their Sr isotopic compositions. Conversely, the eastern region appears isotopically distinct from the other regions, with more radiogenic compositions.

Thus Sr isotopic composition and REE fractionation do not correlate in the same manner as Sr concentration and REE fractionation. A corollary of this is that Sr isotopic composition and Sr concentration also do not correlate in the same way. A portion of the noncorrespondence in these data can be attributed to north-south differences along the batholith axis. This is evident in the observation that some of the most pronounced discrepancies in Figure 42 are among points from southern California and Baja California; rocks from the same REE region north and south of the border appear isotopically distinct. Better correlations exist in local profiles. Viewed in plan (Fig. 43), isopleths of constant initial $^{87}Sr/^{86}Sr$ ratio and the boundaries between the REE-defined regions have local alignment, but deviate substantially from complete parellelism along the length of the

Figure 42. Chondrite-normalized Sm/Dy ratio versus initial $^{87}Sr/^{86}Sr$ ratio for quartz gabbros and tonalites in the Peninsular Ranges. The samples are grouped by geographic region. The symbols differentiate between samples from southern California and Baja California. Sr isotopic values are from unpublished data of Early and Silver. Note the large overlap in initial $^{87}Sr/^{86}Sr$ ratio between the western and central groups, but the distinctly higher values for the eastern group. Within this limited sampling, the rocks from Baja California are less radiogenic than corresponding samples from southern California in the western and central regions.

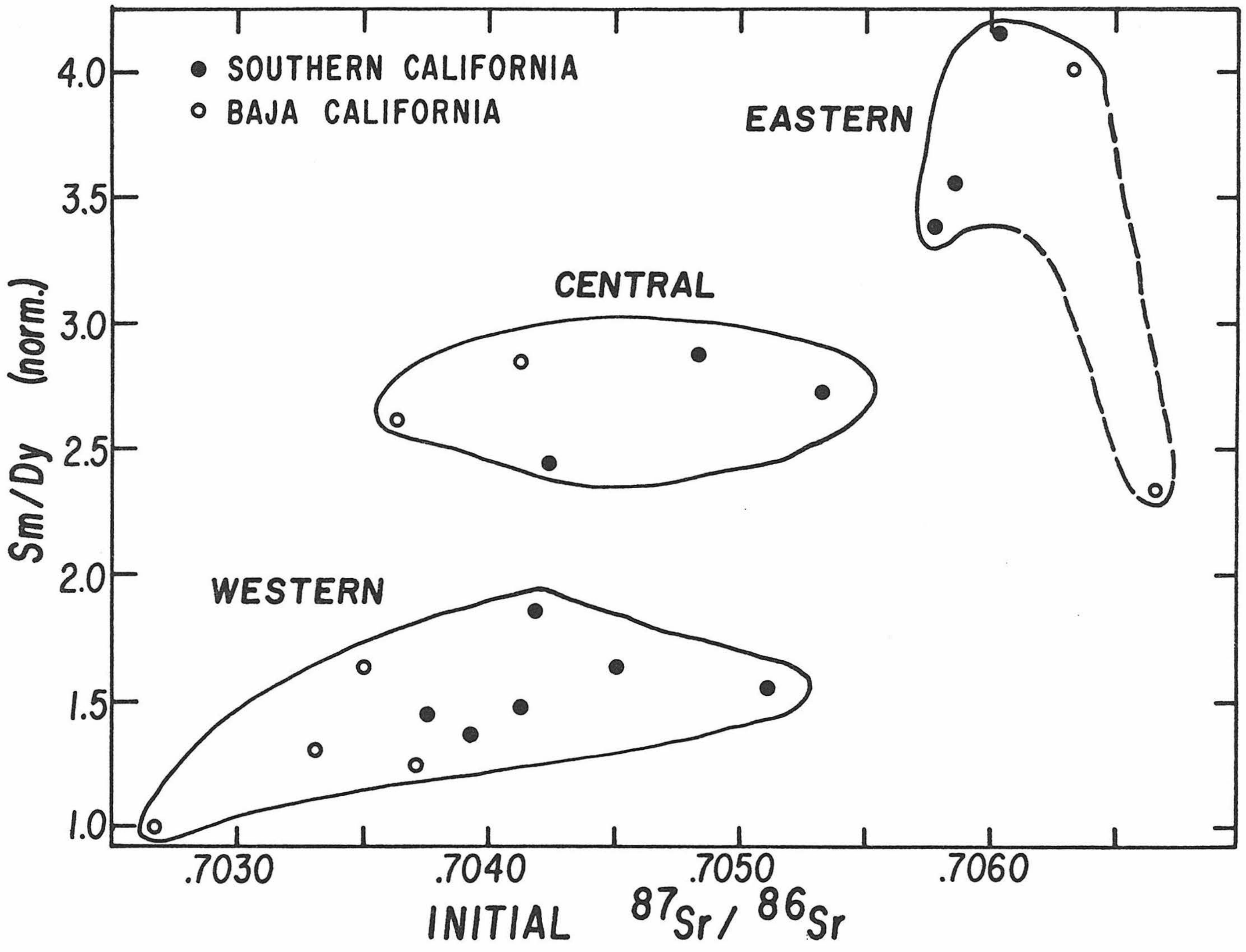
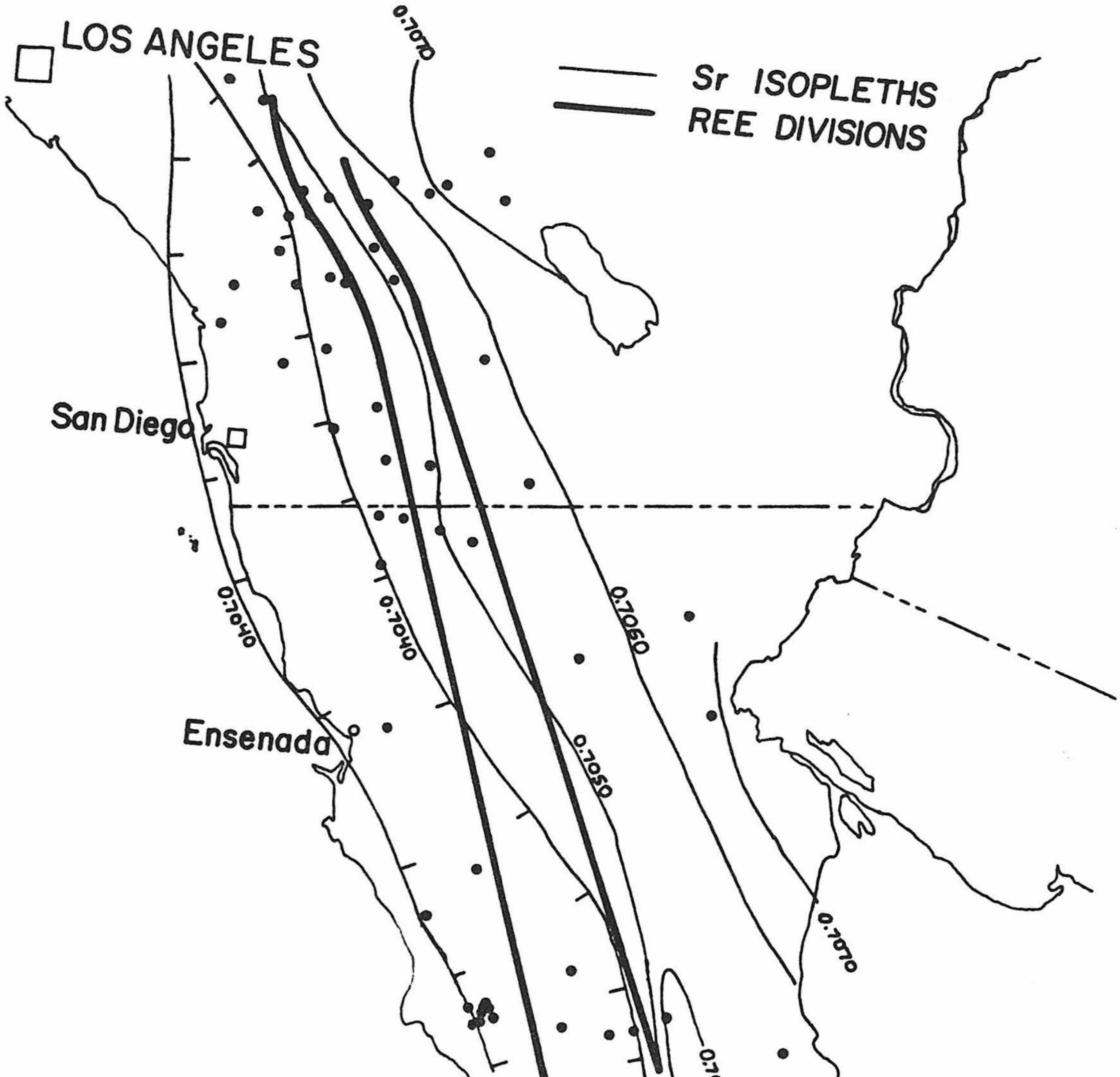

● SOUTHERN CALIFORNIA
○ BAJA CALIFORNIA
EASTERN
CENTRAL
WESTERN
Sm/Dy (norm.)
4.0
3.5
3.0
2.5
2.0
1.5
1.0
.7030
.7040
.7050
.7060
INITIAL $^{87}Sr/^{86}Sr$

Figure 43. Isopleths of constant initial $^{87}Sr/^{86}Sr$ ratio and the boundaries between the REE-defined regions in the Peninsular Ranges. The sample locations are those analyzed for REE. The isotopic isopleths are determined on a larger set of samples, indicated in Figure 2. Note the general parallelism of these varitions over local profiles, but their deviation along the length of the peninsula.


LOS ANGELES
Sr ISOPLETHS
REE DIVISIONS
0.7070
San Diego
Ensenada
0.7040
0.7040
0.7060
0.7050
0.7070
0.7060

peninsula, as do many other important parameters.

The along-axis discrepancies observed for the correlation between REE fractionation and $^{87}Sr/^{86}Sr$ initial ratio is also true for the correlation between REE fractionation and $\delta^{18}O$. This can be viewed as largely a consequence of the general correlation between the isotopic compositions of Sr and O (Taylor and Silver, 1978). The discontinuity in the variations of $\delta^{18}O$ of the batholithic rocks, which paralells the axis of the batholith and does not have a counterpart in the Sr isotopic variations, does not correlate simply to the discontinuities in the REE variations. In most of southern California and northernmost Baja California, the $\delta^{18}O$ step and the abrupt transition in REE characteristics from the western to central regions coincide, or very nearly do so. Further to the south, however, the two diverge, and the western and central regions no longer have contrasting $\delta^{18}O$. At the latitude of the southernmost samples, the step in $\delta^{18}O$ may more closely correspond to the boundary between the eastern regions.

The eastward decrease in ages of the intrusives across the Peninsula suggests that a general correlation between REE fractionation and age may exist, just as with all the previously discussed properties. But as with the isotopic variations, there is considerable deviation from a simple correlation. The step in the age pattern (Silver, per. comm.) does not coincide with a single REE boundary transition, but discontinuities of ages and $\delta^{18}O$ appear to share a common narrow belt parallel to the batholith axis (Taylor and Silver, 1978). This belt is approximately in the region of the central REE fractionation zone.

SUMMARY

REE patterns of plutonic rocks display large, systematic but discontinuous variations from west to east across the Peninsular Ranges batholith, transverse to the long axis and structural trends of this magmatic arc. The consistent, slightly fractionated patterns of the western rocks make an abrupt transition to middle and heavy REE fractionated and depleted patterns for rocks in the central region of the batholith. Further to the east, a second well defined transition to strongly light REE enriched rocks occurs. The slopes of the REE patterns within each of these regions are largely independent of rock type, with the exception that certain gabbros may show divergent trends.

These variations in REE fractionation generally correlate with other transverse asymmetries in petrologic, geochemical, isotopic, and geochronologic parameters. Detailed comparisons reveal distinct and genetically significant divergences. An excellent correspondence exists among REE fractionation, Sr concentration, and petrologic characteristics in the western and central regions, suggesting a common explanation for their origin. The eastern region can be clearly distinguished in both REE fractionation and initial $^{87}Sr/^{86}Sr$ ratio from the central and western regions. The discontinuities in $\delta^{18}O$ and age distributions in the batholith do not consistently correlate with the REE discontinuities; they are locally divergent by the dimensions of one or two plutons.

The REE relationships provide critical information for the evaluation of the origin of all these variations and for understanding of the mechanisms, processes, and sources participating in the generation of the batholithic magmas. The REE data collected on minerals in this study provide important insight into the behavior of REE during crystallization of these rocks and during any high level differentiation processes. All of the above information is applied in the following chapter to a discussion of the problem of the origins of the REE variations.

CHAPTER 3

ORIGIN OF THE REE VARIATIONS

HIGH LEVEL PROCESSES

THE ROLE OF CRYSTAL FRACTIONATION

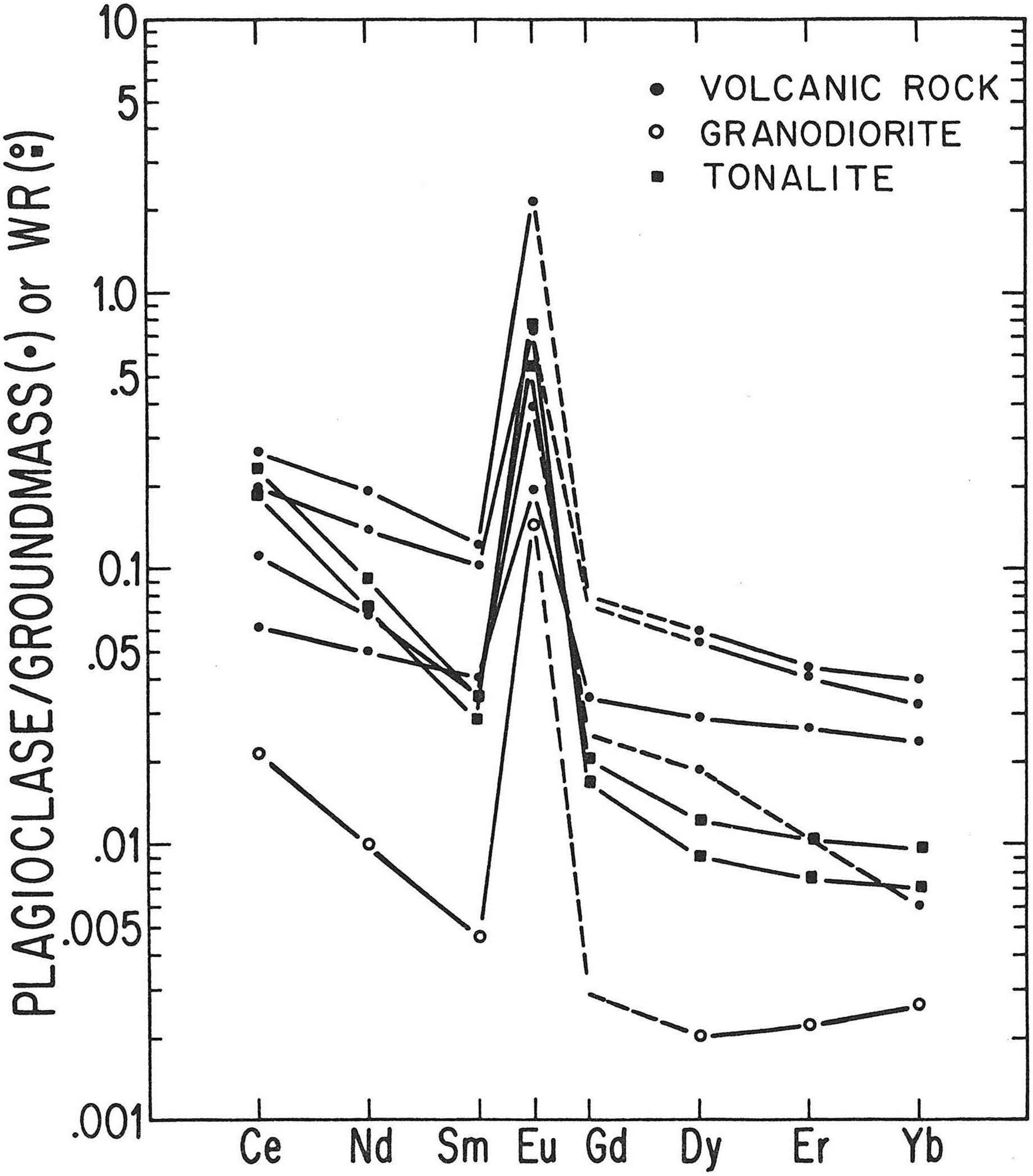
Regional Variations

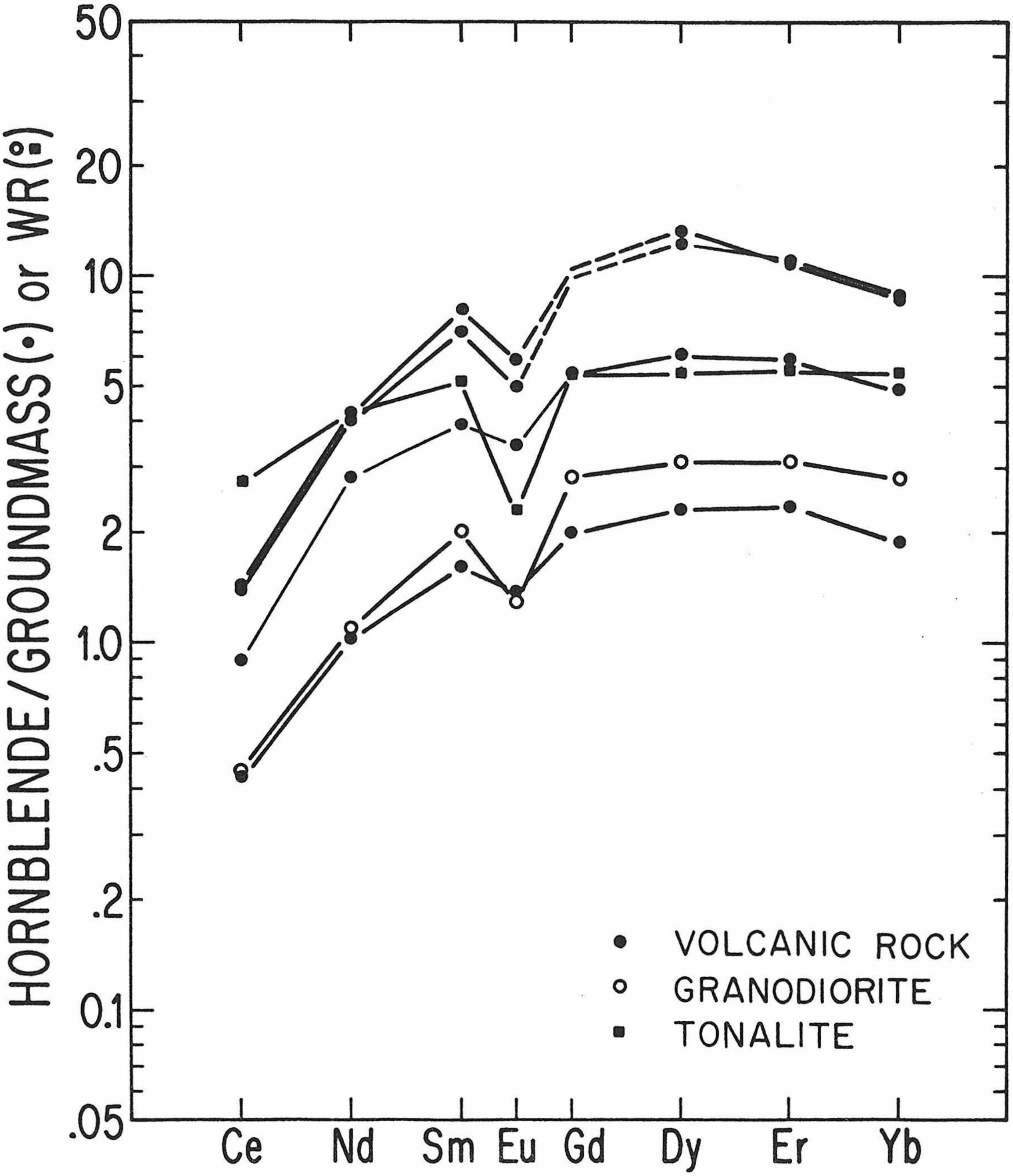
Major Minerals

Textural observations and bulk chemistry indicate the major early crystallizing phases in the abundant tonalites are plagioclase, hornblende, and quartz. This is supported by the common presence of these phases as phenocrysts in volcanic rocks of related chemistry, and their occurrence as liquidus or near liquidus phases in experimental studies conducted at crustal pressures (e.g. Eggler and Burnham, 1973). The general plagioclase/groundmass and hornblende/groundmass REE distributions have been measured by several workers in a wide variety of volcanic rocks, yielding consistent fractionation patterns for the trivalent REE which are taken to approximate equilibrium fractionations. The whole-rock-normalized REE patterns for plagioclase and hornblende in plutonic rocks from the Peninsular Ranges batholith are quite similar in form to these general fractionation patterns (Figs. 44 and 45). This consistent REE behavior supports the operation of straightforward magmatic crystallization processes for the batholithic rocks and lends confidence to the application of REE fractionations in interpretation of the courses of plutonic crystal differentiation processes. It is noted that the absolute values of the mineral/melt distributions are much more variable than their shape. The

Figure 44. Comparison of selected plagioclase/groundmass REE distributions in volcanic rocks and plagioclase/whole rock REE distributions in a western tonalite (sample 1), a central tonalite (sample 50), and an eastern granodiorite (sample 19). These patterns have a strong similarity in shape, but the granodiorite plagioclase displays much lower absolute values.

Figure 45. Comparison of selected hornblende/groundmass REE distributions in intermediate to felsic volcanic rocks and hornblende/whole rock REE distributions in a western tonalite (sample 1) and an eastern granodiorite (sample 19). The volcanic and granodioritic hornblendes have remarkably similar REE distribution pattern shapes, but the tonalitic hornblende pattern is significantly flatter.


PLAGIOCLASE/GROUNDMASS(•) or WR(▪)
10
5
1.0
.5
0.1
.05
.01
.005
.001
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
● VOLCANIC ROCK
○ GRANODIORITE
■ TONALITE


HORNBLENDE/GROUNDMASS(•) or WR(○■)
50
20
10
5
2
1.0
.5
.2
0.1
.05
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
● VOLCANIC ROCK
○ GRANODIORITE
■ TONALITE

following arguments will therefore be developed primarily on the relative fractionations of REE, and only secondarily on the absolute values.

Plagioclase strongly discriminates against all the REE, except Eu, relative to melt. In spite of its great abundance (40%-65%), plagioclase holds only a small proportion of the total REE in almost any rock in the batholith. This is well displayed, for example, in the granodiorite from near Indio, described earlier, in which plagioclase contributes less than 1% of all trivalent REE to the whole rock. Plagioclase will accept the lighter REE more readily than the heavy REE, but does not strongly fractionate among the heavy REE (Fig. 44). Therefore, removal of plagioclase by crystallization will not lead to significant relative fractionations of the heavy REE. The very low fraction of all REE but Eu in plagioclase indicates subtraction of plagioclase will increase the concentration of all trivalent REE in a nearly uniform manner. In effect, plagioclase is a diluent for these REE in batholithic rocks.

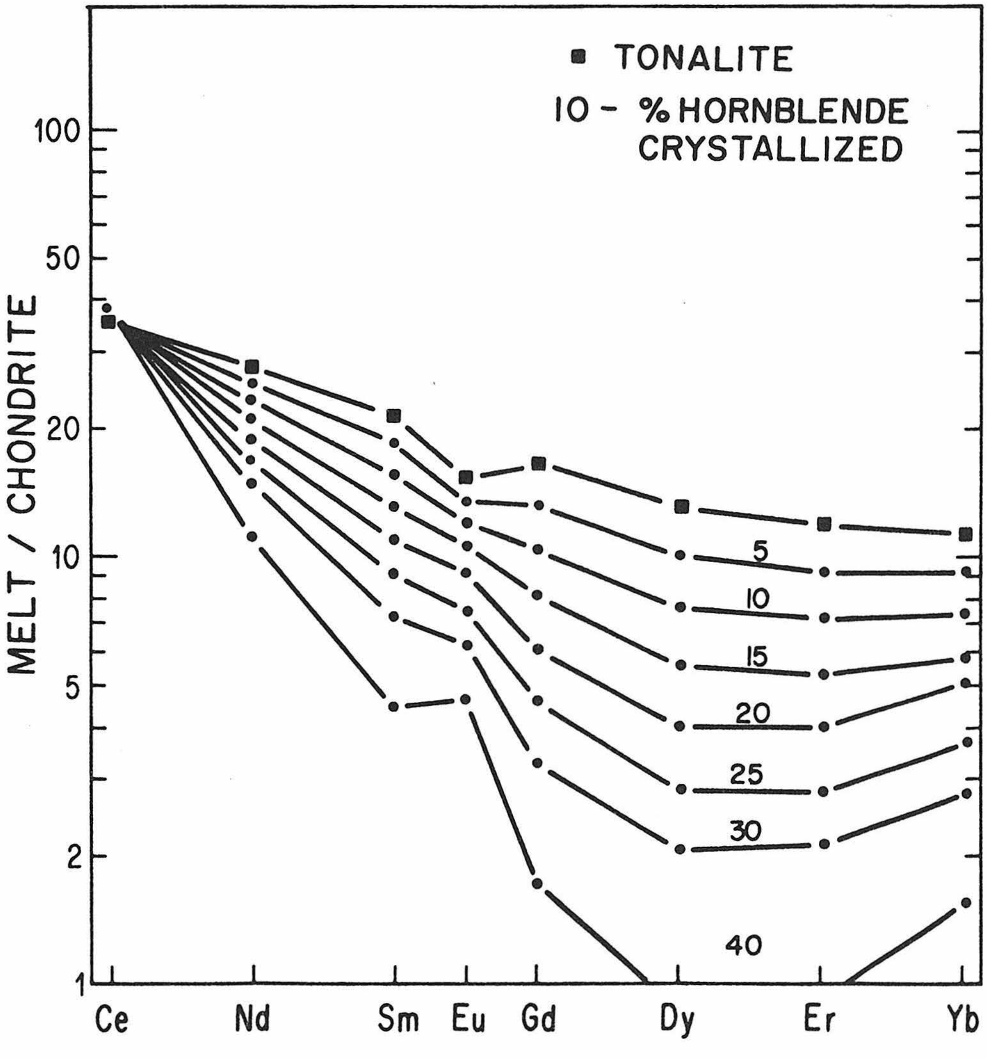
Among the major rock-forming minerals in the batholith, hornblende is capable of fractionating the REE most effectively. Hornblende can be an important site for the REE in these rocks. In the case of the granodiorite near Indio, hornblende contributes up to 15% of the heavier REE to the whole rock (Fig. 17) and is the only major mineral holding a substantial proportion of REE. In the tonalite near Val

Verde, the weight percent of hornblende in the rock (11.8%) indicates that roughly half of the middle (except Eu) and heavy REE of the whole rock are contained in this mineral. The mineral data of Towell et al. (1965) suggest that most of the REE in the sample of San Marcos gabbro which was analyzed was also present in hornblende. Removal of even relatively small proportions of hornblende can lead to modifications in the REE character of the remaining system.

Hornblende/groundmass and hornblende/whole rock distributions are consistently concave-down with steep positive slopes in the lighter REE. Subtraction of hornblende will produce a complementary negatively sloped, concave-up light REE pattern in the residual melt. These effects are illustrated in Figure 46 which displays the REE patterns resulting from progressive crystallization of hornblende from a western tonalite (sample 1) according to the Rayleigh fractionation law. In this particular example, crystallization of hornblende results in depletion of most REE in the melt. In other cases this will vary with the absolute value of the distribution coefficients. The pertinent point here, however, is the analysis of the effect of hornblende crystallization on the <u>shape</u> of the REE pattern of the residual liquid. The increase in slope across the light REE in the liquid for progressive hornblende crystallization results from the steep positive slopes for the light REE in the hornblende distributions. These distributions flatten through the middle to

FIGURE 46. The effect of progressive hornblende crystallization on the REE pattern of a western tonalite (sample 1). Crystal fractionation is performed according to the Rayleigh fractionation law using the following mineral/melt distribution coefficients (from Arth and Barker, 1976): Ce, .90; Nd, 2.80; Sm, 3.99; Eu, 3.44; Gd, 5.47; Dy, 6.19; Er, 5.94; Yb, 4.90.

As hornblende is progressively removed by crystallization, the increased slope of the light REE and the development of positive Eu anomalies and spoon-shaped patterns in the middle to heavy REE are notable.


■ TONALITE
10 - % HORNBLENDE CRYSTALLIZED
MELT / CHONDRITE
100
50
20
10
5
2
1
5
10
15
20
25
30
40
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

heavy REE, resulting in much more limited relative fractionation among these elements. As is seen in Figure 46, sizable relative fractionation across the middle to heavy REE occurs only for large amounts of hornblende crystallization. Dy and Er are more strongly depleted than lighter or heavier REE, producing a concave-up, spoon-shaped pattern in this region. Positive Eu anomalies are produced in the derived liquids.

Comparison of these effects with the observed regional variations in REE character does not suggest that hornblende crystallization has played an important role in producing the regional trends. The transition in REE character from the western to central regions is a strong fractionation in the middle to heavy REE (e.g. Fig. 29), a region where hornblende is least effective or incapable of producing relative fractionations. The minimal changes in slope across the light REE between the western and central tonalites does not correspond well to the steep slopes expected from the hornblende distributions.

The transition from the central to eastern regions involves an increase in abundance and fractionation of the light REE and a slightly increased fractionation of the middle to heavy REE (e.g. Fig. 30). The variations in the light REE might be related to hornblende fractionation, but the increased slope across the middle to heavy REE cannot be explained the same way. Spoon-shaped patterns expected from hornblende crystallization are not generally observed. From the range of observed variations of the behavior of REE in hornblende, its fractionation alone does not appear capable of producing the observed regional variations.

The other major minerals in the intermediate to felsic batholithic rocks are biotite, quartz, and in some cases alkali feldspar. The unfractionated biotite/whole rock REE distribution in the granodiorite near Indio, represented in shape by the biotite pattern in Figure 17, is similar to biotite/groundmass distributions in volcanic rocks. Crystallization of quartz, another diluent for REE, will simply increase the concentration of all REE in the melt. REE distributions for alkali feldspar are generally comparable to plagioclase, and the same arguments apply. In gabbros and quartz gabbros, pyroxenes and in particular clinopyroxene appear as major phases. Clinopyroxene shares a similarly shaped REE distribution with hornblende, and its effect on the REE patterns of derived liquids can be expected to be similar.

On the basis of the REE fractionation characteristics of these minerals, crystal fractionation involving assemblages of the major minerals do not provide an explanation for the regional REE variations.

An independent argument leading to the same conclusion comes from comparing the bulk composition and major mineral assemblages of the tonalites across the batholith. The similarity in bulk composition of tonalites possessing very different REE characteristics and containing hornblende as the only significant REE-bearing major phase constrains high-level fractionation processes to a limited role. The chemical composition and mineral assemblages of selected tonalites may be compared in Appendix I.

Minor Minerals

The precipitation of minor minerals with high concentrations of REE may have a profound effect on the REE characteristics of the melt. The mineral study of the granodiorite near Indio revealed that accessory phases such as allanite and sphene can contain most of the REE in some rocks. Extended observations of accessory minerals in a broad sampling of batholithic rocks by Prof. Silver (per. comm.) indicate this will be generally true of rocks from the central and eastern regions of the batholith (and probably will apply to many other rocks as well). Typically, eastern rocks have highly fractionated patterns with low heavy REE and high light REE contents. Most of the allanite and sphene crystals occur as large, euhedral, independent grains, and are apparently among the earlier minerals to crystallize. If most of the allanite and sphene precipitated from the melt early, it may become strongly depleted and possibly fractionated in REE after a moderate degree of crystallization.

The study of REE distributions among minerals in the granodiorite near Indio is very relevant to this discussion.

The observed REE zoning trends in allanite and sphene (Figs. 18 and 19) suggest that the REE content of melt and perhaps the chemical potential of REE ions in the melt decreased as these minerals crystallized. The argument for a decrease in REE concentration in the melt is firmest for sphene, as explained below, where the REE are not major constituents which influence its stability. As crystallization proceeds in a magma, some change in the partitioning of REE between sphene and melt is anticipated. The melt will progressively become more SiO_2 rich and CaO, MgO, and FeO poor. Deductions concerning compositionally related modifications of melt structure presented earlier indicate that partitioning of REE into Ca bearing minerals such as sphene will be enhanced during this process. If the concentration of REE remained the constant in the melt during crystallization, it might be expected that the REE contents of the sphene would increase core to rim. The roughly 50% decrease in Ce concentration from core to rim in sphene is perhaps the minimum extent that Ce actually decreased in the melt during the growth of this grain.

The zoning in allanite is more difficult to interpret as the light REE are major stoichiometric constituents whose concentrations affect the stability field of this phase. The antithetic increase in Th from core to rim suggests that allanite attempted to maintain its growth by substituting other components to replace the less available REE. This is consistent with the arguments based on melt structure; both Th^{4+} and the trivalent REE would be strongly excluded from a SiO_2-rich melt. These arguments may offer some explanation for the apparent lack of late growth of allanite. As crystallization proceeds, allanite can remain saturated in the melt with progressively lower absolute concentrations of REE in this melt. The energetics of coordinating these cations in the melt may become unfavorable as polymerization increases, thereby tending to maintain the chemical potential of these components necessary for allanite saturation at lower concentration levels. During the last stages of crystallization, the very small amounts of allanite components available for precipitation result in little observed late growth. The core to rim decrease in light REE in allanite would seem to indicate that the chemical potential of these elements actually decreased during their growth, if other effects such as temperature are not sufficient to explain these variations.

In all of these discussions, of course, one must bear in mind that local melt-crystal kinetics may be responsible for all or part of the observed zoning.

Additional evidence suggesting that the REE contents of the magma decreased substantially as the granodiorite solidified is found in the whole-rock-normalized mineral REE patterns. Plagioclases from the three plutonic rocks studied here have whole-rock-normalized pattern shapes, including the Eu anomaly, which are comparable to volcanic data (Fig. 44).

The absolute distribution values are lower, however, particularly for the plagioclase in the granodiorite. There are several possible explanations for this. The plagioclases in the plutonic rocks almost certainly crystallized and equilibrated at lower temperatures than the volcanic plagioclases, perhaps allowing them to more effectively exclude unwanted REE. The rapid growth of volcanic plagioclases may not have allowed an equilibrium distribution to be obtained, in which case the plutonic plagioclases might be a better indicator of the true distribution. However, little explanation for the very low distribution in the granodiorite can be offered from the above.

A major difficulty in such comparisons of whole-rock-normalized patterns with mineral/groundmass patterns is that in the plutonic case, it must be assumed that the whole-rock values are representative of the melt. This may be true only early in the crystallization history of the rock, if even then. Early-formed minerals may yield whole-rock-normalized patterns closely approximating mineral/groundmass distributions, but significant portions of minerals in plutonic rocks have grown from progressively changing melts, probably with much different REE abundances than the whole rock. The much lower whole-rock-normalized plagioclase patterns for the granodiorite as compared to the volcanic data and to plagioclases from the two tonalites studied here suggests that the early appearance of sphene and allanite depleted the melt in

REE at an early stage. The lack of significant Eu anomalies in the whole-rock-normalized shapes of sphene and allanite (seen in Fig. 17) supports the textural evidence of largely early crystallization of these minerals in the granodiorite. The two tonalites do not have parageneses involving early sphene or allanite.

Whole-rock-normalized patterns can be used advantageously to evaluate whether much relative fractionation of REE has occurred during the crystallization of the granodiorite. The general similarity of the whole-rock-normalized REE distributions of the major minerals in the granodiorite to observed mineral/groundmass distributions in volcanic rocks indicates that the REE pattern of the melt did not change greatly in shape during crystallization. This is most readily seen in the minerals plagioclase, hornblende, biotite, and alkali feldspar, whose growth covers the entire crystallization history of this rock. It can be concluded that large variations in the REE patterns of the progressive residual melts have not occurred even though REE-rich accessories seem to have strongly modified REE levels in the melt. Further study of such features is desirable. Whole-rock-normalized REE patterns of plutonic minerals such as sphene and allanite which make dominant contributions to their whole rocks may not reflect equilibrium mineral/melt distributions. Observations on mineral/groundmass distributions in volcanic rocks for these minerals would provide additional, independent comparisons.

If some elements of high-level crystal fractionation processes can be adequately described by these observations and by the considerations of crystallization and the distribution of REE within these few rocks, these processes do not appear to be an effective means of producing the regional variations in REE.

Finally, an independent argument against the production of the observed regional variations by precipitation of minor phases can be made. While it is possible to strongly deplete a melt in any component by crystallization of a small amount of an accessory phase rich in that component, the only way to <u>increase</u> the concentration of a component in a melt through crystal fractionation is to remove major quantities of mineral phases poor in that component. To increase the residual concentration two-fold, a minimum of 50% of a melt would have to crystallize. The close correspondence between REE fractionation and Sr concentration in the tonalites of the western and central parts of the batholith (Fig. 41) suggests a shared response to the same differentiation mechanism, but the more than two-fold increase in Sr concentration cannot be explained by selective crystallization of small amounts of any minerals. Bulk compositional constraints appear to rule out the possibility of extensive crystal fractionation of major minerals to relate these rocks. By similar arguments, the light REE enrichments of the eastern rocks are not the result of these processes.

REE Variations with Bulk Chemistry In Each Linear Region

Introduction

The variations in REE characteristics among rocks of the same linear region provide insights into and constraints on the differentiation processes responsible for the lithologic diversity. The purpose of this section is to evaluate the extent to which high-level crystal fractionation processes may have been responsible for these variations.

Some limits can be placed on the dimensions of a differentiating system. Considerable variations in initial $^{87}Sr/^{86}Sr$ ratio occur in each of the REE-defined regions, and indicate that all rocks within these regions could not have been consanguineous. Across the Peninsula, the systematic gradients in Sr isotopic composition appears to limit the size of a hypothetical differentiating system to the dimensions of a few plutons at the very most. Along the length of the Peninsula, little constraint is provided. These rough dimensions are also supported by the size and shape of Rb-Sr fractionation domains in the Peninsular Ranges (Silver and Early, 1977). REE characteristics tend to be consistent within each defined region, and consideration below is given to individual regions.

In most of the discussion that follows, it is assumed that the bulk composition and REE patterns of the sampled rocks are reasonably accurate reflections of melt characteristics. These assumptions are supported by several lines of evidence. Both the REE patterns and the bulk composition of most of the quartz bearing rocks studied here are similar to common orogenic volcanic rocks. The studies of REE distribution among minerals in plutonic rocks reported here and in Towell et al. (1965) indicate strong fractionation of REE among the minerals. If the bulk composition and REE contents of these rocks were controlled or affected to any significant degree by crystal accumulative processes, the uniformity in the REE patterns of such rocks as the western tonalites and quartz gabbros would be difficult to understand in light of their considerable range in mineral composition. These and other features will, however, be considered in more detail in the discussions here.

Western Region

The quartz gabbros and tonalites of the western region in southern California have remarkably uniform REE characteristics over a significant range in major element chemistry (Fig. 26). This places a very powerful constraint on the differentiation processes which relate them. The variations in major chemical components in these rocks are highly correlated, as evidenced by tight curvilinear arrays in variations diagrams such as those presented by Larsen (1948) and in Harker-type oxide-oxide plots. Some form of simple, systematic differentiation, strongly controlled by phase equilibria, is implied.

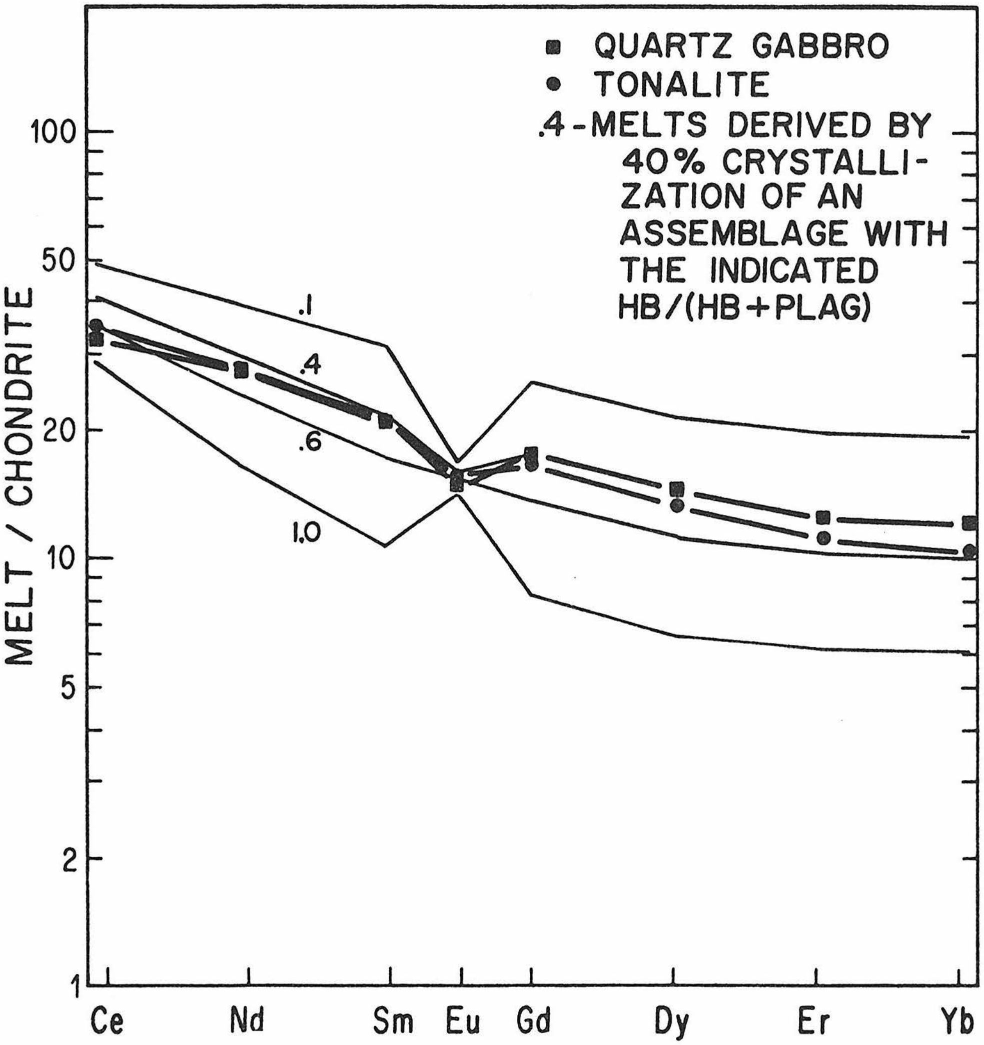
Two rocks which span the larger part of this range are a quartz gabbro (sample 9) and a tonalite (sample 1), common examples of the San Marcos gabbro and Bonsall tonalite of Larsen (1948). These two rocks are used below to illustrate the effects of crystal fractionation on REE patterns. Chemical analyses and petrographic descriptions of these samples are found in Appendix I. Early crystallizing minerals in these rocks are plagioclase, hornblende, clinopyroxene, and, probably, opaque minerals. The shape of the distribution coefficient patterns for the REE for the three silicates are well established, and Fe-Ti oxides will simply be diluents. It appears that there is sufficient information from REE distributions and bulk and mineral chemistry to evaluate crystal fractionation processes.

Utilizing the bulk chemistry of these rocks and a range of reasonable mineral chemistries for the early crystallizing phases (from Larsen, 1948, p. 164, and from my work in linear mixing calculations not presented here) it is indicated that the quartz gabbro composition can only yield approximately 60% to 65% of its mass as a tonalitic magma. A substantial amount of crystal removal is implied. Under conditions when H_2O is present, the assemblage removed probably consists of subequal amounts of plagioclase and hornblende or plagioclase, hornblende, and clinopyroxene. For dry conditions, Larsen (1948) calculated olivine and clinopyroxene as the hypothetical mafic minerals, but it is unlikely that significant quantities of olivine could crystallize from quartz oversaturated compositions such as the quartz gabbro. These early formed assemblages of minerals do resemble the gabbros of the batholith, and the possible relation between the gabbros and the quartz-oversaturated rocks is considered later in this section. These major minerals will dominate the behavior of the REE in high-level differentiation schemes. Minor REE-rich phases such as allanite and sphene are not considered important here because when present, they do not appear to crystallize early from the western quartz gabbros and tonalites.

Calculations summarized graphically in Figure 47 illustrate the effects on the residual liquid REE pattern when appropriate quantities of plagioclase and hornblende are removed by crystal fractionation. An approximation to mineral/melt distributions used in these calculations has been obtained by whole-rock normalization of plagioclase and hornblende REE abundances from western tonalite sample 1. REE analyses of the same minerals in the quartz gabbro would be more appropriate, but were not available. The shape of the REE distributions for clinopyroxene are similar to hornblende, and its influence on the REE can be adequately accounted for by the hornblende distribution. In this calculation, the whole-rock-normalized values for the hornblende from the tonalite have been uniformly adjusted downward to values yielding an optimal

FIGURE 47. The effect of subtraction of hornblende and plagioclase by crystal fractionation on the REE pattern of a quartz gabbro magma. The quartz gabbro REE pattern is that of sample 9. All of the calculated REE patterns are for 40% crystallization, but for variable proportions of hornblende and plagioclase. This degree of crystallization with subequal amounts of these minerals is required to generate a residual liquid with the major element chemistry of a tonalite such as sample 1, whose REE pattern is shown for comparison. The distribution coefficients used in this calculation are as follows:

	Plagioclase	Hornblende
Ce	.175	1.22
Nd	.070	1.93
Sm	.025	2.36
Eu	.700	1.07
Gd	.015	2.50
Dy	.008	2.50
Er	.007	2.50
Yb	.006	2.50


■ QUARTZ GABBRO
● TONALITE
.4-MELTS DERIVED BY 40% CRYSTALLIZATION OF AN ASSEMBLAGE WITH THE INDICATED HB/(HB+PLAG)
MELT / CHONDRITE
100
50
20
10
5
2
1
.1
.4
.6
1.0
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

match in REE abundances for subtraction of an assemblage containing subequal amounts of plagioclase and hornblende. Such lower values for hornblende are observed in more mafic rocks, and clinopyroxenes tend to have lower values as well.

Figure 47 indicates that the behavior of REE in a system differentiating by crystal fractionation of plagioclase and hornblende (and possibly clinopyroxene) is sensitive to the relative proportions of these phases. General agreement is found in REE abundance levels and fractionation in the middle to heavy REE when plagioclase and hornblende are subequal, but an excessively steep fractionation in the light REE is always present in the derived residual liquids when considerable amounts of hornblende (or clinopyroxene) is subtracted. This is quite significant and can not be overlooked. Due to the concave-down shape of their distribution patterns, hornblende or clinopyroxene-bearing assemblages subtracted in the proportions and quantities broadly suggested by the major element chemistry would always produce considerable changes in the relative fractionation of the light REE in the residual liquid. Furthermore, the whole-rock-normalized REE distribution of the hornblende in the tonalite has the least concave-down shape and the least steep slope in the light REE. The volcanic hornblende distributions and the whole-rock-normalized distribution for the hornblende from the granodiorite near Indio have consistently steeper slopes in the light REE (see Fig. 45). The REE distribution of the hornblende used in this calculation in both shape and its lowered absolute value is then the most favorable (although possibly the most unrealistic) for the success of this differentiation scheme.

It therefore appears that the remarkable similarity in the REE patterns of the quartz gabbros and tonalites from the western region in southern California is a major obstacle to models utilizing high-level crystal fractionation processes to explain the major chemical and petrographic differences among most of these closely related rocks.

There is one western tonalite from Baja California (sample 31, Fig. 32) whose REE pattern does differ substantially from the rest. This tonalite has lower REE contents, a steeper light REE fractionation, a positive Eu anomaly, and a reversal in slope in the heavy REE. All of these features are what would be expected if a hornblende-dominated assemblage was, indeed, subtracted in a crystal differentiation process. These REE features are consistent with the presence of abundant hornblende in the final assemblage of this rock. This tonalite, by its uncommon but plausible derivation from the other quartz gabbros and tonalites emphasizes the improbability that the latter were derived from differentiation of a single parental magma by high-level crystal fractionation.

This general discussion of high-level crystal fractionation can be extended to the siliceous leucogranodiorites of the western region. Simple calculations show that a Bonsall-type tonalite can yield about 50% of its mass as a melt of approximate granodiorite composition (e.g. Larsen, 1948, p. 164). The removed mineral assemblage would again be dominated by hornblende and plagioclase in subequal amounts. Since the mineral assemblage and the REE pattern of the starting composition (tonalite in this case) are the same as those used in the quartz gabbro-tonalite discussion, the effects on REE fractionations for this crystallization step can be approximated with the same diagram.

In contrast to the uniformity in the REE patterns of most of the quartz gabbros and tonalites, these leucogranodiorites are quite variable in REE abundances. Many of the granodiorites have steeper slopes in the light REE consistent with the forms of the calculated REE patterns. Some granodiorites also show a reversal in slope in the heavy REE, suggesting the removal of a mineral with a concave-down pattern observed in many hornblendes (particularly like those observed in volcanic rocks and the granodiorite near Indio). The large negative Eu anomalies present in most of the granodiorites, however, would require subtraction of very plagioclase-rich assemblages which would be difficult to reconcile with the required bulk compositional changes. However, steep, concave-up slopes in the light REE, a reversal in slope in the heavy REE, low REE contents and a very limited development of a negative Eu anomaly are observed in one leucogranodiorite (sample 25). These are the REE fractionation effects expected from hornblende and plagioclase crystallization in the proportion and amounts suggested by the major element chemistry.

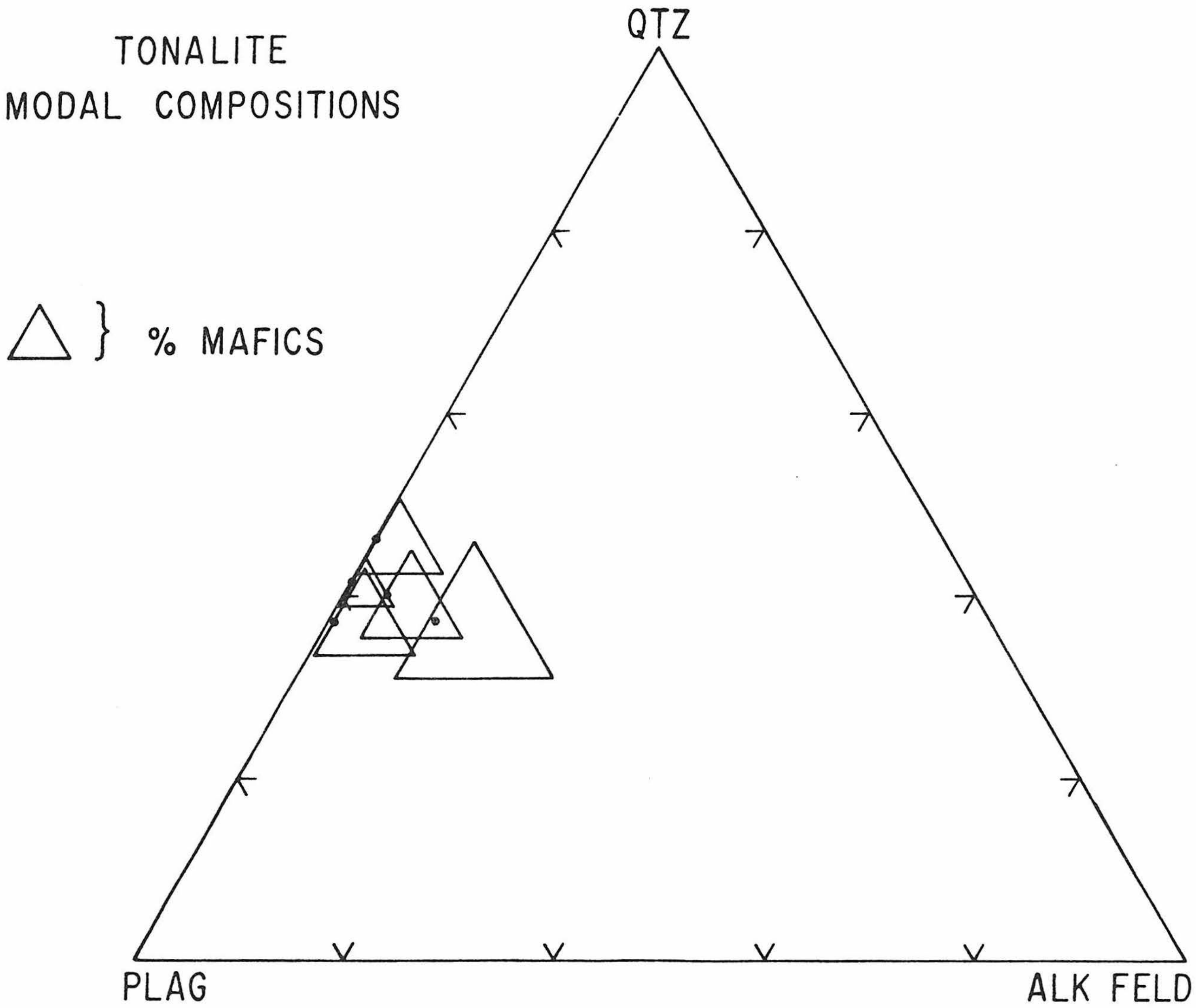
Whether the range in the REE characteristics of the various leucogranodiorites can be attributed to vagaries in the amounts or composition of the mineral assemblages inferred to be subtracted in the hypothetical crystal fractionation process is unresolved. If the minerals considered here are the only phases strongly influencing REE behavior, it appears difficult to derive all the granodiorites from a common parental magma such as might be represented by the tonalites. (A small amount of allanite is present in these rocks, but it is unlikely to have produced the observed qualitative characteristics.) This does not rule out the possibility that some degree of high-level differentiation has occurred, or that some of these leucogranodiorites are related to the tonalites by such a process.

There are other independent lines of petrologic and geologic evidence suggesting that the western leucogranodiorites are not related to the more mafic rocks by high-level crystal fractionation mechanisms. The histogram of Larsen indices for the western region in southern California (Fig. 40) is characterized by strong modalities in abundances of the lithologies. Both the quartz gabbros-tonalites and the leucogranodiorites are abundant rock types, but relatively few examples of intermediate compositions are present. In a crystal fractionation process, the amount of residual liquid must decrease continuously throughout the sequence. The very large proportion of leucogranodiorite in western San Diego and Riverside counties is incompatible with this. Additionally, field relationships indicate that both tonalites and leucogranodiorites form independent, largely uniform intrusive bodies (Larsen, 1948). Whatever differentiation processes may relate these rocks, most of them must have occurred at depth.

The question of high-level crystal differentiation has been examined in more detail within the San Telmo ring complex in Baja California which has been studied by Prof. Silver and his students. Multiple intrusive rings of tonalitic rocks were emplaced in a single event. Modal compositions of four tonalitic rocks and one low-K_2O granodiorite from the central and western parts of the complex are summarized in a ternary quartz-plagioclase-alkali feldspar diagram (Fig. 48). The height of the triangles is scaled to the percentage of mafic minerals; their generally small size reflects the felsic nature of these rocks. The modal abundance of plagioclase is limited to the range 45-55%. The limited range in the modal characteristics is an indication of the cogenetic nature of these rocks, and is accompanied by uniform plagioclase compositions. The plagioclase is zoned from interior sodic andesine or calcic oligoclase, to

FIGURE 48. Quartz-plagioclase-alkali feldspar ternary for some tonalites of the San Telmo pluton, Baja California. The heights of the triangles correspond to the percentage of mafic minerals, and the dots within the triangles to the position of the samples within the ternary if the three components are summed to 100%. These rocks display a very limited petrographic range.

SAN TELMO RING COMPLEX



a rind of sodic oligoclase.

The REE patterns of all these rocks are parallel (except for Eu), but range by a factor of 2.5 in concentrations (Fig. 34b). The rock with the highest REE contents has a very large negative Eu anomaly (55%). In the other rocks, the small Eu anomalies do not correlate with REE concentrations. Sr concentrations in these rocks also vary by a factor of 2, but not smoothly with REE content or Eu anomaly. The amount of crystal differentiation implied by these variations far exceeds the amount permitted by the limited modal variations and, in particular, the uniform plagioclase composition. The zoning present in plagioclase implies melt composition changed significantly during final crystallization, establishing the potential for large differentiation effects, but the uniformity in plagioclase among these various rocks demonstrates that appreciable differentiation did not occur. The fact that nearly all the Sr in these rocks is in a single major phase - plagioclase - rules out the possibility that the observed variations in Sr and REE could be due to small but varying amounts of late melt component in these rocks. The amount of dilution required to vary REE and Sr contents by a factor of two is excessive. The trace element variations described here must have been present in the magmas from which these rocks formed prior to emplacement and solidification. It is concluded that these variations were inherited from the source regions of these magmas, and were not obscured by emplacement or crystallization processes.

This discussion of the western region is concluded with the consideration of recently proposed hypotheses involving a different type of chemical differentiation. The high liquidus temperatures of tonalitic melts, above 1000°C even with excess H_2O (Lambert and Wyllie, 1970), has led some workers to suggest that tonalites never were complete melts. This stems in part from the belief that these rocks have their origin within the crust, where it is difficult to achieve such high temperatures, and in part from the interpretation of some petrographic features of some of these rocks. The essence of these hypotheses (e.g. Wyllie et al., 1976; White and Chappell, 1977) is that tonalites are mixtures of granite or granodioritic liquid with unmelted mafic and calcic components. Mafic xenoliths so commonly found in tonalites, and in some cases cores of the mafic and calcic mineral grains, are interpreted as transported relicts from the source regions. The chemical variation in the granitic suite is attributed to different degrees of unmixing of melt and this residual material (restite) from a mixture which formed during ultrametamorphism of the source material.

In this hypothesis, the Woodson Mountain granodiorite of this batholith would presumably be the melt component. The quartz gabbros and tonalites, particularly those with abundant mafic inclusions, would be restite-rich mixtures.

Several problems in the application of this hypothesis to the rocks of the western region of the Peninsular Ranges batholith become apparent when the REE, bulk chemical, and petrographic features of these rocks are considered. While most western quartz gabbros and tonalites have uniform REE patterns, there is no uniform REE pattern for the western leucogranodiorites (Fig. 28). The bulk composition of the tonalites suggests that they hypothetically contain subequal amounts of leucogranodiorite melt and restite, yet the variability in the leucogranodiorites is not seen to influence the remarkably consistent REE patterns of the tonalites. Further, the range in bulk chemistry among the quartz gabbros and tonalites implies variable amounts of melt and restite components in these rocks. The similarity in the REE patterns of the quartz gabbros and tonalites does not reflect the differences in the REE patterns between the leucogranodiorites and the quartz gabbros-tonalites (the leucogranodiorites show a general relative depletion of the heavy REE and increased fractionation in the light REE; see Fig. 28).

White and Chappell (1977) suggest that a restite component such as plagioclase may be recognized in many granitic rocks as complexly zoned grains or grains with calcic cores. If the tonalites of the Peninsular Ranges batholith contain 50% restite component as suggested by the bulk chemistry, this requires, hypothetically, that the inner

80% of the diameter of the average mineral grain ($0.8^3 = 0.5$) is restite component. The few calcic cores of plagioclase observed in tonalites could not account for this much material. Rather, most grains have preserved delicate growth zoning, often oscillatory. Hornblendes are often well formed prismatic grains, suggesting growth from a melt.

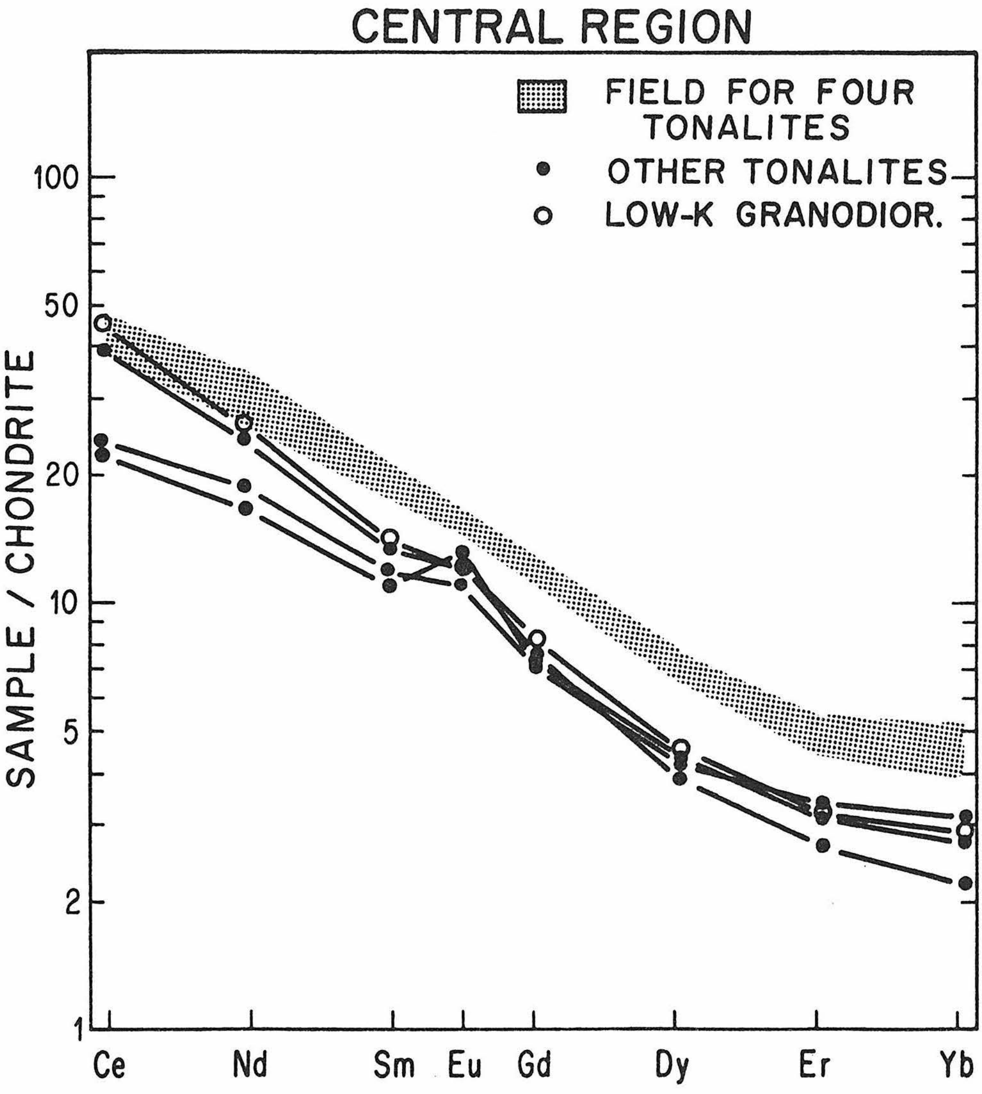
This general hypothesis may have merit, however, if tonalite is the liquid component and the restite component consists of minerals in proportions similar to those found in the tonalite. It may then be possible to relate the tonalites and the quartz gabbros by a melt-restite relationship, although evidence of restite components in the quartz gabbros should be documented. The formation of the more siliceous liquids such as the leucogranodiorites would, of course, require some other process.

Central Region

The restricted range in the diversity of lithologies represented in the central region does not permit as extensive treatment of the role of crystal fractionation as for the western region. The rocks of the central region are comprised almost entirely of tonalite and low-K_2O granodiorite. Most of the tonalites have a very consistent fractionated REE pattern with no Eu anomaly. A few of the tonalites display lower abundances of REE coupled with small to moderate positive Eu anomalies. The two low-K_2O granodiorites, both from the large La Posta pluton, also have lower REE contents except Ce, and a small positive Eu anomaly. These two granodiorites and one tonalite also have slightly steeper slopes in the light REE. These characteristics are summarized in Figure 49.

If a crystal fractionation hypothesis relating these rocks to a common magma is to explain these variations, either removal of hornblende and/or unusual accumulation of plagioclase is required to produce the rocks with low REE content. Crystallization and removal of a hornblende-rich assemblage from magmas having REE patterns like the tonalites without Eu anomalies could account for most of the REE features of the rocks with lower REE contents. It is inconsistent, however, that the most extreme examples of rocks with low REE contents and positive Eu anomalies are tonalites, not the presumably more differentiated granodiorites, and that some of these rocks are as mafic or more mafic than the other tonalites. Most of these rocks are not more plagioclase-rich than the other tonalites from this region. Excess plagioclase could not have been the cause of these variations as the amount required is far in excess of what is observed petrographically. Accumulation of early formed plagioclase from a much larger reservoir than appears to be represented by the plutons seems unlikely as most of these plutons are fairly uniform. The remarkable modal and chemical homogeneity of the San Jose pluton (Murray, 1978) would require the

FIGURE 49. Summary of the REE characteristics of the central region rocks. A field for four tonalites from both southern and Baja California which do not have Eu anomalies is shown. Other tonalites have positive Eu anomalies and generally lower REE contents. Two low-K_2O granodiorites, represented by a single analysis, have REE patterns resembling these latter tonalites.


CENTRAL REGION
FIELD FOR FOUR TONALITES
OTHER TONALITES
LOW-K GRANODIOR.
SAMPLE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

supposed plagioclase accumulation to take place prior to emplacement and to be thoroughly homogenized over the volume of granitic magma underlying 117 km^2, a difficult task for such a viscous material. The essentially identical REE patterns of the two low-K_2O granodiorites from the La Posta pluton, spaced 25 km apart, indicate the uniformity of this pluton. Uniform intrusive bodies, at least as evaluated from the bulk compositional and mineral relationships, appears to be a common occurrence in the Peninsular Ranges batholith (Larsen, 1948; Silver, per. comm.). However, more information on the variability of REE patterns of whole rocks within individual plutons and how these relate to petrographic features is desirable.

Eastern Region

The mineral study of the granodiorite presented earlier has direct application to REE behavior during crystallization and differentiation of rocks of the eastern region. There are two lines of evidence suggesting that these rocks cannot be related to a common parental magma by a crystal fractionation mechanism. As indicated previously, the early appearance of allanite and sphene with high REE content, the REE zoning patterns of these phases, and the unusually low whole-rock normalized values for some major minerals all indicate that REE were being depleted in the melt during the earlier stages of solidification of this granodiorite. The same mineralogy and textures of the other tonalites and granodiorites suggest similar characteristics. The granodiorites in southern California consistently have higher middle and heavy REE contents than the tonalites, a trend opposed to the early depletion.

The strong localization of light REE in allanite and middle to heavy REE in sphene indicates that the relative proportions of these phases can exert strong control on the shape of the REE pattern of the rock. If this suite of rocks is presumed to be related by high-level crystal fractionation, their parallel to subparallel patterns requires limited variation in the relative abundances and fractionations of these REE-rich accessories. This is not indicated by petrographic observations, nor by yields from heavy mineral separates (Table 7). Allanite/sphene ratios for the selected samples vary by a factor of five. Variation in the Sm/Dy ratio provided for comparison is not correlated. The variation in the abundance of sphene does not correlate with the concentration of Yb, a heavy REE, measured in the whole rock.

It does not appear possible to relate the relative and absolute abundances of REE-rich accessories to the REE characteristics of these eastern rocks. This suggests that high-level crystal fractionation processes have not determined or significantly altered the highly fractionated REE patterns of these rocks. Rather, these REE patterns appear to be primary features of the granitic magmas which are retained

TABLE 7

DATA ON SELECTED ACCESSORY MINERALS IN SOME ROCKS

Rock Type	Allanite/Sphene (wgt. ratio)	Sm/Dy (chondrite normalized) in rock	Sphene (g/kg rock)	Yb ppm in rock
Tonalite	.03	4.2	2.6	.76
Granodiorite	.15	2.8	2.5	1.12
Tonalite	.10	3.4	1.7	.63
Granodiorite	.08	2.7	1.6	2.04

in the whole rock through largely closed-system crystallization.

Gabbros

The gabbros of the batholith present some very intriguing problems. Many of the gabbros possess features which have been interpreted to be the result of local high-level differentiation. Perhaps the most important characteristic is the variability observed in the field. Often several different rock types are intimately and sometimes gradationally associated within a single gabbro body. Some of the gabbros have compositions which are very unlikely to have been melts (e.g. peridotite in Cuyamaca gabbro, Creasey, 1946; peridotite and anorthosite in the gabbro of Los Pinos Mountain, Walawender, 1976). There are other gabbros which form more uniform bodies and possess relatively simple igneous textures, suggestive of direct crystallization from gabbroic magmas.

While some workers (e.g. Nishimori, 1976) have proposed that these gabbros represent appropriate material to subtract from a presumably basaltic parental magma to produce the regular chemical trends of the main batholithic suite, there are several characteristics of the gabbros which are at odds with such an interpretation. The REE characteristics of these gabbros, with low REE contents compared to quartz-bearing rocks and either large positive Eu anomalies or strongly concave-down shapes, are inappropriate for the material whose fractional removal would have presumably moved the batholithic differentiation from quartz gabbro through tonalite. The uniformity of the REE patterns of these rocks strongly disfavors this possibility. In contrast, the shapes and even the variability of the REE patterns of the leucogranodiorites of the western region and the gabbros seem complementary. This potential correspondence encounters the difficulty in following the quartz-poor gabbros in the field by gradation into any of the quartz-bearing lithologies, particularly the siliceous leucogranodiorites (Silver, per. comm.). Additionally, the minerals of some of these gabbros are often too magnesian and too calcic to have crystallized from the quartz gabbros (Miller, 1937), the most mafic rocks of the continuous trend of batholithic differentiation.

The REE data collected on several of these gabbros, taken by themselves, are compatible with accumulation of the phases present in the rocks. The noritic gabbro from Pala (sample 8) and the olivine gabbro from Thomas Mt. (sample 16) both have the low REE contents and strong positive Eu anomalies expected of plagioclase-dominated accumulation. The strongly concave-down patterns and small positive Eu anomalies of two of the three hornblende gabbros (Fig. 38) are quite unusual in this study, and are qualitatively what would be expected from hornblende and plagioclase accumulation. None of these samples show the obvious textures and structures of cumulate rocks, with the possible exception of the Pala gabbro. Some appear to have been recrystallized and may have lost original textures.

The gabbros included in this study are not sufficiently representative of the range of gabbros observed in the batholith (e.g. Miller, 1937; Creasey, 1946; Walawender, 1976) to adequately characterize the entire range of variations in REE patterns that may exist. At this stage of observations, however, it does not appear likely that the limited abundance of gabbros observed and the diversity of REE trends are consistent with the simple model proposed by Larsen in which the batholithic lithologies were derived by fractional crystallization of a parental gabbroic magma at depth.

DISCUSSION OF MAGMA CONTAMINATION WITH UPPER CRUSTAL MATERIAL

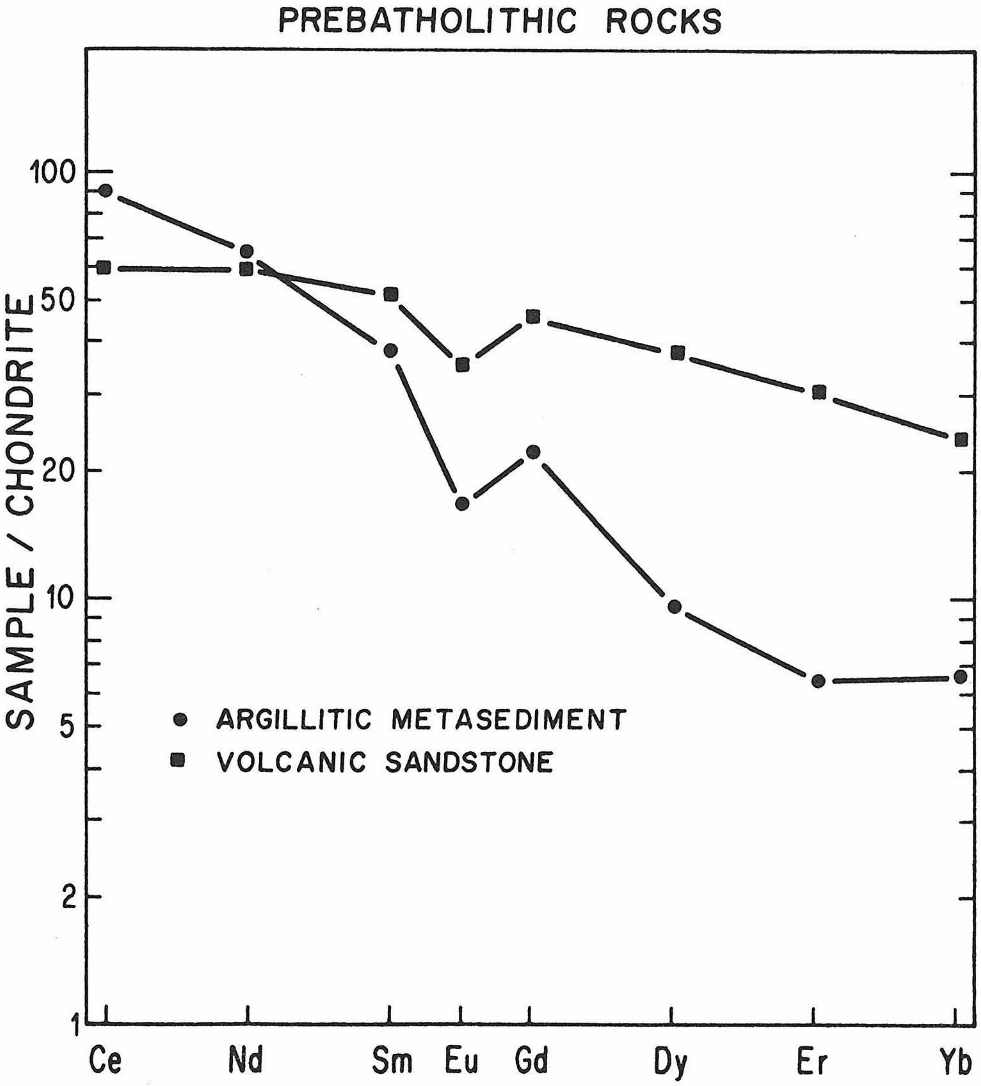
The batholith was emplaced in high-level crustal rocks, which in the northern and eastern parts of the batholith include abundant clastic sedimentary rocks containing older continental components. The possibility of magma contamination by these components must be addressed. Some workers have attributed the more radiogenic Sr observed in the cratonward portions of certain Cordilleran batholiths to crustal contamination processes (e.g. Armstrong et al., 1977). It might similarly be suggested that the more fractionated nature of the REE patterns of the central and eastern regions of the Peninsular Ranges batholith have been influenced by magma contamination with crustal materials, which are usually strongly REE fractionated.

We will consider first the systematic variations in REE and Sr isotopic composition in tonalites across the batholith. In the western regions, the REE patterns are rather flat and $^{87}Sr/^{86}Sr$ initial ratios low (0.703-0.705) while strongly fractionated REE patterns and high $^{87}Sr/^{86}Sr$ initial ratios (0.705-0.707) are characteristic of the eastern regions. The eastern rocks are therefore the rocks presumably most likely to have been affected by interaction with continental crust. The REE data on the tonalites from southern California have been summarized in Figure 30. From west to east, there is a reduction in the heavy REE by a factor of 3 to 5. If this feature is to be explained by contamination, massive amounts of diluting contaminant would be required. In the most favorable case of a contaminant with no heavy REE, the resultant rock would consist of up to 80% contaminant. Yet, these eastern tonalites have bulk compositions similar to the western tonalites. A contaminant process appears even more difficult when candidate upper crustal materials are considered. Shale composites from globally distributed stable platform deposits are often cited as representative of the upper continental crust. These composites possess a striking uniformity in REE character (Fig. 9). The light REE in shales have similar abundances and fractionations to the eastern tonalites, but the shales have much higher abundances of heavy REE, which are unfractionated at 15-20 times chondrites indicating they are inappropriate as a mixing component to form eastern tonalites from western tonalites.

The question of upper crustal contamination has been considered further by analyzing two prebatholithic metasedimentary rocks for REE. REE patterns for an argillitic metasediment from the French Valley Formation (described by Schwarcz, 1969) and a volcanic derived sandstone from the vicinity of the San Jose pluton in Baja California are presented in Figure 50. They are located within the western and central REE regions, respectively. It is not known how representative these patterns are of similar lithologies in the prebatholithic section.

The argillitic metasediment has a pattern much different from that of the shale composites (see Fig. 9). Its strongly fractionated pattern across the light to middle REE becomes flat in the heaviest REE. Strong light REE enrichment is expected upon assimilation of this material. The modest light REE contents and fractionations of most of the western and central region rocks which intrude this prebatholithic unit do not appear to reflect its light REE properties. The operation of an assimilative process to produce the characteristic light REE enrichment and fractionation observed in the western leucogranodiorites is precluded by $^{87}Sr/^{86}Sr$ ra-

Figure 50. REE patterns of two prebatholithic rocks. The analyses of these rocks were presented in Table 6. The argillitic metasediment occurs within the western region in southern California. The volcanic sandstone is located in the central region in Baja California.


PREBATHOLITHIC ROCKS
SAMPLE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb
● ARGILLITIC METASEDIMENT
■ VOLCANIC SANDSTONE

tios in excess of 0.710 (Early and Silver, per. comm.) in this rock at the time of plutonic emplacements. Eastern region rocks with light REE enriched patterns are spatially removed from the presently known distribution of this particular unit.

The distinct concave-down REE pattern of the volcanic sandstone is a feature not observed in any of the granitic rocks, although this rock was probably derived from the closely associated volcanic arc to the west (Silver, per. comm.). Its very high heavy REE contents are in contrast to the low values characterizing the nearby plutonic rocks of the central region. Its relatively flat pattern would not aid in the production of the REE patterns of the central and eastern region rocks.

These few data on the prebatholithic rocks are not comprehensive in representing wall rocks with which plutons of the batholith might have interacted, but they are important types. Simple assimilation of at least much of the prebatholithic metasedimentary rocks is unlikely to have contributed to the regional REE patterns and to the major lithologies of the batholith. The above arguments do not rule out the addition of crustal materials to the deeper source regions of the batholithic magmas.

It has been argued that contamination with upper crustal materials is not responsible for the systematic regional Sr isotopic variations (Early and Silver, 1973; Silver et al., 1975). At any one geographic position, the $^{87}Sr/^{86}Sr$ initial ratio is independent of rock type. Tonalites have the same values as adjacent quartz monzonites or leucogranodiorites; thus the petrologic diversity cannot be attributed to contamination with older radiogenic crustal components. Silver et al. (1975) have noted that the isotopic variations cannot be related to the distributions of prebatholithic lithologies. Some of the least radiogenic plutons are intruded into metasedimentary formations which had $^{87}Sr/^{86}Sr$ ratios greater than .710 at the time of emplacement of the batholith. Similar contrasts are observed in $\delta^{18}O$ values between the plutonics and the country rocks (Turi and Taylor, 1971; Taylor and Silver, 1978). These features, along with the similarity in bulk composition among the tonalites across the batholith, effectively rule out the possibility of producing the regional variations in Sr and O isotopic compositions by massive contamination at high levels.

The possibility of contamination with a small quantity of extremely radiogenic Sr and heavy O, perhaps by exchange, must be considered. Almost all of the rocks with high $^{87}Sr/^{86}Sr$ initial ratios have high Sr contents (greater than 400 pp, and up to 700 ppm). These rocks should not be as susceptible to exchange as the lower Sr rocks to the west. The rocks sampled in this study, with very few exceptions, do not show petrographic evidence of extensive hydrothermal alteration. Taylor and Forester (1971) have pointed out that exchange may occur only at very high temperatures in some hydrothermal systems, and not produce petrographically observable effects. However, the recent work of Taylor and Silver (1978) has demonstrated that the major regional ^{18}O variations are not due to high-level exchange. Additionally, the regularity in the regional variations and the uniformity at any one geotraphic position argue against an exchange origin for the Sr isotopic variations. The exchange process may operate locally. Given the great heterogeneity in upper crustal materials, it is difficult to visualize the operation of controlled processes capable of producing such consistent variations.

SOURCE REGION HETEROGENEITIES

OVERVIEW

The preceding arguments concerning high-level crystal fractionation and contamination processes strongly suggest that such processes have not been a significant factor in the production of the regional variations in REE characteristics, and probably not even most of the local variability. The same is also true of the Sr and O isotopic variations, with the possible exception of exchanges within external H_2O systems along the shallow west side of the batholith. The first-order variations must originate at deeper levels, most likely within the source regions of these rocks.

The profound transverse asymmetry in these variations relate to the structural elements of the plutonic arc, and reflect the dynamics of batholith formation at a convergent plate boundary. The magma generating process has sampled heterogeneous regions within the source terrains of the batholithic rocks. The term source region heterogeneities is used here in the broadest sense, and includes major element, trace element, isotopic, and mineral assemblage characteristics of the source materials.

In the following discussions, the extent to which these aspects of source region heterogeneities contribute to the observed variations is explored.

ISOTOPIC VARIATIONS

The regional variations in the isotopic composition of Sr and O must result from simple inheritance of these properties from the source regions as suggested by Early and Silver (1973), Silver et al. (1975), Silver and Early (1977), and Taylor and Silver (1978). Recent studies of diffusion in melts (e.g. Hofmann and Hart, 1978) strongly suggest that isotopic equilibration over distances of millimeters to centimeters may occur well within the time scales required for partial melting of source terrains on the regional scale of a magmatic arc. Disequilibrium partitioning of Sr isotopes and trace elements among mineral grains and melt during melting is considered highly unlikely in such an environment. In contrast, equilibration over distances of many meters to kilometers does not take place over many millions of years. Unless a plutonic magma was very well mixed by some physical process prior to emplacement, at least partial preservation of some isotopic differences among various materials within the source volume of the magma may be expected. These isotopic variations may occur, conceivably, within a magma of uniform bulk composition as this general property may be determined by phase equilibria considerations rather than bulk and trace compositional uniformity.

Variations in the Sr isotopic composition of the source regions within the mantle or crust-mantle system in the region of the batholith would be anticipated from its geologic setting. The batholith was developed on or near a continental margin, and a transition from oceanic to continental properties within the lithologic components of the potential sources might occur across it. This transition would be normal to the continental margin. The smooth regional regularity in the nature of this transition bears directly on the origin of the isotopic differences. They might be derived from regionally controlled, long-term differences in Rb/Sr ratio within the source regions, or remarkably regular physical mixing on a regional scale of two or more components which have not otherwise been closely related over a long period of geologic time.

As pointed out by Taylor and Silver (1978), high ^{18}O reservoirs have to date only been recognized in rocks which were once near the earth's surface. On this basis, the ^{18}O data appear to suggest the presence of crustal oxygen components in the sources of the eastern part of the batholith. The occurrence of rocks essentially identical in major and trace element chemistry but with contrasting $\delta^{18}O$ values across much of the batholith indicates simple contamination cannot provide a satisfactory explanation.

REE AND SR CONCENTRATION VARIATIONS

Inferences from Rb/Sr - Sr Isotopic Systematics

The conclusions drawn about the isotopic variations raises the question of whether the REE and Sr concentration variations may also be properties inherited from the source regions. This would imply for the western region a source with a flat REE pattern and low Sr content, and central and eastern sources with a distinctive fractionated REE pattern and higher Sr content. This further assumes either no or minimal fractionations, or parallel fractionations occurred during melting of the sources.

One means of evaluating the sources and timing of fractionations is through examination of Rb-Sr relationships which must be consistent with the isotopic properties of Sr. Early and Silver (1973) have observed that within a limited geographic area, initial $^{87}Sr/^{86}Sr$ ratios are uniform and independent of lithology although the various lithologies may display a large range of Rb/Sr values. It can be deduced that fractionations in Rb relative to Sr accompanied whatever differentiation process was responsible for the lithologic diversity within a region, and that this fractionation and differentiation is constrained to have taken place at or very near the time of melting and emplacement ($<10^7$ years).

Early and Silver (1973) and Silver and Early (1977) have shown that the Rb contents of tonalites have a more or less constant range across the batholith, but Sr is much more abundant in the central and eastern parts of the batholith. The Rb/Sr ratios for the central and eastern tonalites, correspondingly, are lower than those for the western region. The eastern tonalites, however, have higher initial $^{87}Sr/^{86}Sr$ ratios. Taking the batholith as a whole, the Rb/Sr ratios of these rocks do not reflect long-term differences of Rb/Sr ratio in the sources implied by the regional Sr isotopic variations (Early and Silver, 1973; Silver and Early, 1977). These differences in Rb/Sr ratio are quite large, and if they were present in the source regions, the regional regularities in Sr isotopic composition would have been obscured in a short period of time.

It is concluded that differential fractionation of Rb and Sr must have developed during or very close to the time of production of the magmas from their sources. These arguments suggest that the variations in Rb/Sr ratio, which are largely variations in Sr content, could be related to fractionations by mineral assemblages, not large differences in Rb/Sr of source materials. The differential fractionation presumably results largely from changing mineral assemblages and possibly mineral proportions in the residual assemblage coexisting with melt. Possibly the residual mineralogy of the western tonalites retained appreciable Sr, while the residuum of the central and eastern tonalites retained much less. The possibility of some of the regional variation in Sr contents reflects inherent differences in the Sr contents of the source materials cannot be ruled out, however.

These inferences are further developed below with regard to the REE variations. These variations are addressed in the context of the two regionally observed transitions in REE character: the western to central region and the central to eastern region transitions.

Western to Central Region Transition

The transition from the western to the central REE-defined regions of the batholith is marked by changes in the following general properties: a depletion and fractionation in middle to heavy REE; a disappearance of negative Eu anomalies in tonalites; an increase in Sr contents of tonalites; a restriction in the diversity of lithologies. These differences between the regions are summarized in Table 8.

The fact that these four significant properties appear to change together strongly suggests they have a related origin. In contrast, the isotopic and age variations do not reflect the rather sharply defined west to central transitions in several of the above properties. This observation, coupled with arguments against the direct inheritance of the Rb/Sr and Sr concentration variations from the source regions, argues that the discontinuities in these features are not reflections of what is compositionally present in the sources, but are probably dominantly process related.

A single process which appears to satisfy the above observations is a phase transition in the residual mineralogy of the source materials from a plagioclase-bearing assemblage to a garnet-bearing, plagioclase-poor assemblage.

The low Sr contents and high Rb/Sr ratios of the western rocks may be interpreted as the result of retention of a Sr-bearing phase (plagioclase) in the residual mineralogy of the

TABLE 8

SUMMARY OF DIFFERENCES BETWEEN THE WESTERN AND CENTRAL REGIONS

WESTERN REGION	CENTRAL REGION
Unfractionated heavy REE pattern	Fractionated and depleted heavy REE pattern
Negative Eu anomalies in tonalites	No or positive Eu anomalies in tonalites
Low Sr contents in tonalites, with comparatively high Rb/Sr ratio	High Sr contents in tonalites, with comparatively low Rb/Sr ratio
Wide diversity of lithologies; distinct peaks in abundances of tonalites and SiO_2-rich granodiorites; gabbros common	Restricted range of lithologies; single peak in abundance of tonalite-low K granodiorite; gabbros rare

source of these rocks. The feldspars are the only major rock-forming mineral capable of storing Sr. As a corollary effect, the higher Sr contents of the central rocks may be a reflection of the elimination or large reduction in abundance of the feldspars.

The negative Eu anomalies of the western tonalites also suggests the involvement of feldspar in the evolution of these rocks. The absence of negative Eu anomalies in the tonalites of the central region is consistent with a much less important role for feldspar in the derivation of these rocks. Although the anomalous behavior of Eu compared to the rest of the REE may be eliminated under sufficiently oxidizing conditions, the lack of consistent appreciable differences in $Fe^{3+}/(Fe^{3+} + Fe^{2+})$ between the western and central rocks (Silver, per. comm.) does not support this mechanism to explain the differences in Eu anomalies. However, this does not disprove variations in pO_2 in the sources.

The assemblage feldspar-amphibole-pyroxene-biotite-quartz cannot be expected to produce much relative fractionation among the heavy REE. The fractionated and depleted heavy REE patterns of the central region rocks requires the involvement of a phase or phases that would preferentially fractionate as well as retain in abundance the heavy REE. Objections to the suitability of accessory phases such as allanite and sphene to accomplish this rest mostly with the difficulty in producing large and consistent fractionations with small and presumably highly variable amounts of these phases. Garnet is the only major rock-forming mineral having the required properties. Garnet is stable at high pressures and is important volumetrically across broad compositional fields of most of the batholithic rocks and their plausible source materials (Green and Ringwood, 1968; Green, 1972; Wyllie et al., 1976). Its participation implies formation of the batholithic magmas of the central region under high pressure conditions. The appearance of garnet is complementary to the previous inference of the disappearance of plagioclase as a major phase, which is expected at higher pressures. Pressure and temperature estimates for the transition from plagioclase to garnet-bearing assemblages and their implications for the geometry of the zone of melting will be discussed in the following sections.

The transition from a multimodal to a unimodal frequency distribution of lithologies lends support of a different kind to the concept of a major change in the phase assemblages. The difficulties in producing the multimodal distribution of the western region by fractional crystallization or contamination processes have already been discussed. One improbable alternative is to derive the magmas of contrasting composition by melting sources in which the observed differences have already existed. There is a great contrast in Rb/Sr ratios between the tonalites and granodiorites, yet their $^{87}Sr/^{86}Sr$ initial ratios show no compositional dependence. This relationship constrains the time of formation of the proposed differences in source composition to no less than near contemporaneity with the formation of the melts (Silver and Early, 1977).

The diversity of the western rocks is attributed to phase equilibria controls. No other mechanism is envisioned which can satisfy the other observations and still produce the smooth and consistent covariations of chemical components displayed in the variation diagrams of Larsen (1948). Partial melting at invariant points or points of reduced degrees of freedom can provide several recurring compositions with a paucity of intermediate types.

The appearance of a more limited compositional range in the central region can be explained by a major readjustment in the stable phase assemblages. The loss of plagioclase and the incoming of garnet would require a major perturbation in the phase relations, probably including the number and location of pseudo-invariant points. However, an important constraint is that the contrasting assemblages must be capable of producing tonalites of similar bulk composition.

Central to Eastern Region Transition

This transition is defined by an increase in the abundances and fractionation of the light REE. These characteristics exist along the length of the peninsula from the San Jacinto Mountains to San Filipe. A low sample density makes correlations to other parameters difficult to establish, but initial $^{87}Sr/^{86}Sr$ ratios of the rocks of this region are always high, greater than about 0.7050. An increased proportion of granodioritic lithologies is commonly observed in at least parts of the eastern region, but this does not appear to be a sharp transition nor correlate to the position of the REE transition. The step in ^{18}O variations across the batholith does not coincide with the REE transition along the length of the peninsula.

Although there are so few other directly comparable discontinuities recognized across the central to eastern boundary and this transition is not yet as well defined as the western-central transition, it is not necessarily less important or significant. The central-eastern transition appears to show correlation with isotopic and some compositional variations suggesting more of a dependence on longer term, more subtle differences in compositional characteristics and history of the source material.

The general increase in the light REE content of the eastern rocks, by a factor of 2 to 3 for Ce compared to the rocks of the central region, could be produced by either a much smaller degree of partial melting in a constant source, the breakdown of a hitherto stable light REE-rich phase in the eastern source, or a higher light REE content in the eastern source.

A smaller degree of partial melting for the eastern rocks is not supported on several grounds. Tonalites and closely related low-K_2O granodiorites display the higher light REE contents. The bulk composition and general Rb-Sr equivalence of these rock types to the central region rocks implies roughly the same degree of partial melting. Additionally, this process requires residual phases in the source region very low in light REE content in order to permit enrichment of these elements in the melt. A residual assemblage containing a very small proportion of the total reservoir of REE cannot produce significant relative fractionations among the strongly excluded elements. The steep linear slopes of the light REE in most of the eastern rocks compared to the concave-down curvature of the same REE in the central region rocks cannot be achieved except by the most extreme differentiation processes.

The possibility of breakdown of a light REE-rich phase in eastern sources requires an accessory phase because there are no major rock-forming minerals which preferentially incorporate the light REE. Allanite appears to be a possible mineral; apatite and sphene discriminate against light REE compared to middle and heavy REE. Allanite is a conspicuously uncommon accessory mineral in metamorphic rocks of granulite and higher grade facies and is implausible (Silver, per. comm). In addition, control of the uniformity of the light REE abundances and fractionations of the tonalites within both the western and central regions by a single dominating accessory phase is very difficult to model.

The most attractive explanation of the high light REE contents of the eastern region is that the source of the eastern rocks had intrinsically higher light REE contents than the sources to the west.

The granodiorites of the northeastern region of the batholith in southern California have negative Eu anomalies, in contrast to the absence of Eu anomalies in the associated tonalites. Rb-Sr domains in the northeastern region have negatively-sloping arrays on a Rb versus Sr diagram (Silver and Early, 1977). These features all suggest some greater degree of involvement of plagioclase in the evolution of the granodioritic rocks.

In contrast, the few analyzed granodiorites in the eastern region of Baja California do not have negative Eu anomalies, although one of the tonalites does. Slopes of Rb versus Sr covariation for domains in Baja California tend to be vertical (Silver and Early, 1977), not sloped as to the north. These features suggest a more limited role for plagioclase in rocks of the eastern region in Baja California.

FURTHER PETROLOGIC CONSIDERATIONS

WESTERN REGION

Partial melting at invariant points to produce contrasting compositions or an apparently unique composition has been advanced by Yoder (1973) to explain contemporaneous basaltic and rhyolitic magmas. Presnall et al. (1973)

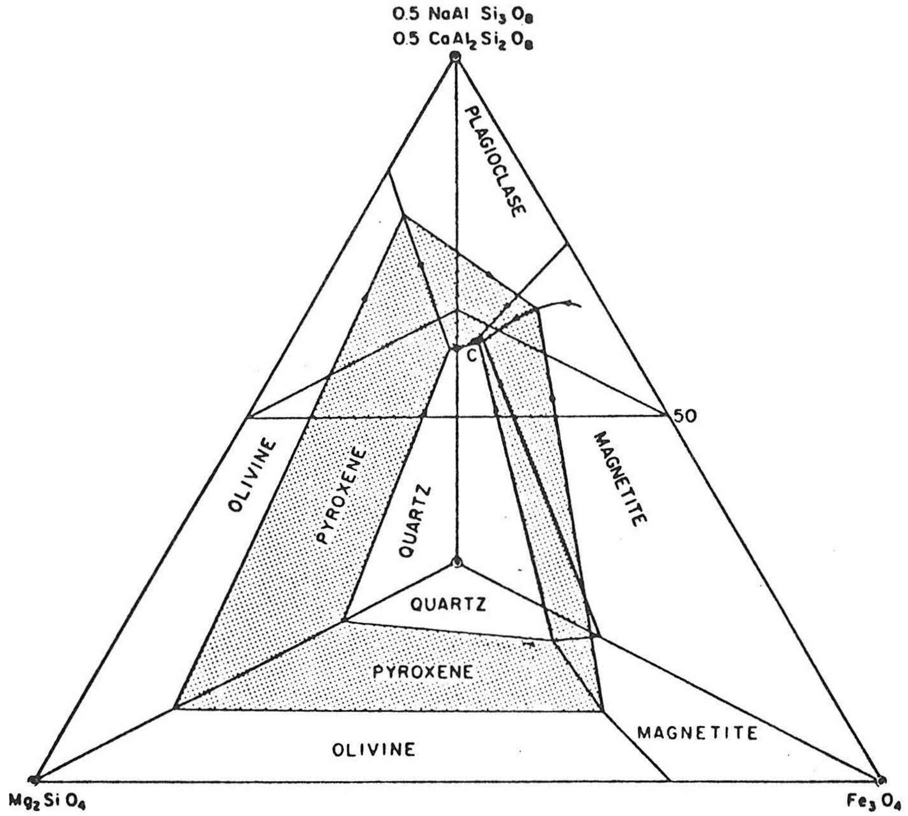
appealed to a phase-change cusp on the solidus of peridotite, a feature related to a point of reduced variance, to explain the chemical uniformity of tholeiitic basalts from mid-oceanic ridges. The question is whether pseudo-invariant points of appropriate composition exist in the complex petrological systems of the source regions of the batholith rocks. These systems will probably never be known in precise detail. Experimental studies on natural compositions have indicated, however, that liquids approaching tonalites and granodiorites can be produced from hydrous basaltic materials by partial melting (Green and Ringwood, 1968; Green, 1972; Holloway and Burnham, 1972; Eggler and Burnham, 1973; Helz, 1973, 1976).

The batholith source must be capable of producing large quantities of melts as mafic as the quartz gabbros and tonalites. In the unlikely event of 100% melting of a source, the source must be at least as mafic as these compositions. For smaller degrees of melting, it must be more mafic. On the other hand, the silica-oversaturated quartz gabbros and tonalites with 5-25% quartz and Mg/(Mg+Fe) ratios of 0.5 to 0.4 cannot be derived directly from peridotite (e.g. Green, 1973). Thus, if the quartz gabbros and tonalites are direct partial melts from their sources, the plausible composition of that source is approximately basaltic.

Perhaps the most realistic and sophisticated interpretation of phase relationships in appropriate synthetic systems has been presented by Eggler (1974). Eggler's study delimited the position of some pseudo-invariant points for the assemblage orthopyroxene-quartz-magnetite-plagioclase-melt-vapor within a portion of the system $CaAl_2Si_2O_8$-$NaAlSi_3O_8$-SiO_2-MgO-Fe-O_2-H_2O-CO_2. Several important parameters had to be held constant (such as An content of normative plagioclase, pH_2O, and pO_2) in order to obtain experimental determinations of pseudo-invariant points. Although orthopyroxene appeared rather than clinopyroxene, amphibole, and biotite, the basic positions of the pseudo-invariant points are likely to prevail.

Figure 51 is an example of the relationships presented by Eggler (his Fig. 5). Point C represents the composition of an H_2O-saturated melt containing normative An_{50} plagioclase (by definition) in equilibrium with quartz, plagioclase, orthopyroxene, magnetite, and vapor at 1 kb pressure. The composition of point C, along with the compositions of other pseudo-invariant points within this system under various conditions, are also given in Figure 51 (Eggler's Table 3). Most noteworthy is the very siliceous composition of an invariant point such as point C, and the importance of H_2O content of the melt as indicated in points D, E, and to a certain extent C. The precise applicability of Eggler's study to the rocks under investigation here is uncertain. Certainly, the pressure of most of these experiments is less than that of the source region of the batholithic rocks and the assumptions about plagioclase composition need to be more thoroughly

Figure 51. H_2O-saturated phase relations in a join of the system $CaAl_2Si_2O_8$-$NaAlSi_3O_8$-Mgo-SiO_2-FeO-O-H_2O at 1 kb total pressure at the NNO buffer, from Eggler (1974). Only relations about the point C have been determined. Compositions of some liquids at pseudo-invariant points in this system, including point C, are presented in the ajoining table.



Compositions of various liquids at pseudo-invariant points in the system $NaAlSi_3O_8$-$CaAl_2Si_2O_8$-SiO_2-MgO-FeO-Fe_2O_3-H_2O-CO_2

Point	Assem-blage*	Composition						Total pressure bars	Tempera-ture °C	Log fO_2 bars	Weight % H_2O in melt
		SiO_2	Al_2O_3	Fe as FeO	MgO	CaO	Na_2O				
A	I	66.7	14.4	7.3	3.4	5.1	3.0	1	1145	− 7.5	0.0
B	I	64.9	13.9	9.8	3.5	5.0	2.9	1	1135	− 8.5	0.0
F	I	55.5	16.2	13.0	6.1	9.1	—	1	1155	− 7.0	0.0
G	I	51.7	14.2	19.5	6.1	8.1	—	1	1133	− 9.0	0.0
C	I	72.2	13.0	6.2	0.6	4.7	2.7	1000	925	−11.8	4.3
D	I	69.8	14.4	7.0	0.7	5.1	3.0	5400	1005	−10.5	4.1
E	I	64.3	14.7	11.3	1.4	5.2	3.1	5400	1030	−10.0	2.0
K	II	53.5	15.1	17.7	5.2	5.4	3.1	1	1157	− 8.2	0.0
L	II	53.4	15.6	14.6	7.6	5.6	3.3	1	1181	− 7.0	0.0
d	III	61.9	19.3	—	7.6	7.0	4.1	1	1228	—	0.0
b	IV	71.7	15.3	—	4.3	5.5	3.2	1	1187	—	0.0
e	V	68.7	13.9	9.5	—	5.0	2.9	1	1185	− 7.8	0.0

*Assemblage I = opx-mt-trid (qtz)-plag-liq-(vapor)
II = opx-oliv-mt-plag-liq
III = opx-oliv-plag
IV = opx-trid-plag
V = plag-mt-trid

evaluated. However, the simultaneous formation of liquids corresponding to leucocratic granodiorites and tonalites of the western region of the batholith by melting at such pseudo-invariant points appears conceivable.

The compositions of the leucocratic granodiorites and particularly the tonalites and quartz gabbros of the western region of the batholith are not tightly fixed, but do show some range. This is not viewed as inconsistent with a mechanism of the above type. The compositions of the pseudo-invariant points only become invariant when all of the assumptions employed to reduce the multivariancy of the assemblage are met. Clearly they are not in the real case. Compositional positions of pseudo-invariant points would vary depending on the actual composition of solid solution minerals, pH_2O, and other variables. These variations may be limited enough, however, to yield melt compositions of a restricted range.

The granodiorites of the Woodson Mountain type may be produced by fairly small degrees of partial melting, their composition being largely controlled by the hypothetical pseudo-invariant point. Experimental studies of melting relations in natural basalts (Holloway and Burnham, 1972; Helz, 1973, 1976) at 5 to 8 kb, H_2O-saturated and undersaturated conditions, have produced melt compositions similar in many respects to the granodiorites for about 20% melting. Selected results are presented in Table 9. The high SiO_2

TABLE 9

SELECTED EXPERIMENTAL AND ROCK COMPOSITIONS

LEUCOGRANODIORITIC COMPOSITIONS

	875°C * 4.9 kb $pH_2O<P_t$	825°C ** 5 kb $pH_2O=P_t$	Average Woodson Mtn. Granodiorite Larsen (1948)
SiO_2	72.0	72.1	73.4
TiO_2	0.5	0.2	0.2
Al_2O_3	15.0	17.6	14.1
FeO	2.8	1.50	2.5
MgO	0.7	0.20	0.4
CaO	4.1	4.63	2.1
Na_2O	3.4	2.5	3.4
K_2O	1.2	1.2	3.5
Calculated melt (in %)	18	18	

TONALITIC COMPOSITIONS

				TONALITES Larsen (1948)	
	970°C ** 5 kb $pH_2O=P_t$	999°C * 5.2 kb $pH_2O<P_t$	995°C * 7.8 kb $pH_2O<P_t$	Average Bonsall	Bonsall of San Luis Rey
SiO_2	64.3	63.8	63.0	62.2	64.9
TiO_2	0.71	1.2	1.0	0.7	0.55
Al_2O_3	19.4	18.6	19.9	16.2	16.12
FeO	3.3	4.3	3.0	6.0	4.8
MgO	0.93	0.7	0.2	2.7	2.70
CaO	7.14	6.9	6.4	5.7	5.26
Na_2O	2.8	2.8	4.7	3.4	3.38
K_2O	0.92	1.0	1.0	1.6	1.64
Calculated melt (in %)	34	30	29		

* Holloway and Burnham (1972)

** Helz (1976)

and low MgO and FeO contents of the experimental glasses formed at 800°C to 900°C are in good agreement with the granodiorite. The low K_2O contents of the experimental melts are a problem, but probably reflect the K_2O contents of the starting material. These liquids coexist with hornblende, plagioclase, Fe-Ti oxides, and one liquid is in equilibrium with clinopyroxene. The low mafic contents of these melts despite equilibration with mafic phases is an indication of the very low solubility of these minerals in such melts, and supports deductions concerning the structure of siliceous melts presented earlier.

With more extensive partial melting, magmas with tonalitic to quartz gabbroic compositions could be derived from equivalent source materials. The composition of melts experimentally derived from basalts and coexisting with amphibole, clinopyroxene, and sometimes olivine, are also presented in Table 9. Temperatures indicated are about 1000°C, with about 30% melt present. These compositions are noticeably more aluminous and very poor in MgO compared to typical tonalites. These compositions are significantly removed from the melts produced at lower temperatures, perhaps due to the disappearance of a phase or phases in the residual assemblage. Possibly such compositions would be determined by a pseudo-univariant line or a different pseudo-invariant point. For example, local variations in pH_2O within an otherwise compositionally uniform source could exert control on the position of a pseudo-invariant point, as suggested by Eggler's study.

In addition to the discrepancies in chemistry noted above, another serious problem exists in reconciling these experimental data with the observations on REE and Sr behavior in the batholith. The melts corresponding to tonalites in these experimental studies do not coexist with plagioclase. Indeed, even the granodioritic melts listed previously coexist with only a small amount of plagioclase (<10%). In both cases, this is far too little to produce the fractionations which might be attributed to them. It is not apparent whether this is a major shortcoming of this hypothesis or a reflection of inappropriate selection of experimental conditions. For example, pH_2O in both these experimental studies ($pH_2O = 0.6\ P_t$: Holloway and Burnham, 1972; $pH_2O = P_t$: Helz, 1976) is probably much higher than can be expected for tonalitic magma. A high pH_2O will suppress the thermal stability of plagioclase relative to ferromagnesian minerals (e.g. Yoder and Tilley, 1962), a feature which is supported by a comparison of the observations of Holloway and Burnham (1972) and Helz (1973, 1976). Whether this trend, when carried down to reasonable pH_2O levels, would result in a more favorable residual assemblage and melt composition is speculative. The enhanced breakdown of amphibole by incongruent melting to pyroxene and melt might release more mafic components into the melt, bringing this aspect of the composition of experimental glasses more in line with natural systems. At the same time, increased residual plagioclase might lower the excess alumina and CaO of these melts. An alternative mechanism for increasing the mafic and calcic components of magmas permits them to carry some residual material from their source regions, as suggested by the restite hypothesis (White and Chappell, 1977).

The extent to which each of these various processes may have participated in the formation of the batholithic magmas is not fully resolved. However, derivation of magmas approaching the observed compositions of the batholithic rocks of the western region directly from gabbroic materials at pressures up to 10 kb or so appears plausible in light of experimental studies.

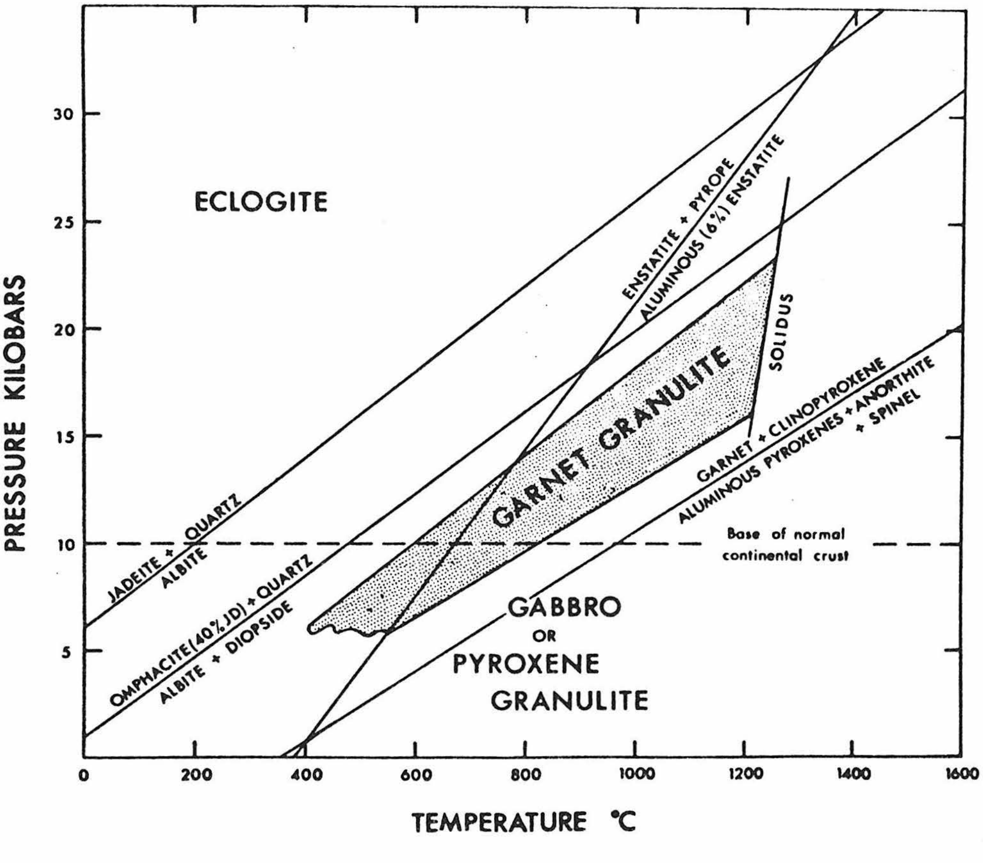
CENTRAL AND EASTERN REGIONS

Experimental data bearing on the composition of melts in equilibrium with garnet-bearing assemblages are much less available. The composition of possible invariant points are unknown. The results of several studies have indicated that partial melting at high pressure of material basaltic in composition can produce melts of broadly andesitic character under both anhydrous and hydrous conditions (Green and Ringwood, 1968; Green, 1972; Green and Ringwood, 1972). Summarizing these works, the most important features of a garnet-bearing assemblage in either a crystal fractionation or a partial melting process are: a) the absence of a strong Fe enrichment in derivative liquids because of the similarity of Mg/(Mg+Fe) ratio in garnet and melt, and b) the subsiliceous character of the garnet, which is very efficient in increasing the SiO_2 contents of derivative liquids. Determinations of actual compositions of melts are very limited. Discussion has been provided by the above authors, Gill (1974) and, dissenting, Stern (1974).

An important consideration which must be addressed is the location in P-T-a_{H_2O} space of the transition from garnet-free to garnet-bearing assemblages in possible source regions of the batholith. Arguments favoring a basaltic source region composition for the batholith have already been discussed. Most experimental work on the gabbro to garnet granulite or to eclogite transition has been for the anhydrous, subsolidus case. This transition under hydrous and supersolidus conditions is poorly known; available information is discussed below.

The garnet granulite assemblage - marked by the coexistence of plagioclase and garnet - is intermediate between lower pressure gabbroic (plagioclase-pyroxene) and higher pressure eclogitic (garnet-omphacite) assemblages under anhydrous, subsolidous conditions (Fig. 52). Both the first appearance of garnet in upper amphibolite facies and the final disappearance of plagioclase are sliding reactions which progress across the width of the stability field. Both transitions have positive P-T slopes. For a range of

Figure 52. Extrapolated anhydrous stability fields of pyroxene granulite-gabbro, garnet granulite, and eclogite for a quartz tholeiite composition, from Ringwood (1975).


ECLOGITE
ENSTATITE + PYROPE
ALUMINOUS (6%) ENSTATITE
SOLIDUS
GARNET GRANULITE
GARNET + CLINOPYROXENE
ALUMINOUS PYROXENES + ANORTHITE + SPINEL
JADEITE + QUARTZ
ALBITE
OMPHACITE(40% JD) + QUARTZ
ALBITE + DIOPSIDE
Base of normal continental crust
GABBRO OR PYROXENE GRANULITE
PRESSURE KILOBARS
5
10
15
20
25
30
TEMPERATURE °C
0
200
400
600
800
1000
1200
1400
1600

basalt compositions, the first appearance of garnet occurs at 10-15 kb at 1100°C (summarized in Ringwood, 1975). The pressure range over which plagioclase and garnet coexist is compositionally dependent, being wider for Na_2O-rich compositions. Anhydrous garnet granulite corresponding to quartz tholeiite starts to melt at approximately 1200°C and 15 kb, its lowest pressure on the solidus (Fig. 52). However, more than minor amounts of garnet are likely to be present with melt only at pressures substantially above 15 kb.

In the hydrous case, the phase relationships are more poorly understood due to the complex interplay between the stability fields of amphibole and garnet. Under subsolidus H_2O-saturated conditions, amphibolite is stable relative to eclogite for P_t less than 10 kb (Yoder and Tilley, 1962). At pressures of 20 to 25 kb, largely independent of T, amphibolite transforms to a garnet-bearing assemblage (Essene et al., 1970; Lambert and Wyllie, 1970). Above the solidus, relations are less well defined. Essene et al. (1970) suggested garnet amphibolite transforms to eclogite within the pressure range of 15 to 20 kb (limits of reversal in experiments) at temperatures in excess of 800°C. At temperatures the order of 1000°C, amphibole reaches its thermal stability limits. The work of Allen et al. (1972) indicates the liquidus nearly coincides with the breakdown of amphibole for a range of basalts. Again, garnet only appears above 15 kb. Only qualitative estimates of the proportions of phases in some assemblages and no estimates of the amount of melt are given in these studies.

H_2O-undersaturated conditions are assumed to have prevailed during the generation of the batholithic magmas. A rise in solidus temperatures and in the P-T stability field of plagioclase is expected, with some adjustment in the stability of amphibole. No experimental data exist on the high pressure phase assemblages of basalts in their melting range under H_2O-undersaturated conditions except very near their liquidii, where garnet appears for some H_2O-rich cases (10-25 wgt.% H_2O) (Nicholls and Ringwood, 1973). Basaltic melts might be expected to crystallize garnet when cooled further below their liquidii.

The problem of source assemblages can also be approached by examining the phases that crystallize from tonalitic to granodioritic melts under high pressure H_2O-undersaturated conditions. Studies on andesitic (Green, 1972; Stern et al., 1975) and rhyodacitic (Green and Ringwood, 1972) compositions (Table 10) similar to some tonalites and granodiorites from the eastern and central regions of the batholoth provide some useful information.

The phase relations determined in the vicinity of the liquidus of andesite + 2% H_2O (Fig. 53) indicate clinopyroxene and garnet appear close to the liquidus for $P_t \geq 15$ kb. Plagioclase begins to crystallize at lower temperatures and pressures, while amphibole does not appear until tempera-

Table 10

COMPOSITIONS USED IN EXPERIMENTAL STUDIES

	Andesite	Tonalite	Rhyodacite
SiO_2	60.24	59.9	68.33
TiO_2	.69	.80	.38
Al_2O_3	16.98	18.5	15.40
Fe_2O_3	.88	2.2	.59
FeO	5.35	3.7	3.53
MgO	3.14	2.5	1.74
CaO	7.22	6.0	3.30
Na_2O	3.91	3.8	2.79
K_2O	1.26	2.2	3.65
P_2O_5	.20	.30	.22
	Green (1972)	Stern et al. (1975)	Green and Ringwood (1972)

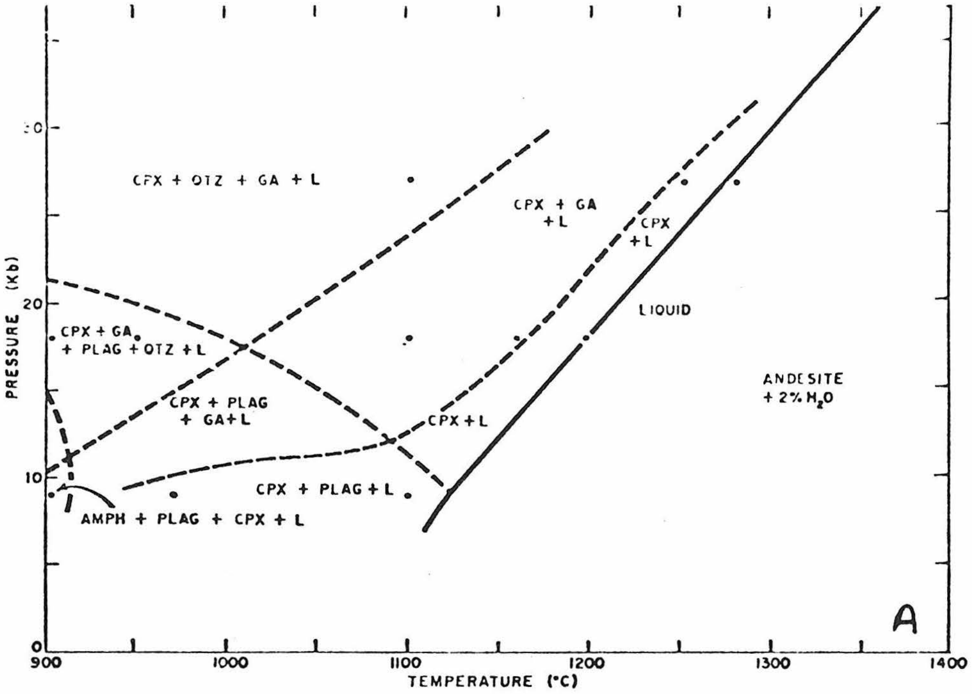
Figure 53. Pressure-temperature diagrams for the crystallization of andesite under hydrous conditions:

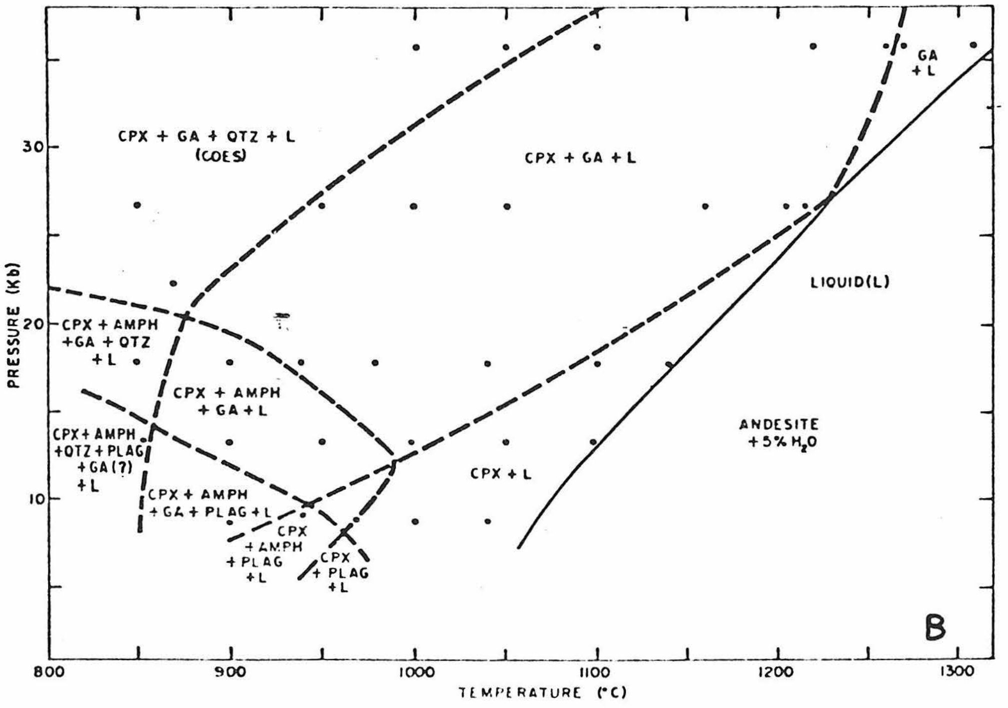
a) Andesite + 2% H_2O

b) Andesite + 5% H_2O

Note that with an increase in the amount of H_2O, amphibole stability is increased and plagioclase stability is decreased.

These diagrams are from Green (1972).


CPX + QTZ + GA + L
CPX + GA + L
CPX + L
LIQUID
CPX + GA + PLAG + QTZ + L
CPX + PLAG + GA + L
CPX + L
CPX + PLAG + L
AMPH + PLAG + CPX + L
ANDESITE + 2% H_2O
A
PRESSURE (Kb)
0
10
20
30
900
1000
1100
1200
1300
1400
TEMPERATURE (°C)


GA + L
CPX + GA + QTZ + L (COES)
CPX + GA + L
LIQUID(L)
CPX + AMPH + GA + QTZ + L
CPX + AMPH + GA + L
CPX + AMPH + QTZ + PLAG + GA (?) + L
CPX + AMPH + GA + PLAG + L
CPX + AMPH + PLAG + L
CPX + PLAG + L
CPX + L
ANDESITE + 5% H_2O
B
PRESSURE (Kb)
10
20
30
800
900
1000
1100
1200
1300
TEMPERATURE (°C)

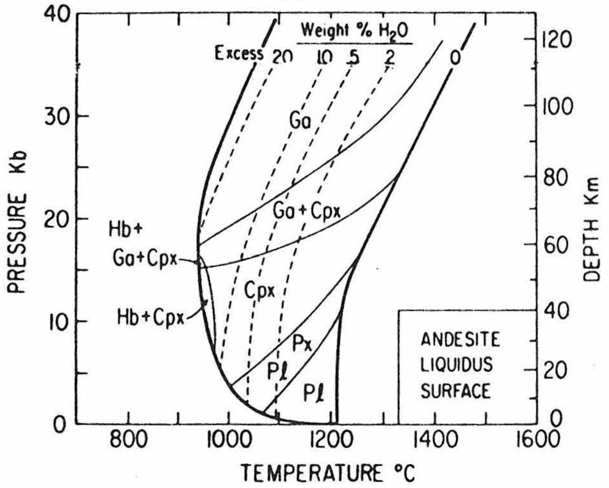
tures fall to 900°C. For the same andesite with 5% H_2O added, amphibole stability expands and plagioclase contracts. Clinopyroxene and garnet exist in the same position relative to the liquidus, but a plagioclase-free clinopyroxene-amphibole-garnet assemblage coexists with melt to pressures perhaps as low as 10 kb, at 150°C below the liquidus. This may be a garnet-poor assemblage; no mineral proportions provided. The appearance of plagioclase is restricted to $T \leq 950^\circ C$ for this composition, and to $T \leq 1100^\circ C$ for the 2% H_2O composition.

The experimental work of Stern et al. (1975) on a tonalitic composition also indicates the presence of garnet and clinopyroxene near the liquidus at 15-20 kb for a wide range of H_2O contents. With some assumptions, these authors have constructed H_2O-undersaturated liquidus surfaces for this composition, summarizing the liquidus and near-liquidus mineralogy (Fig. 54). From this diagram it appears possible to derive tonalitic melts from a gabbroic composition under both high pressure (garnet-clinopyroxene) and low pressure (pyroxene-plagioclase) conditions.

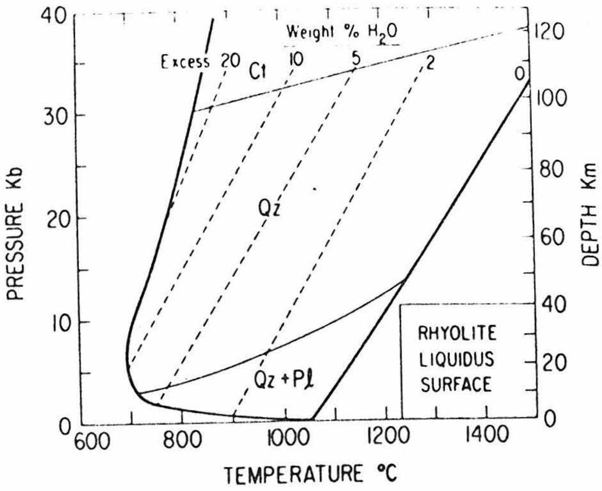
However, similar experiments on rhyodacitic to rhyolitic compositions do not suggest that these compositions can be derived directly from a garnet-pyroxene source. Rhyodacites with 2% and 5% H_2O have garnet-quartz and quartz, respectively, on their liquidii for pressures greater than 13.5 kb (Green and Ringwood, 1972). A H_2O-undersaturated

Figure 54. Contoured map of H_2O-undersaturated liquid surfaces for andesite (A) and rhyolite (B), from Stern et al. (1975).

A


Weight % H_2O
Excess 20 10 5 2 0
Ga
Ga + Cpx
Hb + Ga + Cpx
Cpx
Hb + Cpx
Px
Pl
Pl
ANDESITE LIQUIDUS SURFACE
PRESSURE Kb
0
10
20
30
40
TEMPERATURE °C
800
1000
1200
1400
1600
DEPTH Km
0
20
40
60
80
100
120

B


Weight % H_2O
Excess 20 10 5 2 0
Cl
Qz
Qz + Pl
RHYOLITE LIQUIDUS SURFACE
PRESSURE Kb
0
10
20
30
40
TEMPERATURE °C
600
800
1000
1200
1400
DEPTH Km
0
20
40
60
80
100
120

liquidus surface for a SiO_2-rich rhyolite (also in Fig. 54) suggests that this composition could not be in equilibrium with garnet and pyroxene over the pressure range of the experiments.

DISCUSSION AND SUMMARY

The experimental observations presented above appear to permit and in some cases support inferences made previously about contrasts in source region mineralogies across the batholith. To illustrate the adequacy of a gabbroic and eclogitic source material for the tonalites of the western and central regions, respectively, calculated REE patterns for melts produced from a basaltic composition in these mineralogies are presented (Fig. 55). In both cases the parental material material used has a slightly light REE depleted pattern typical of mid-ocean ridge tholeiites or island arc tholeiites. The selection of a source whose REE pattern is light REE depleted relative to chondritic abundances receives support from a few Nd isotopic measurements on rocks from the western region of the batholith (DePaolo, 1978).

To produce tonalites, 30% melting of the basaltic source has been chosen to be consistent with experimental data discussed above. Calculations are performed according to the equation of Shaw (1970) (Table 11). Partial melting under equilibrium conditions is assumed, leaving a residue of 50% hornblende-50% plagioclase or 50% hornblende-50% garnet. The effect of the presence of some clinopyroxene in

Figure 55. Calculated REE patterns of melts derived from a parental material with a REE pattern resembling oceanic tholeiites by 30% partial melting. Residual assemblages are either equal amounts of hornblende and plagioclase or hornblende and garnet. Equations and partition coefficients used are presented in Table 11.

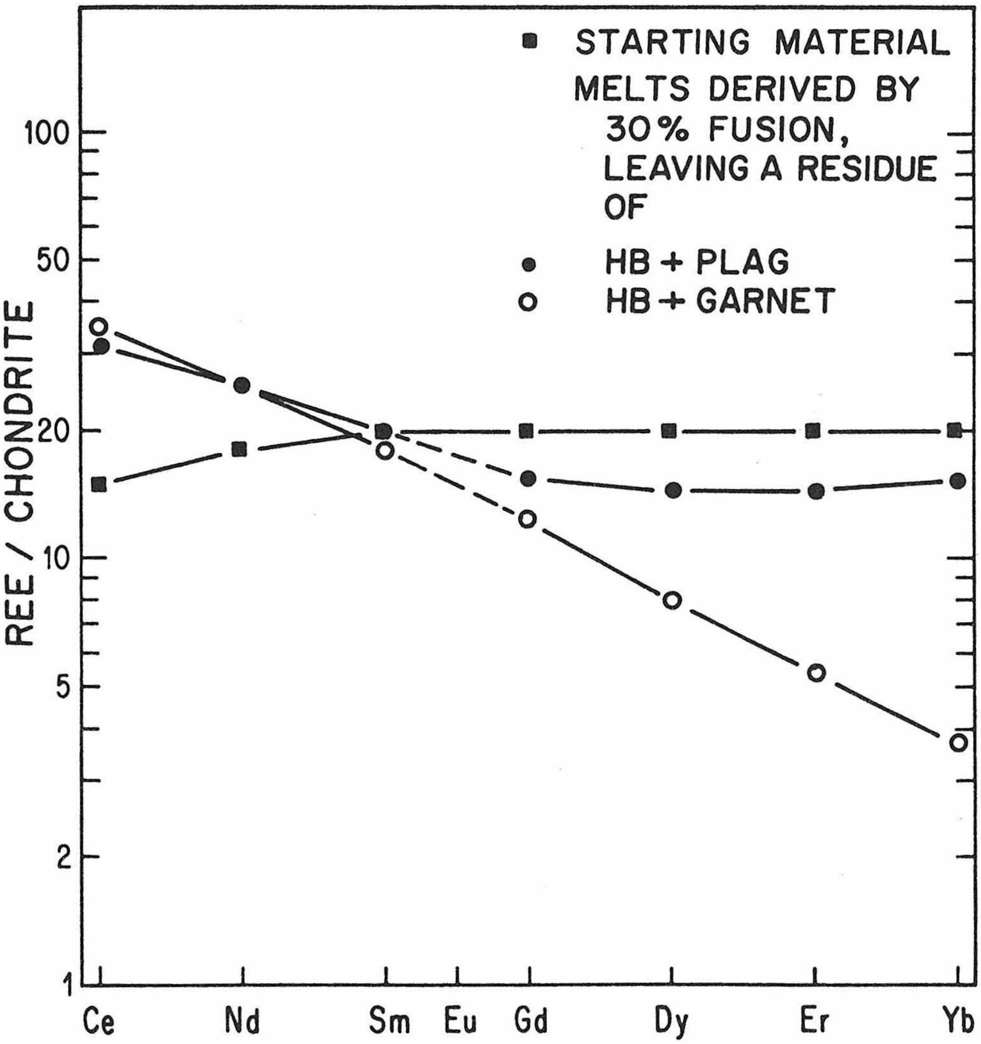

■ STARTING MATERIAL
MELTS DERIVED BY 30% FUSION, LEAVING A RESIDUE OF
● HB + PLAG
○ HB + GARNET
REE / CHONDRITE
100
50
20
10
5
2
1
Ce
Nd
Sm
Eu
Gd
Dy
Er
Yb

Table 11

DISTRIBUTION COEFFICIENTS USED IN CALCULATIONS FOR FIGURE 55

	HORNBLENDE	PLAGIOCLASE	GARNET
Ce	.35	.175	.03
Nd	1.10	.069	.07
Sm	2.01	.026	.29
Eu	-	-	-
Gd	2.82	.015	.97
Dy	3.15	.008	3.17
Er	3.17	.007	6.56
Yb	2.84	.007	11.5

EQUATION FOR TRACE ELEMENT FRACTIONATION DURING MELTING

$$C_1/C_o = 1/(D_o + F(1-D_o)) \qquad \text{Shaw (1970)}$$

where

C_1 = concentration of element in liquid

C_o = concentration of element in initial material

D_o = weighted bulk distribution coefficient (mineral/melt). This value is a constant.

F = proportion of melt

the place of hornblende in either assemblage will be minimal as REE behave similarly in these minerals. Distribution coefficients employed are considered appropriate in this compositional system, and are listed in Table 11. Values for Eu have not been assigned as these are quite dependent on several variables that cannot be sufficiently defined, the most important of which is fO_2.

The results presented in Figure 55 indicate that these assemblages are capable of producing REE patterns corresponding reasonably to the tonalites of the western and central regions. It is stressed that the above exercise does not provide unique solutions, but simply demonstrates that conclusions drawn on the basis of qualitative relationships in the REE data, on petrographic observations, and on appropriate experimental work are in acceptable agreement with model calculations.

The tonalites of the eastern region require greater enrichment of the light REE. The similarity in the bulk composition of these tonalites to those to the west, including K_2O (and Rb), suggest that the source composition and the degree of partial melting were broadly equivalent and that, most probably, the source itself was richer and more fractionated in light REE. Basalts with moderately light REE enriched patterns commonly occur in island arcs and continental regions, suggesting the possible availability of such sources.

In the western region, melting at pseudo-invariant points or loci of reduced variance are attractive mechanisms to produce a diverse lithologic suite with high relative abundances of particular compositions. The small degree of melting of a gabbroic source necessary to produce siliceous leucogranodiorites of the Woodson Mountain type allows trace element fractionations to be sensitive to oscillations in amount of melt produced and in mineral proportions. These factors may contribute to the large variation in REE characteristics observed in these rocks. The more extensive melting suggested for tonalitic to quartz gabbroic compositions leads to less sensitivity to these parameters, and may be in part or perhaps largely responsible for the limited variation in REE and other trace element characteristics of these rocks.

A garnet-bearing but plagioclase-poor or free residual assemblage for the source of the tonalites of the central and eastern regions is permitted by the experimental studies on andesite with reasonable amounts of H_2O. Pressures above 15 kb will give an appropriate assemblage in equilibrium with tonalitic melts, with temperatures of $1000^{\circ}C$ and above. However, the negative Eu anomalies and negatively sloped arrays on Rb-Sr diagrams for granodioritic rocks of the eastern region in southern California suggest the involvement of plagioclase in the evolution of these rocks. This can be accounted for in several ways. One possibility is that these features are produced during differentiation at a level higher than the source region, the lower pressure permitting plagioclase to be stable. However, large masses of H_2O-undersaturated magma do not intersect their liquidus curves upon adiabatic rise until they reach shallow levels. Appreciable differentiation during ascent is not expected, but is plausible near the level of emplacement. Bearing on this consideration is the observation that a small proportion of the rocks of the central and eastern regions are quite SiO_2-rich. This can be seen for the northeasternmost part of the batholith in the small number of rocks with high Larsen indices (see Fig. 40), but is also known from more southerly parts of the batholith (Silver, per. comm.). Experimental data discussed above suggest that SiO_2-rich melts could not form directly from garnet-clinopyroxene sources, and implies low-pressure differentiation has participated in the evolution of at least these lithologies. The low proportion of SiO_2-rich lithologies may suggest that high-level differentiation processes have been largely ineffective in the central and eastern parts of the batholith, although it must be kept in mind that the deeper levels of erosion in these areas may have removed a more differentiated top to the batholith.

It is equally important to emphasize that most of the central and eastern region rocks, including many of the granodiorites, are not SiO_2-rich and possess some petrographic features not expected in strongly differentiated rocks. For example, the granodiorite near Indio (sample 19) has an intermediate SiO_2 content and an andesine plagioclase (Appendix I), both of which are also characteristic of tonalites of the batholith. The major difference between this type of granodiorite and the tonalites is the amount of alkali feldspar. Therefore, many of the rocks now exposed possibly have crystallized directly from melts derived at depth.

If some of the granodiorites are direct partial melts of their sources, they may have formed by smaller degrees of partial melting and/or possibly at shallower depths than the tonalites. This permits the appearance of plagioclase in the residual mineralogy if physical conditions are appropriate (i.e. within or very near the garnet granulite stability field). If this were true, an important limitation to the depth of the eastern source regions in southern California is obtained. An alternative and possibly complementary hypothesis must be considered. The importance of pH_2O in the control of subsolidus and residual mineralogies has been noted in the experimental investigations cited above. Modest variations in pH_2O within the source regions strongly influence the relative stabilities of plagioclase, amphibole, and pyroxene. In light of this, consider that the granodiorites analyzed in Baja California generally do not display negative Eu anomalies and have near vertical arrays on Rb-Sr diagrams, suggesting a much less important role for plagioclase. A wetter source region for many of the rocks of the eastern region in Baja California compared to southern California would tend to eliminate supersolidus plagioclase. Hence, for all degrees of partial melting, no negative Eu anomalies would be developed and Rb and Sr would fractionate differently. This possibility of a wetter source also receives support from other REE features. Many of the tonalites and granodiorites from the central and eastern regions in Baja California have comparatively low REE contents and small positive Eu anomalies. These features are consistent with a more amphibole-rich residual mineralogy, which can be a direct consequence of higher pH_2O. Additionally, the anomalous western tonalite in Baja California (sample 31, Fig. 32) might have been derived from a more hornblende-rich source relative to the other tonalites rather than by hornblende-dominated fractional crystallization. The hornblende-rich character of this rock is in accord with the former.

Thus it appears that there are several processes which are consistent with the experimental investigations, and which may have contributed to the evolution of these rocks and their REE patterns. The relative importance of these processes, particularly those which participated in the differentiation that gave rise to the eastern granodiorites, are not fully resolved. These questions may yield to more detailed investigations.

The above do have some implications for the origin of the gabbros, previous discussions of which indicated that the magmas from which these rocks crystallized may have already possessed some of their unusual characteristics. The difficulties in relating these rocks to the quartz gabbros and tonalites has been discussed, but in the western region a rather interesting complementarity among the REE patterns of some of the siliceous leucogranodiorites and the gabbros has been noted. This complementariness is particularly well expressed in the light REE and in Eu anomalies, but also extends to the large variability observed within both rock types. These rocks are not thought to be related by high-level processes as they occur as distinct bodies in the field and often do not have mutually compatible mineral assemblages. However, if the dominant process of differentiation in the batholith was variable partial melting in the source, as suggested by previous considerations, the remelting of source material which has already produced some melt provides a simple mechanism for generating magmas with depleted characteristics. Admittedly, this requires the attainment of extreme temperatures within the source, but a gabbroic magma capable of precipitating fosteritic olivine and anorthitic plagioclase requires such temperatures.

The suggestion raised here is that the siliceous leucogranodiorites represent an initial melt, leaving a residual source. This residual material is remelted to give at least a component of the gabbroic magmas. The initial melts might be expected to be among the first magmas to be emplaced. Geochronological studies of Silver and his coworkers (Banks and Silver, 1966; Silver, per. comm.) have shown siliceous leucogranodiorites to be included among the earliest of the intrusives, but they actually span the range of ages in the western region.

This hypothesis may offer some explanation for the presence of gabbro and siliceous leucogranodiorite together in the western region in southern California, but their near absence in the central and eastern regions. The formation of a small amount of partial melt may be controlled by an invariant point, as suggested previously. If this invariant point is destroyed by a major phase change, the generation of both lithologies may be suppressed. The fact that this lithologic distinction (at least within southern California) occurs at the same geographic position as the abrupt change in REE characteristics reflects favorably on this concept.

A perplexing aspect of the gabbros is their tendency to sometimes be slightly lower in $^{87}Sr/^{86}Sr$ initial ratios and ^{18}O than adjacent quartz-bearing rocks. A difference in Sr isotopic composition among rocks derived during a limited time interval from the same homogeneous source reservoir is not expected. The variability in the magnitude of the Sr isotopic discrepancy (Silver, per. comm.) requires an additional contribution to the local system. The possibility that this hypothetical component was magmatic and supplied the heat necessary to cause melting of a residual source is explored in a separate discussion.

Summarizing, different residual assemblages and source materials appear to provide the dominating influence on the western-central and central-eastern region transitions, respectively. Integration of observations on REE and other geochemical and petrologic characteristics with experimental studies has established the plausibility of a gabbroic residual mineralogy for the western region and a garnet granulitic or eclogitic residual mineralogy for the central and eastern regions. This change from low to high pressure conditions appears to explain why several different and presumably independent parameters undergo coupled transitions.

These variations occur in a regular, systematic arrangement across the batholith, reflecting its structure. These variations do not appear to be simply part of a single, unified process of which the rocks of the western and eastern regions are endmembers mixed to form the central region rocks. Indeed, the REE patterns of the central region rocks cannot be expressed as a combination of the REE patterns of western and eastern rocks. The transitions in REE character are abrupt, and as developed above appear to require explanations of different natures. This leads to the question of what is the mechanism which caused these different assemblages and materials to occur in their orderly arrangement across the batholith, and how closely is this tied to the geometry and dynamics of the subduction process and to the character of the environments across which the batholithic arc was developed. These questions are most advantageously addressed along with further development of arguments for the depth constraints, thermal requirements, and geologic context of the source materials.

INFERENCES CONCERNING THE NATURE OF THE MELTING PROCESSES AND THE ORIGIN OF THE SOURCE MATERIALS

LOCATION AND THERMAL REQUIREMENTS OF THE SOURCE REGIONS

The experimental considerations, if correct, place some constraints on the depth of the source regions. Plagioclase feldspar is likely to be present in significant abundances in a supersolidus assemblage of basaltic bulk composition at pressures of 10-15 kb or less. The maximum depths at which the western source regions are inferred to have equilibrated are therefore 30 to 45 km. The source regions of the central and eastern regions require pressures of 15 kb and greater to stabilize garnet. Pressures cannot greatly exceed this value if some plagioclase is to be present in the source under subsolidus to slightly supersolidus conditions; otherwise, no upper limit on pressure is provided.

These depth constraints have important implications for both the location of source materials within the lower crust or upper mantle and in the thermal requirements for production of batholithic melts. Presently the crustal thickness as averaged seismically along the length of the peninsula is approximately 26 km, with some suggestion of a west-to-east thickening giving values as high as 43 km beneath the Sierra Juarez (Thatcher, 1972). Several geologic and petrologic features in the western Peninsular Ranges (including: discordant intrusive contacts; occasional miarolitic cavities; low grade of regional metamorphism and largely undisturbed prebatholithic section) suggest no more than several kilometers of crust have been eroded since emplacement of the batholith, while in the eastern parts of the province perhaps 10-15 km or possibly more erosion has exposed a much deeper level of emplacement. Deduced post-emplacement crustal thicknexxes of roughly 30 km for the western region and 35 to 45 km (but perhaps locally thicker than 50 km) must be reduced to obtain true prebatholithic thicknesses as the emplacement of the batholith constituted a major addition of material to the crust.

It appears that pressures prevalent in the source regions of the batholith inferred from the REE data and petrologic considerations place the sources largely if not entirely below the base of the crust for all regions of the peninsula, to the best that its location can be estimated.

Geothermal gradients estimated from experimental knowledge of the temperature of batholithic magmas (in excess of

1000°C for H_2O-undersaturated tonalite or granodiorite melt) and their inferred depth of origin are at least 100°C/kb to 50°C/kb for the western and eastern regions, respectively. These values are unrealistically high for a regionally developed geothermal gradient in a continental environment, or even an oceanic environment for the case of the western region. The simple fact that batholiths such as the Peninsular Ranges do not underlie all continental areas but rather are localized in curvilinear arcs is an indication of the unusual thermal requirements needed to form these magmatic rocks. An intense, highly localized thermal regime must have existed during their formation.

CONSIDERATION OF THE ROLE OF CONVERGENT PLATE MOTION

The increasing depth to the east implied for the source regions is suggestive of the asymmetry of a subduction zone, but direct derivation of batholithic magmas from the subducted slab is improbable. A cool slab is not expected to be within its melting range at the depths in question. The slab almost certainly must be deeper than the source regions implied above, but may have contributed to them less directly. Substantial quantities of heat energy must have been transported to the source regions, most probably by a transient heat source from below. This heat source could have been in various forms, but its inferred long curvilinear shape seems to relate it closely to convergent plate motions and perhaps the subduction process. The possibilities include magmas derived from the vicinity of the subduction zone, such as basalts derived from the fluxing of the mantle overlying the subducted slab with H_2O and possibly other slab components. Alternatively, hydrous pyroxenite diapiric masses may be produced above the subduction zone by reaction of slab components and overlying peridotite (Ringwood, 1974) could rise largely as solid masses and transport heat upward.

Such processes, being controlled by convergent plate motions, may offer some explanation for the spatial distribution of the batholithic magmas. A critical constraint is the time progression of placements across the peninsula (Silver et al., 1975). Recent evaluations (Silver, per. comm.) have shown that a broad locus of magmatism was static in the western part of the Peninsular Ranges for 20-25 my, after which the locus became transgressive and progressed eastward. This evolution in magmatic style almost certainly reflects some change in the dynamics and perhaps geometry in the interaction of converging plates. This transition appears to closely but not in all places precisely follow the western to central region REE transition, and seems to imply some dynamic control on REE characteristics. The abruptness of the western to central REE transition supports this. A smoothed, gradational variation would be expected to result from generation within a steady state or static setting, as the phase changes which are implied occur by sliding reactions.

To pursue further what additional controls on the characterization of the source regions and melting processes might be provided, a comparison of the REE variations observed in the Peninsular Ranges to those in island and continental margin arcs is made. In the westernmost Peninsular Ranges in Baja California, distinctive flat REE patterns with slight relative depletions of the light REE observed in the San Telmo pluton strongly resemble the early, near trench volcano-plutonic elements of oceanic island arcs (e. g. Gill, 1970). The geologic setting of the San Telmo pluton, particularly its emplacement as a high-level ring complex into a thick, little disturbed volcaniclastic section, is also directly analogous to oceanic island arcs.

The REE fractionation patterns of the western tonalites and quartz gabbros are remarkably similar to those of circum-Pacific high-alumina basalts (see Fig. 11) with the exception of Eu anomalies. It would appear that this form of pattern is a fundamental characteristic of many mafic to intermediate rocks of circum-Pacific magmatic arcs and that they may owe their origin to the same basic processes or materials. An important contrasting feature of the REE patterns of these batholithic rocks is the consistent moderate negative Eu anomaly, which is not generally present in the high-alumina basalts. This difference, presumably attributable to some differentiation process involving plagioclase, may be related to the generally more siliceous composition of the quartz gabbros and tonalites. However, simple fractional crystallization of high-alumina basalt to produce the western quartz gabbros through tonalites seems unlikely due to the uniformity of the REE patterns among these plutonic rocks, especially the Eu anomaly.

In contrast, the heavy REE depleted patterns of the rocks of the central and eastern parts of the batholith have not been observed in oceanic island arcs, and have only occasionally been reported in continental margin arcs.

The strong light REE enrichments which typify the rocks of the eastern region of the batholith are observed away from the associated trench in both oceanic and continental margin magmatic arcs. Therefore materials and processes capable of producing this particular aspect of eastern patterns appear to be available and operative in analogous positions in oceanic and continental margin arcs and the batholith.

Thus, many aspects of the variation of REE characteristics across the batholith appear to reflect environments which occur across the continental margin. The similarity in many of these features to those observed in modern magmatic arcs lends additional support to the previously established relationship between the batholith and a convergent plate boundary. However, the striking contrast in the abundance of heavy REE depleted rocks in the batholith as compared to their absence in oceanic arcs appears to reflect a fundamental difference in their setting.

This leads to consideration of how continental crust and lithosphere might influence and participate in the generation of the magmatic products at convergent plate boundaries. It is appropriate to readdress the observations on isotopic variations in the batholith. The increasing initial $^{87}Sr/^{86}Sr$ ratios across the batholith appear to be an intrinsic property of the source regions. These variations may arise either by long-term differences of Rb/Sr within the source regions, or by physically mixing two or more components within the source. Long-term differences of Rb/Sr in the source regions may be reasonable as the mantle is not known to be homogeneous, either vertically or laterally. However, some systematic control on these variations is required to explain the variations observed in the batholith. Geophysical studies have suggested that subcontinental mantle is a rigid part of the lithosphere to depths of several hundred kilometers (e.g. Jordan, 1975). The continental crust and lithosphere may have acted as a thick, rigid, low density layer which served to localize the formation and/or segregation of middle to heavy REE depleted and fractionated magmas. It might be speculated that the thickness of the crust at a convergent plate boundary is an important factor in determining the depth of the source regions.

Production of the variations in ^{18}O observed in the batholith by the above processes remains difficult (Taylor and Silver, 1978). This warrants consideration of crustal components in the source. Crustal materials more sialic than basalt may participate in the generation of batholithic magmas as a component in the source, but with some firm limitations. Consider the observation that all the quartz-bearing rocks of the region of the batholith display characteristic REE, Sr, and O isotopic properties. Incorporation of sialic crustal components in the source regions cannot be responsible for both the local lithologic variation (such as by having more sialic component present to obtain a more granitic derived melt) and the regional trace element and isotopic variations. Rather, the sialic component must, if present, be a fairly uniformly distributed component of the source and contribute to all lithologies equally if it is to explain the trace element and isotopic variations. Therefore this hypothetical sialic component must be of a calcic nature or low enough in abundance such that the source remains capable of producing the calcic compositions that characterize the batholith. Tonalites and low-K_2O granodiorites of similar bulk composition are the dominant rock types across the Peninsular Ranges. A major change in the bulk composition of the source regions from which these rocks were derived is therefore not expected.

This argument is supported by the observtions on Sr isotopic compositions and REE patterns. The involvement of large proportions of old radiogenic crustal materials appears ruled out by the comparatively unradiogenic character of the entire batholith, including the eastern regions. Average upper crustal material is likely to have high absolute but unfractionated heavy REE abundances, such as represented by shales (see Fig. 9). Most of the eastern region rocks have low and fractionated heavy REE abundances, much like the central region rocks, despite the more radiogenic Sr and higher $\delta^{18}O$ values of the eastern rocks. Increased involvement of such crustal materials in the source regions of the eastern rocks is not suggested.

At this time, it does not appear possible to offer an explanation which adequately reconciles the firm constraint provided by the bulk composition of the batholith, the limitations provided by the observations on Sr isotopic composition and REE, and known sources of high-^{18}O materials.

CONCLUSIONS AND IMPLICATIONS

The most attractive working hypothesis for the origin of the REE and other variations across the batholith appears to be a combination of mechanisms which occur within the source regions. High-level processes including crystal fractionation and contamination are of no more than minor importance in producing the basic geographic trends.

The decoupling of REE and Sr concentration variations from the isotopic trends across the western to central REE transition argues for distinct origins of these features. An abrupt change from plagioclase to garnet-bearing assemblages in the source regions is capable of explaining the REE fractionations, elimination of negative Eu anomalies, increases in Sr concentration, and possibly the restriction in lithologic diversity observed across the boundary between the western and central regions. That these four significant parameters appear to change together is a strong argument for such a related origin. However, it has not yet been demonstrated that plagioclase and garnet-bearing sources can produce as similar bulk compositions as are seen within all regions of the batholith (e.g. tonalites and low-K_2O granodiorites). This inferred change in phase assemblage appears to have been superimposed upon a source region already zoned in $^{87}Sr/^{86}Sr$, $\delta^{18}O$, light REE abundances, and possibly Sr content. These variations may reflect long-term or inherent properties of the source regions, or they may be produced by some mixing process within the sources, or both processes to some degree. Components participating in a mixing process may be of both mantle and crustal origin, but must be capable of generating tonalitic and low-K_2O granodioritic melts over the entire region. Additionally, the observed regularities in the regional variations of the above properties require that components be mixed in a very systematic manner, the controls on which are poorly defined.

The energy needed to generate melt in the source regions is inferred to be supplied by convergent plate motions. Temperature and pressure constraints on the source regions indicate the batholithic magmas cannot be directly derived from subducted slab materials or from mantle components immediately overlying the slab. It may be speculated that magmas generated in the vicinity of the subduction zone rose and contributed heat and possibly material to the source regions of the batholith. Perhaps this mechanism determined the elongate form of the batholith as well as the general alignment of all the variations along its length.

The REE-fractionated rocks of the central and eastern regions appear to require sources at pressures high enough to yield garnet-bearing assemblages. The abruptness of the western to central REE transition seems inconsistent with the gradual phase transformation expected from gabbroic to garnet granulitic to eclogitic assemblages in basaltic source materials with increasing depth. The abruptness of the REE transition and its fairly close correlation to a transition from a static to transgressive arc suggests that this feature is related to a change in the dynamics and possibly geometry of the interaction of plates at a convergent boundary. The static arc tapped relatively shallow sources whereas the transgressive arc sampled deeper garnet-bearing sources as it moved across the continental margin. Comparison to REE variations in island arcs suggests thicker continental lithosphere may play an important role in providing sites at appropriate depths for the formation and/or segregation of middle to heavy REE depleted and fractionated batholithic magmas. Such sites do not appear to be present in oceanic island arc environments.

Details of the nature of the heat sources remain largely unresolved. Further study of the gabbroic rocks may yield additional information as these rocks are the most probable to have been intimately associated with the heat sources. The Sr and O isotopic compositions of the gabbros, which often differ from adjacent quartz-bearing rocks, may be an indication of material contributions. The complex petrologic features of the gabbros, particularly those developed at high levels, must be better understood, however, before deeper-seated characterisitcs can be confidently inferred.

It may be possible to obtain a better understanding of the variations within the batholith by extending sampling to the east. The present investigation has been limited to the locus of intense magmatic activity within the Peninsular Ranges. Regional studies (e.g. Silver et al., 1975) have established that the same cycle of magmatic activity transgressed eastward, onto the Precambrian craton in Sonora, Mexico. A major change in the petrologic character of the plutonic rocks across the cratonic boundary has been noted

(Silver et al., 1975). If the batholith had sources within the continental lithosphere, plutons which were emplaced into the craton should be more uniquely indicative of lithospheric components. It is not yet known whether these rocks display variations in geochemical characteristics which are more extreme than rocks to the west. Would the batholithic rocks have REE patterns approximating shale composites? Geochemical studies of these rocks should be pursued so that the origin of the variations across all segments of the batholith can be more effectively evaluated.

The growth and evolution of the continents by orogenic episodes is well established. Granitic batholiths and associated volcanic and volcanic-derived rock masses are major if not dominating components of the continental crust. The observed variations in REE and other characteristics across the Peninsular Ranges batholith provide an example of the contributions of a simple and relatively recent orogenic episode to the North American continent. This is the first report of systematic variations in REE characteristics across a granitic batholith whose geologic relationships have established its formation at a convergent plate boundary. It must be asked whether batholithic arcs of the same age and analogous setting display similar variations, and if older arcs also show related features.

The present study emphasizes that REE investigations of other orogenic terrains might be evaluated in light of the large but systematic variations observed across a single magmatic arc. These variations appear to have a simple geometric relationship to a convergent plate boundary and inferred subduction zone. Complex orogenic terrains which are comprised of multiple superimposed plutonic arcs in differing orientations may display complex or apparently nonsystematic variations in REE and other geochemical characteristics related to the geometry of the plate boundary. In terrains where field and geochronologic studies have identified and resolved the multiple arc components, the transverse REE variations observed in the Peninsular Ranges batholith may be useful in establishing the orientation of the plate boundaries and for comparing magmatic arc genesis in different age regions. This may be particularly valuable in Precambrian terrains which have had extended deformational and metamorphic histories and where much of the geologic evidence for original setting and genesis has been lost. In regions where an adequate geochronologic framework does not exist, REE variations may be of value in aiding the resolution of superimposed arcs. In settings such as the southwestern North American continent, where large lateral shears have separated and juxtaposed segments of various magmatic arcs, REE and other variations should provide additional constraints for the identification and reconstruction of tectonic components.

LIST OF REFERENCES

Allen, J. C., Modreski, P. J., Haygood, C., and Boettcher, A. L., 1972, The role of water in the mantle of the Earth: The stability of amphiboles and micas: Proc. 24th Inter. Geol. Congr., Sec. 2, 231-240.

Allison, E. C., 1955, Middle Cretaceous gastropoda from Punta China, Baja California, Mexico: Jour. Paleontology 29, 400-432.

Anderson, T. H. and Silver, L. T., 1977, Geochronometric and stratigraphic outlines of the Precambrian rocks of northwestern Mexico (abs.): Geol. Soc. Amer. Abstracts with Programs 9, 880.

Armstrong, R. L., Taubeneck, W. H., and Hales, P. O., 1977, Geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho: Geol. Soc. Amer. Bull. 88, 397-411.

Arth, J. G. and Barker, F., 1976, Rare-earth partitioning between hornblende and dacitic liquid and implications for the genesis of trondhjemitic-tonalitic magmas: Geology 4, 534-536.

Arth, J. G. and Hanson, G. N., 1972, Quartz diorites derived by partial melting of eclogite or amphibolite at mantle depths: Contr. Mineral. Petrol. 37, 161-174.

Arth, J. G. and Hanson, G. N., 1975, Geochemistry and origin of the early Precambrian crust of northeastern Minnesota: Geochim. Cosmochim. Acta 39, 325-362.

Baird, A. K., Baird, K. W., and Welday, E. E., 1974a, Chemical trends across Cretaceous batholithic rocks of southern California: Geology 4, 493-495.

Baird, A. K., Morton, D. M., Woodford, A. O., and Baird, K. W., 1974b, Transverse Ranges Province: A unique structural-petrochemical belt across the San Andreas fault system: Geol. Soc. Amer. Bull. 85, 163-174.

Balashov, Yu. A., Ronov, A. B., Migdisov, A. A., and Turanskaya, N. V., 1964, The effect of climate and facies environment on the fractionation of the rare earths during sedimentation: Geochim. Intern. 2, 951-969.

Banks, P. O. and Silver, L. T., 1966, Evaluation of the decay constant of uranium-235 from lead isotope ratios: J. Geophys. Res. 71, 4037-4046.

Banks, P. O. and Silver, L. T., 1969, U-Pb isotope analyses of zircons from Cretaceous plutons of the Peninsular and Transverse Ranges, southern California (abs.): Geol. Soc. Amer. Spec. Paper 121, 17-18.

Barker, F., Arth, J. G., Peterman, Z. E., and Friedman, I., 1976, The 1.7 to 1.8 b.y. old trondhjemites of southwestern Colorado and northern New Mexico: geochemistry and depths of genesis: Geol. Soc. Amer. Bull. 87, 189-198.

Bateman, P. C., Clark, L. D., Huber, N. K., Moore, J. G., and Rinehart, C. D.,1963, The Sierra Nevada batholith - a synthesis of recent work across the central part: U. S. Geol. Survey Prof. Paper 414-D, D1-D46.

Bateman, P. C. and Dodge, F. C. W., 1970, Variations of major chemical constituents across the central Sierra Nevada batholith: Geol. Soc. Amer. Bull. 81, 409-420.

Buma, G., Frey, F. A. and Wones, D. R., 1971, New England granites: trace element evidence regarding their origin and differentiation: Contr. Mineral. Petrol. 31, 300-320.

Chappell, B. W., 1968, Volcanic greywackes from the Upper Devonian Baldwin Formation, Tamworth-Barraba District, New South Wales: J. Geol. Soc. Aust. 15, 87-102.

Condie, K. C. and Hayslip, D. L., 1975, Young bimodal volcanism at Medicine Lake volcanic center, northern California: Geochim. Cosmochim. Acta 39, 1165-1178.

Condie, K. C. and Swenson, D. J., 1973, Composition variation in three Cascade stratovolcanoes: Jefferson, Rainier, and Shasta: Bull. Volcan. 37, 205-230.

Coryell, C. D., Chase, J. W., and Winchester, J. W., 1963, A procedure for geochemical interpretation of terrestrial rare-earth abundance patterns: J. Geophys. Res. 68, 559-566.

Creasey, S. C., 1946, Geology and nickel mineralization of the Julian-Cuyamaca area, San Diego county, California: Cal. Jour. Mines Geology 42, 15-29.

DePaolo, D. J., 1978, Study of magma sources, mantle structure and the differentiation of the earth from variations of $^{143}Nd/^{144}Nd$ in igneous rocks: Ph.D. thesis, Cal. Inst. Tech., 373 p.

Drake, M. J. and Weill, D. F., 1975, Partition of Sr, Ba, Ca, Y , Eu^{2+}, and Eu^{3+}, and other REE between plagioclase feldspar and magmatic liquid: an experimental study: Geochim. Cosmochim. Acta 39, 689-712.

Duval, C., 1963, *Inorganic Thermogravimetric Analysis*, 2nd ed.: Elsevier Publishing Co., Amsterdam.

Dymek, R. F., 1977, Mineralogic and petrologic studies of Archaean metamorphic rocks from west Greenland, lunar samples, and the meteorite Kapoeta: Ph.D. thesis, Cal. Inst. Tech., 392 p.

Early, T. O. and Silver, L. T., 1973, Rb-Sr isotopic systematics in the Peninsular Ranges batholith of southern and Baja California (abs.): EOS (Trans. Amer. Geophysical Union) 54, 494.

Eggler, D. H., 1974, Application of a portion of the system $CaAl_2Si_2O_8$-$NaAlSi_3O_8$-SiO_2-MgO-Fe-O_2-H_2O-CO_2 to genesis of the calc-alkaline suite: Amer. J. Sci. 274, 297-315.

Eggler, D. H. and Burnham, C. W., 1973, Crystallization and fractionation trends in the system andesite-H_2O-CO_2-O_2 at pressures to 10 kilobars: Geol. Soc. Amer. Bull. 84, 2517-2532.

Essene, E. J. B., Hensen, B., and Green, D. H., 1970, Experimental study of amphibolite and eclogite stability: Phys. Earth Planetary Interiors 3, 378-384.

Everhart, D. L., 1951, Geology of the Cuyamaca Peak quadrangle, California: Cal. Div. Mines Bull. 159, 51-115.

Ewart, A., Bryan, W. B., and Gill, J. B., 1973, Mineralogy and geochemistry of the younger volcanic islands of Tonga, S. W. Pacific: J. Petrology 14, 429-465.

Fife, D. L., Minch, J. A., and Crampton, P. J., 1967, Late Jurassic age of Santiago Peak Volcanics, California: Geol. Soc. Amer. Bull. 78, 299-304.

Frey, F. A., Chappell, B. W., and Roy, S. D., 1978, Fractionation of rare-earth elements in the Tuolumne Intrusive Series, Sierra Nevada batholith, California: Geology 6, 239-242.

Frey, F. A., Haskin, M. A., Poetz, J. A., and Haskin, L. A., 1968, Rare-earth abundances in some basic rocks: J. Geophys. Res. 73, 6085-6098.

Gastil, R. G., Krummenacher, D., Doupont, J., and Bushee, J., 1974, The batholith belt of southern California and western Mexico: Pacific Geol. 8, 73-78.

Gastil, R. G., Phillips, R. P., and Allison, E. C., 1975, Reconnaissance geology of the state of Baja California: Geol. Soc. Amer. Mem. 140, 170 p.

Gill, J. B., 1970, Geochemistry of Viti Levu, Fiji, and its evolution as an island arc: Contr. Mineral. Petrol. 27, 179-203.

Gill, J. B., 1974, Role of underthrust oceanic crust in the genesis of a Fijian calc-alkaline suite: Contr. Mineral. Petrol. 43, 29-45.

Green, D. H., 1973, Experimental melting studies on a model upper mantle composition at high pressures under water-saturated and water-undersaturated conditions: Earth Planet. Sci. Lett. 19, 37-53.

Green, T. H., 1972, Crystallization of calc-alkaline andesite under controlled high-pressure hydrous conditions: Contr. Mineral. Petrol. 34, 150-166.

Green, T. H. and Ringwood, A. E., 1968, Genesis of the calc-alkaline igneous rock suite: Contr. Mineral. Petrol. 18, 105-162.

Green, T. H. and Ringwood, A. E., 1972, Crystallization of garnet-bearing rhyodacite under high pressure hydrous conditions: J. Geol. Soc. Aust. 19, 203-212.

Hamilton, W., 1969, Mesozoic California and the underflow of Pacific mantle: Geol. Soc. Amer. Bull. 80, 2409-2430.

Hanna, M. A., 1926, Geology of the La Jolla quadrangle, California: Univ. Cal. Dept. Geol. Bull. 16, 187-246.

Haskin, L. A. and Frey, F. A., 1966, Dispersed and not-so-rare earths: Science 152, 299-314.

Haskin, L. A., Haskin, M. A., Frey, F. A., and Wildeman, T. R., 1968, Relative and absolute terrestrial abundances of the rare earths: in Origin and Distribution of the Elements (Ed. L. H. Ahrens), Pergamon, New York, pp. 889-912.

Haskin, L. A. and Schmitt, R. A., 1967, Rare earth distributions: in Researches in Geochemistry, ed. P. H. Abelson, Wiley, New York, pp. 234-258.

Haskin, L. A., Wildeman, T. R., Frey, F. A., Collins, K. A., Keedy, C. R., and Haskin, M. A., 1966, Rare earths in sediments: J. Geophys. Res. 71, 2091-6105.

Helmke, P. A. and Haskin, L. A., 1973, Rare-earth elements, Co, Sc and Hf in the Steens Mountain basalts: Geochim. Cosmochim. Acta 37, 1513-1529.

Helz, R. T., 1973, Phase relations of basalts in their melting range at PH_2O=5kb as a function of oxygen fugasity, Part I. Mafic phases: J. Petrology 14, 249-302.

Helz, R. T., 1976, Phase relations of basalts in their melting ranges at PH_2O=5kb, Part II. Melt compositions: J. Petrology 17, 139-193.

Hess, P. C., 1977, Structure of silicate melts: Can. Mineral. 15, 162-178.

Higuchi, H. and Nagasawa, H., 1969, Partition of trace elements between rock-forming minerals and the host volcanic rocks: Earth Planet. Sci. Lett. 7, 281-287.

Hirschi, H. and de Quervain, F., 1933, Beiträge zur Petrographie von Baja California (Mexico) (Fortsetzung und Schluss): Schweiz. Min. Pet. Mitt. 13, 232-277.

Hofmann, A. W. and Hart, S. R., 1978, An assessment of local and regional isotopic equilibrium in the mantle: Earth Planet. Sci. Lett. 38, 44-62.

Holloway, J. R. and Burnham, C. W., 1972, Melting relations of basalt with equilibrium water pressure less than total pressure: J. Petrology 13, 1-29.

Hurlbut, C. S. Jr., 1935, Dark inclusions in a tonalite of southern California: Amer. Mineral. 20, 609-630.

Jahn, B., Shih, C., and Murthy, V. R., 1974, Trace element geochemistry of Archean volcanic rocks: Geochim. Cosmochim. Acta 38, 611-627.

Jakes, P. and Gill, J. B., 1970, Rare-earth elements and island arc tholeiitic series: Earth. Planet. Sci. Lett. 9, 17.

Jakes, P. and Taylor, S. R., 1974, Excess eruopium content in Precambrian sedimentary rocks and continental evolution: Geochim. Cosmochim. Acta 38, 739-745.

Jakes, P. and White, A. J. R., 1972, Major and trace element abundances in volcanic rocks of orogenic areas: Geol. Soc. Amer. Bull. 83, 29-40.

Jordan, T. H., 1975, The continental tectosphere: Rev. Geophys. Space Phys. 13, 1-12.

Kay, R., Hubbard, N. J., and Gast, P. W., 1970, Chemical characteristics and origin of oceanic ridge volcanic rocks: J. Geophys. Res. 75, 1585-1613.

Kistler, R. W., 1974, Phanerozoic batholiths of western North America: a summary of some recent work on variations in time, space, chemistry, and isotopic compositions: Annual Review Earth Planet. Sci. 2, 403-418.

Kistler, R. W., Evernden, J. F., and Shaw, H. R., 1971, Sierra Nevada plutonic cycle: Part I, Origin of composite granitic batholiths: Geol. Soc. Amer. Bull. 82, 853-868.

Kistler, R. W. and Peterman, Z. E., 1973, Variations in Sr, Rb , K , Na, and initial Sr^{87}/Sr^{86} in Mesozoic granitic rocks and intruded wall rocks in central California: Geol. Soc. Amer. Bull. 84, 3489-3512.

Krummenacher, D., Gastil, R. G., Bushee, J., and Doupont, J., 1975, K-Ar apparent ages, Peninsular Ranges batholith, southern California and Baja California: Geol. Soc. Amer. Bull. 86, 760-768.

Kuno, H., 1966, Lateral variation of basalt magma type across continental margins and island arcs: Bull. Volcan. 29, 195-222.

Lambert, I. B. and Wyllie, P. J., 1970, Melting in the deep crust and upper mantle and the nature of the low velocity layer: Phys. Earth Planetary Interiors 3, 316-322.

Lanphere, M. A. and Reed, B. L., 1973, Timing of Mesozoic and Cenozoic plutonic events in circum-Pacific North America: Geol. Soc. Amer. Bull. 84, 3773-3782.

Larsen, E. S. Jr., 1948, Batholith and associated rocks of Corona, Elsinore, and San Luis Rey quadrangles, southern California: Geol. Soc. Amer. Mem. 29, 182 pp.

Lindgren, W., 1915, The igneous geology of the Cordilleras and its problems: in Problems of American Geology, New Haven, Yale University, Silliman Foundation, pp. 234-286.

Lopez-Escobar, L., Frey, F. A., and Vergara, M., 1977, Andesites and high alumina basalts from the central-south Chile high Andes; geochemical evidence on their petrogenesis: Contr. Mineral. Petrol. 63, 199-228.

Masuda, A., Nakamura, N., Kurasawa, H., and Tanaka, T., 1972, Precision determination of rare-earth elements in the Apollo 14 and 15 samples: Geochim. Cosmochim. Acta, Suppl. 3, 1307-1313.

Masuda, A., 1968, Geochemistry of lianthanides in basalts of central Japan: Earth Planet. Sci. Lett. 4, 284.

Masuda, A., Nakamura, W., and Tanaka, T., 1973, Fine structures of mutually normalized rare-earth patterns of chondrites: Geochim. Cosmochim. Acta 37, 239-248.

McGetchin, T. R. and Silver, L. T., 1972, A crustal-upper-mantle model for the Colorado plateau based on observations of crystalline rock fragments in the Moses Rock dike: J. Geophys. Res. 77, 7022-7037.

Merriam, R. H., 1946, Igneous and metamorphic rocks of the southwestern part of the Ramona quadrangle, San Diego, California: Geol. Soc. Amer. Bull. 57, 223-260.

Miller, F. S., 1937, Petrology of the San Marcos gabbro, southern California: Geol. Soc. Amer. Bull. 48, 1397-1426.

Miller, W. J., 1935, A geologic section across the southern Peninsular Range of California: Cal. Jour. Mines Geology, State Mineral. Rept. 31, 123-126.

Miller, W. J., 1946, Crystalline rocks of southern California: Geol. Soc. Amer. Bull. 57, 457-540.

Miller, D. G., 1973, Plate-edge deformation and crustal growth, Gulf of California structural province: Geol. Soc. Amer. Bull. 84, 1883-1906.

Moore, J. G., 1959, The quartz diorite boundary line in the western United States: Jour. Geology 67, 198-210.

Morris, R. V. and Haskin, L. A., 1974, EPR measurement of the effect of glass composition on the oxidation states of europium: Geochim. Cosmochim. Acta 38, 1433-1445.

Murray, J. D., 1978, The structure and petrology of the San Jose pluton, northern Baja California, Mexico: Ph.D. thesis, Cal. Inst. Tech., 709 pp.

Nagasawa, H., 1970, Rare-earth concentrations in zircons and apatites and their host dacites and granites: Earth. Planet. Sci. Lett. 9, 359-364.

Nagasawa, H. and Schnetzler, C. C., 1971, Partitioning of rare-earth alkali and alkaline earth elements between phenocrysts and acidic igneous magma: Geochim. Cosmochim. Acta 35, 953-968.

Nance, W. B. and Taylor, S. R., 1976, Rare-earth element patterns and crustal evolution-I. Australian post-Archaen sedimentary rocks: Geochim. Cosmochim. Acta 40, 1539-1551.

Nance, W. B. and Taylor, S. R., 1977, Rare-earth element patterns and crustal evolution-II. Archaen sedimentary rocks from Kalgoorlie, Australia: Geochim. Cosmochim. Acta 41, 225-231.

Nicholls, I. A. and Ringwood, A. E., 1973, Effect of water on olivine stability in tholeiites and the production of silica-saturated magmas in the island arc environment: Jour. Geology 81, 285-300.

Nishimori, R. K., 1976, The petrology and geochemistry of gabbros from the Peninsular Ranges batholith, California, and a model for their origin: Ph.D. thesis, Univ. Cal. San Diego.

Osborn, E. F., 1939, Structural petrology of the Val Verde tonalite, southern California: Geol. Soc. Amer. Bull. 50, 921-950.

Philpotts, J. A., Martin, W., and Schnetzler, C. C., 1971, Geochemical aspects of some Japanese lavas: Earth Planet. Sci. Lett. 12, 89-96.

Philpotts, J. A., Schnetzler, C. C., and Thomas, H. H., 1972, Petrogenetic implications of some new geochemical data on eclogitic and ultrabasic inclusions: Geochim. Cosmochim. Acta 36, 1131-1166.

Potts, M. J., Early, T. O., and Herrmann, A. G., 1973, Determination of rare-earth element distribution patterns in rocks and minerals by neutron activation analysis: Z. Anal. Chem., Band 263, Heft 2, 97-100.

Presnall, D. C., O'Donnel, T. H., and Brenner, N. L., 1973, Cusps on solidus curves as controls for primary magma compositions: a mechanism for producing oceanic olivine tholeiites of uniform composition: Geol. Soc. Amer. Abstracts with Programs 5, 771-772.

Reed, B. L. and Lanphere, M. A., 1973, The Alaska-Aleutian Range batholith: Geochronology, chemistry, and relation to circum-Pacific plutonism: Geol. Soc. Amer. Bull. 84, 2583-2510.

Reed, B. L. and Lanphere, M. A., 1974, Chemical variations across the Alaska-Aleutian Range batholith: Jour. Research U.S.G.S. 2, 343-352.

Ringwood, A. E., 1974, The petrological evolution of island arc systems: J. Geol. Soc. 130, 183-204.

Ringwood, A. E., 1975, Composition and Petrology of the Earth's Mantle: McGraw-Hill, Inc., New York, 618 pp.

Saleeby, J., 1977, Fracture zone tectonics, continental margin fragmentation and emplacement of the Kings-Kaweah ophiolite belt, southwest Sierra Nevada, California: in North American Ophiolite Volume: Oregon Dept. Geol. and Min. Industries (Ed. R. G. Coleman and W. P. Irwin), Bull. 95, 141-160.

Schnetzler, C. C. and Philpotts, J. A., 1968, Partition coefficients of rare-earth elements and barium between igneous matrix material and rock forming mineral phenocrysts -I: in Origin and Distribution of the Elements (Ed. L. H. Ahrens), Pergamon, New York, pp. 929-938.

Schnetzler, C. C. and Philpotts, J. A., 1970, Partition coefficients of rare-earth elements between igneous matrix material and rock-forming mineral phenocrysts-II: Geochim. Cosmochim. Acta 34, 331-340.

Schwarcz, H. P., 1969, Pre-Cretaceous sedimentation and metamorphism in the Winchester area, northern Peninsular Ranges, California: Geol. Soc. Amer. Spec. Paper 100, 63 pp.

Shannon, R. D. and Prewitt, C., 1969, Effective ionic radii in oxides and fluorides: Acta Cryst. B25, 925-946.

Shannon, R. D. and Prewitt, C., 1970, Revised values of effective ionic radii: Acta Cryst. B26, 1046-1048.

Sharp, R. V., 1965, Geology of the San Jacinto fault zone in the Peninsular Ranges of southern California: Ph.D. thesis, Cal. Inst. Tech., 184 pp.

Sharp, R. V., 1967, San Jacinto fault zone in the Peninsular Ranges of southern California: Geol. Soc. Amer. Bull. 78, 705-729.

Shaw, D. M., 1970, Trace element fractionation during anatexis: Geochim. Cosmochim. Acta 34, 237-243.

Shimizu, N., 1975, Rare-earth elements in garnets and clinopyroxenes from garnet lherzolite nodules in kimberlites: Earth Planet. Sci. Lett. 25, 26-32.

Silberling, N. J., Schoellhamer, J. E., Gray, C. H. Jr., and Imlay, R. W., 1961, Upper Jurassic fossils from Bedford Canyon Formation, southern California: Amer. Assoc. Petroleum Geologists Bull. 45, 1746-1748.

Silver, L. T., 1968, Precambrian batholiths of Arizona (abs.): Geol. Soc. Amer. Spec. Paper 121, 558-559.

Silver, L. T., 1974, Presentation for IGCP circum-Pacific plutonic field conference, southern and Baja California: unpubl. proceedings.

Silver, L. T., Allen, C. R., and Stehli, F. G., 1969, Geological and geochronological observations on a portion of the Peninsular Ranges batholith of northwestern Baja California, Mexico: Geol. Soc. Amer. Spec. Pap. 121, 279.

Silver, L. T. and Anderson, T. H., 1974, Possible left-lateral early to middle Mesozoic disruption of the southwestern North American craton margin: Geol. Soc. Amer. Abstracts with Programs 6, 955-956.

Silver, L. T., Anderson, C. A., Crittenden, M., and Robertson, J. M., 1977, Chronostratigraphic elements of the Pre-Cambrian rocks of the southwestern and far western United States: Geol. Soc. Amer. Abstracts with Programs 9, 1176.

Silver, L. T. and Early, T. O., 1977, Rubidium-strontium fractionation domains in the Peninsular Ranges batholith and their implications for magmatic arc evolution (abs.): EOS (Trans. Amer. Geophysical Union) 58, 532.

Silver, L. T., Early, T. O. and Anderson, T. H., 1975, Petrological, geochemical and geochronological asymmetries of the Peninsular Ranges batholith: Geol. Soc. Amer. Abstracts with Programs 7, 375-376.

Silver, L. T., Stehli, F. G., and Allen, C. R., 1956, Lower Cretaceous prebatholithic rocks of northern Baja California, Mexico (abs.): 20th Internat. Geol. Congr., Mexico City, Resumenes de los Trabajos Presentados, 30.

Silver, L. T., Stehli, F. G., and Allen, C. R., 1963, Lower Cretaceous prebatholithic rocks of northern Baja California, Mexico: Amer. Assoc. Petroleum Geologists Bull. 47, 2054-2059.

Stern, C. R., 1974, Melting products of olivine tholeiite basalt in subduction zones: Geology 2, 227-230.

Stern, C. R., Huang, W. L., and Wyllie, P. J., 1975, Basalt-andesite-rhyolite-H_2O: crystallization intervals with excess H_2O and H_2O-undersaturated liquidus surfaces to 35 kilobars, with implications for magma genesis: Earth Planet. Sci. Lett. 28, 189-196.

Streckeisen, A. L., 1973, Plutonic rocks: classification and nomenclature recommended by the IUGS Subcommission on the Systematics of Igneous Rocks: Geotimes 18, 26-30.

Stuart, C. J., 1974, The San Onofre breccia in northwestern Baja California and its regional tectonic implications: in _Geology of Peninsular California_, Pacific Sections, AAPG, SEPM, and SEG, pp. 80-90.

Sun, S. S. and Hanson, G. N., 1975, Origin of Ross Island basanitoids and limitation upon the heterogeneity of mantle sources for alkali basalts and nephelinites: Contr. Mineral. Petrol. 52, 77-106.

Taylor, H. P. and Forester, R. W., 1971, Low-^{18}O igneous rocks from the intrusive complexes of Skye, Mull and Ardnamurchan, western Scotland: J. Petrology 12, 465.

Taylor, H. P. and Silver, L. T., 1978, Oxygen isotope relationships in plutonic igneous rocks of the Peninsular Ranges batholith, southern and Baja California: in Short Papers of the Fourth International Conf., Geochronology, Cosmochronology, Isotope Geology, 1978 (Ed. R. E. Zartman), Geological Survey Open-File Report 78-701, 423-426.

Thatcher, W., 1972, Surface wave propagation and source studies in the Gulf of California region: Ph.D. thesis, Cal. Inst. Tech., 147 pp.

Towell, D. G., Winchester, J. W., and Spirn, R. V., 1965, Rare-earth distributions in some rocks and associated minerals of the batholith of southern California: J. Geophys. Res. 70, 3485-3496.

Turi, B. and Taylor, H. P., 1971, An oxygen and hydrogen isotope study of a granodiorite pluton from the southern California batholith: Geochim. Cosmochim. Acta 35, 383.

Walawender, M. J., 1976, Petrology and emplacement of the Los Piños pluton, southern California: Can. Jour. Earth Sci. 13, 1288-1300.

Watson, E. B., 1976, Two-liquid partition coefficients: experimental data and geochemical implications: Contr. Mineral. Petrol. 56, 119-134.

White, A. J. R. and Chappell, B. W., 1977, Ultrametamorphism and granitoid genesis: Tectonophysics 43, 7-22.

Wildeman, T. R. and Haskin, L. A., 1974, Rare earths in Precambrian sediments: Geochim. Cosmochim. Acta 37, 419-438.

Woodford, A. O. and Harriss, T. F., 1938, Geological reconsaissance across Sierra San Pedro Martir, Baja California: Geol. Soc. Amer. Bull. 49, 1297-1336.

Wyllie, P. J., Huang, W. L., Stern, C. R., and Maalóe, S., 1976, Granitic magmas: possible and impossible sources, water contents, and crystallization sequences: Can. Jour. Earth. Sci. 13, 1007-1019.

Yajima, T., Higuchi, H., and Nagasawa, H., 1972, Variation of rare-earth concentrations in pigeonitic and hyperstenic rock series from Izu-Hakone region, Japan: Contr. Mineral. Petrol. 35, 235-244.

Yoder, H. S. Jr., 1973, Contemporaneous basaltic and rhyolitic magmas: Amer. Mineral. 58, 153-171.

Yoder, H. S. Jr. and Tilley, C. E., 1962, Origin of basalt magmas: an experimental study of natural and synthetic rock systems: J. Petrology 3, 342-532.

APPENDIX I

PETROGRAPHIC DESCRIPTIONS AND CHEMICAL ANALYSES OF SELECTED SAMPLES

Petrographic descriptions and chemical analyses of eleven selected samples representative of the range of lithologies in the Peninsular Ranges batholith are presented below. These samples are:

Sample	Rock Type	Region	
1	Tonalite	Western	Southern California
5	Tonalite	Central	Southern California
9	Quartz Gabbro	Western	Southern California
16	Calcic Gabbro	Eastern	Southern California
19	Granodiorite	Eastern	Southern California
20	Low-K_2O Granodiorite	Eastern	Southern California
23	Leucogranodiorite	Western	Southern California
29	Tonalite	Eastern	Southern California
32	Low-K_2O Granodiorite	Central	Baja California
43	Leucocratic Tonalite	Western	Baja California
50	Tonalite	Central	Baja California

Sample: 1

Rock Type: Tonalite

Location: From the tunnel dumps of the Val Verde Aquaduct, 14.5 km east of Lakeview, San Diego Co., California

General description: A medium-coarse grained biotite-hornblende tonalite. This rock is moderately foliated and carries several percent mafic inclusions. From a large intrusive mass studied in detail by Osborn (1939). A characteristic western tonalite of the Bonsall type (Larsen, 1948), which underlies large areas.

Mineralogy and Textures

50.1 Plagioclase Mostly subhedral grains up to 5 mm in an interlocking texture; seriate size distribution; complex twinning; normal and oscillatory zoning with a euhedral growth pattern; some appear slightly blurred; average interiors about An_{40}; some more calcic zones and rare cores present with a maximum difference of 20 An units, but typically core to rim zoning does not exceed 10 units; rare grains have inclusions of hornblende; slight dusting of grains, and occasional cores have sausseritic alteration.

26.3 Quartz Generally equant anhedral grains up to 3 mm; slightly undulose extinction; interstitial to plagioclase and to a lesser extent to hornblende; often irregular, slightly ragged grain boundaries with biotite.

8.9 Hornblende Subhedral and some occasional near euhedral prismatic grains up to 5 mm; pleochroic in light brown to olive green, but tinges of blue-green especially near penetrating biotite and sphene; twinning common; many grains carry a few inclusions of subhedral plagioclase; subhedral opaque also common, anhedral sphene less so; molding around plagioclase often observed.

12.4 Biotite Irregular anhedral to equant subhedral grains up to 3 mm; pleochroic in light to dark brown; molded around and occasionally slightly penetrating plagioclase and hornblende; relatively free of inclusions, sparse apatite and zircon the most common; occasional grains with partial replacement by chlorite in interlayer

		and flame structures, often accompanied by a small amount of anhedral sphene.
0.6	Alkali feldspar	Anhedral interstitial grains less than 1 mm; very slight dusting.
0.5	Epidote/Allanite	Anhedral grains usually less than 1 mm, but rarely up to 2 mm; pleochroic in pale yellow to pale green; occurs most commonly as a reaction product in and on hornblende and biotite, usually associated with an assemblage of sphene, chlorite, and to a lesser extent opaque; less often seen as a primary mineral filling interstices; also as an alteration of plagioclase; one large 2 mm grain observed to have an anhedral .2 mm patch of allanite.
<0.5	Sphene	Small anhedral grains up to .4 mm; usually part of and alteration assemblage on mafics; occasional grains are interstitial.
<0.1	Opaque	Small subhedral grains up to .2 mm; dispersed in hornblende.
0.5	Chlorite	Anhedral grains as part of an alteration assemblage on mafic minerals.
<0.3	Apatite	Euhedral prisms less than .1 mm; in mafics.
tr	Zircon	Squat prisms less than .1 mm; in mafics.
tr	Calcite/White mica	Traces in alteration of plagioclase.

Mode determined on 12 thin sections, 20,000 points, by L.T. Silver.

Sample: 5 Location: 9 km N of Aguanga, Riverside Co., California
Rock Type: Tonalite

General description: Medium to coarse grained hornblende-biotite tonalite. Evenly distributed, well-formed independent prisms of hornblende, euhedral to subhedral, and equant books of biotite, both up to 1 cm in dimension; weakly foliated. This rock is a characteristic lithologic and textural type of the central Peninsular Ranges, and forms very large plutons.

Mineralogy and Textures

60.8	Plagioclase	Mostly subhedral grains, many perfect crystal faces; seriate size distribution, .2 to 7 mm; complex twinning; normal and delicate euhedral oscillatory zoning commonly observed, mostly limited to the range of An_{40} to An_{35}, core to rim. Occasional calcic cores and zones of sodic labradorite.
21.2	Quartz	As large equant to subequant anhedral masses up to 6 mm in dimension, and as small interstitial grains; undulose extinction.
9.8	Biotite	Anhedral plates up to 5 mm, with highly irregular, characteristically ragged grain boundaries; pleochroic, pale to olive to reddish brown; interstitial to plagioclase; small amount of chlorite in interlayers; inclusions of opaque common.
7.0	Hornblende	Subhedral, stout prisms up to 5 mm; rarer anhedral grains still with prismatic form; pleochroic in pale brown to olive green with tinges of blue-green along some grain margins and some opaque and sphene grains; some larger grains are unusually poikilitic toward their interiors, anhedral quartz the most abundant inclusion but apatite and opaque are also common.
.21	Sphene	Anhedral to rare subhedral grains, up to but usually much less than 1 mm; adjacent to or as inclusions in hornblende and biotite.
.11	Alkali feldspar	Rare interstitial grains not exceeding a few tenths of a mm; clear.

.32	Opaque	Largely equant grains, anhedral to subhedral, up to .4 mm; strongly associated with biotite and hornblende.
.32	Chlorite	Fibrous masses in limited replacement of biotite and hornblende.
tr	Apatite	Small euhedral prisms mostly in biotite and hornblende.
.16	Epidote	Anhedral grains up to several tenths of mm; as both primary and secondary grains: interstitial cavity fillings and replacement of mafics, respectively.
.01	Allanite	One .5 mm grain observed; pleochroic in light brown to light reddish brown; epidote along rim, anhedral sphene traverses center of grain.
.01	Zircon	Stubby euhedral to subhedral crystals up to .2 mm, mostly in mafic minerals.

Mode determined on 1 thin section, 1886 points.

Sample: 9　　Location: In the San Marcos Mountains 7.5 km SSE of Bonsall, San Diego Co., California

Rock Type: Quartz Gabbro

General description: Medium grained, unfoliated quartz gabbro with an inequigranular, hypidiomorphic texture. The rimming of pyroxenes by hornblende and the presence of quartz and biotite are notable features. From the type locality of the San Marcos gabbro (Larsen, 1948), although no rock is typical of this heterogeneous formation.

Mineralogy and Textures

50 Plagioclase Mostly subhedral grains up to 3 mm; complexly twinned; oscillatory zoned interiors and normally zoned rims; calcic zones and cores have An_{52-68}; normal zoning of rim usually does not exceed 10 An units and total zoning is usually 15-20 An units; occasional inclusions of, and in places molded around, orthopyroxene; very slight dusting of grains, occasionally concentrated in calcic zones.

8 Clinopyroxene Anhedral and rare subhedral grains usually to 1.5 mm, rare grains to 3 mm; $2V \approx 40^\circ$ (+); often twinned; exsolution of fine grained rod-shaped opaque and possibly plates of rutile, crystallographically oriented; most grains are discontinuously rimmed, slightly penetrated and replaced by hornblende; often molded around plagioclase; rare inclusions of orthopyroxene.

3 Orthopyroxene Subhedral to anhedral grains up to 1 mm; $2V \approx 50^\circ$ (-); very slight pink-green pleochrism in some grains; rarely in granular aggregates (recrystallized?); many grains partly rimmed by hornblende, with some penetration and replacement; rare inclusions of subhedral plagioclase.

13 Hornblende Anhedral interstitial grains up to 2 mm; pleochroic in light to medium green; most grains are highly irregular and rimming pyroxenes; inclusions of subhedral plagioclase and pyroxene quite common; in some larger grains, reaction with and replacement of pyroxene is indicated.

10	Biotite	Anhedral plates up to 3 mm; pleochroic in light to dark brown; commonly but not always associated with hornblende; molded around and interstitial to most other major minerals except quartz; inclusions of plagioclase and opaque common, and also inclusions of pyroxenes, apatite, and zircon.
15	Quartz	Anhedral interstitial grains up to 1.5 mm; slightly undulose extinction.
<1	Opaque	Anhedral grains up to .3 mm; distributed in all mafic phases; in the rare granular aggregates of orthopyroxene, opaque is found in abundance as wormy to wispy grains.
tr	Zircon	Rare euhedral prisms in biotite, <.1 mm.
tr	Apatite	Rare euhedral prisms <.1 mm; as needles in plagioclase.

Approximate mode determined by visual estimation.

Sample: 16
Location: Southern part of Thomas Mountain, 5 km NE of Anza, Riverside Co., California
Rock Type: Calcic Gabbro

General description: A medium to medium-coarse grained eucritic gabbro. This rock has a well defined corona texture which always isolates anorthitic plagioclase from olivine with pyroxenes, spinel, and pale amphibolite. This rock is from one of the rare gabbroic bodies of the east side of the batholith, and is somewhat coarser than most gabbros. This body is highly heterogeneous on the scale of tens of meters, and includes anorthositic phases.

Mineralogy and Textures

55 Plagioclase Subhedral to anhedral, tabular to equant grains up to 5 mm; complexly twinned, unzoned, An_{92}; many grains have an abundance of small rod-shaped inclusions, probably an amphibole, occurring both randomly oriented and crystallographically oriented; occasionally inclusions are arranged in euhedral zones, suggesting their distribution is related to growth and possibly relict zoning; plagioclase appears to be replaced to a limited extent by amphibole along some margins as well as by the growth of rods.

5 Olivine Anhedral grains up to 2 mm; very slight reddish color; $2V>70(-)$; always found in the central parts of complex mafic aggregates which separate olivine from plagioclase.

20 Clinopyroxene Equant anhedral to subhedral grains up to 4 mm; some twinned; $2V=50-60^{\circ}(+)$; crystallographically controlled exsolution of red-brown Fe-Ti oxide (?) as small plates; often surrounded and partially replaced by orthopyroxene.

15 Orthopyroxene Highly irregular anhedral poikilitic grains up to 6 mm; pale green, slightly pleochroic; $2V\approx 90(-)$; molded around plagioclase; surrounds olivine and clinopyroxene, often invading the contacts between these two minerals, most likely at the expense of the latter; in such cases, optical continuity between clinopyroxene and orthopyroxene is sometimes observed; found in wormy intergrowths with spinel at an intermediate zone in the corona texture.

5	Pale amphibole	Irregular anhedral grains up to 3 mm; 2V large (+); probably cummingtonite; molded around and to a limited extent replacing plagioclase; always in an exterior position in the mafic aggregates of the corona texture.
2	Spinel	Small wormy grains; green, high relief (+); intergrown with orthopyroxene.
<1	Opaque	Anhedral grains up to .5 mm; found as equant grains randomly distributed among and around other mafic minerals, as fracture lining or filing in olivine, and along cleavage in pyroxenes.
tr	Phlogopite	Subhedral plates up to .5 mm; pleochroic in pale brown; only a few well formed plates observed; late growth on other mafics.

Approximate mode determined by visual estimation.

Sample: 19
Location: From road cut on SR 111, 8 km east of Palm Desert, Riverside Co., California
Rock Type: Granodiorite

General description: A medium grained hornblende-biotite granodiorite. Moderately foliated with prominent subhedral hornblende and well distributed, euhedral green sphene crystals. The essential amounts of alkali feldspar found in thin section are difficult to recognize in hand specimen, and the rock resembles a tonalite. This rock has been the subject of the detailed study of the distribution of REE among minerals.

Mineralogy and Textures

46.2 Plagioclase Subhedral grains up to 5 mm; complexly twinned; weakly zoned, gradationally from An$_{35-40}$ core to An$_{30}$, rarely An$_{25}$ rim; rare grains with antiperthitic exsolution in checkered patches restricted to part of the grain interior; myrmekitic intergrowths with quartz along most contacts with alkali feldspar; very slight dusting of some grains with fine grained alteration, occasionally along twins.

26.4 Quartz Anhedral grains, usually equant and in aggregates, up to 5 mm; undulose extinction, but not highly strained; commonly molded around or interstitial to plagioclase, hornblende, sphene, and allanite.

3.94 Hornblende Subhedral grains in prismatic form to 6 mm, rarely to 1 cm; pleochroic in light brown, olive, and blue-green; 2V 35-40(-); commonly twinned; some in aggregates with biotite, limited reaction; molded around plagioclase and sphene, and carries sparse inclusions of anhedral quartz, subhedral biotite usually near rims, and apatite; rare 1 cm grains are oikiocrystic: many inclusions of anhedral, irregular quartz and to a lesser extent sphene and biotite in the interior of grains; this quartz has inclusions of minute apatite needles; sphene and biotite may be secondary.

9.81 Biotite Subhedral to anhedral plates up to 5 mm; pleochroic in light olive brown to dark brown; apatite and zircon are occasional inclusions; rare grains have an abundance of fine apatite needles; other rare grains appear to have inherited inclusions of anhedral quartz with apatite inclusions by partial replacement of oikiocrystic hornblende.

11.9	Alkali feldspar	Equant to subequant anhedra up to 3 mm; largely clear grains, little dusting with fine alteration; incipient microcline twinning in occasional grains; modestly interstitial to most of the other major minerals, but not consistently; some grains have a small area, never on the rim, in which a small amount of a fine grained feldspar(?) has exsolved(?).
.54	Myrmekite	Fine wormy intergrowths of plagioclase and quartz often but not always developed in plagioclase where against alkali feldspar; some variation in coarseness observed.
.50	Sphene	Euhedral to subhedral grains up to 2 mm; twinning common; most are independent grains, some of which have a small number of opaque inclusions; a very small proportion of sphene is as small anhedral grains which appear to be a late reaction product on hornblende and biotite.
.41	Epidote	Mostly anhedral grains less than .5 mm; clear; found as an overgrowth on allanite and as a reaction product on hornblende.
.094	Apatite	Euhedral to subhedral prisms up to .1 mm; commonly as inclusions in biotite and hornblende.
.055	Allanite	Euhedral, independent grains ranging to several mm, but most less than 1 mm; zoned pleochroism with brownish yellow cores and yellow rims; zoning pattern is usually somewhat irregular and gradational, but is occasionally oscillatory and sharply euhedral; allanite grains are often discontinuously overgrown by clear anhedral epidote with a sharp contact between them; epidote overgrowths are always well developed against biotite, less so against hornblende, and are usually absent against quartz.
.033	Chlorite	Anhedral grains less than .1 mm; associated with epidote in the alteration of mafics.

.022	Opaque	Rare grains of hematite-ilmenite occur as inclusions in sphene.
.006	Zircon	Euhedral grains less than .1 mm, in hornblende and biotite.

Mode determined on 10 thin sections, 18,041 points.

Sample: 20

Rock Type: Low-K$_2$O Granodiorite

Location: At the easternmost foot of the Santa Rosa Mountains, 18 km south of Indio, Riverside Co., California

General description: A medium to medium-coarse grained hornblende-biotite granodiorite. Large euhedral, poikilitic alkali feldspar megacrysts up to 2 cm are prominent. A moderate foliation is defined by alignment of mafic grains and to a lesser degree by the orientation of alklai feldspar megacrysts. This rock type is exclusively eastern in character, and is recognized only in a few localities even there.

Mineralogy and Textures

45 Plagioclase Subhedral grains up to 5 mm; weak, diffuse normal and oscillatory zoning, average composition about An$_{30}$; some penetration and replacement by biotite and to a lesser extent by epidote; slight granulation along occasional grain boundaries; occasional bending of twin lamellae; rare grains show closely spaced lamellae of variable width which appear to strain related; most grains clear.

25 Quartz Anhedral grains up to several mm, and many grains in aggregates up to 7 mm; strongly undulose extinction; locally recrystallized to fine grained (a few tenths mm) polygonal aggregates; smaller grains commonly occupy interstitial positions relative to plagioclase.

15 Alkali feldspar Mostly as large grains up to 6 mm; anhedral, but most grains are near equant; incipient microcline twinning in many grains; fine grained exsolution, probably plagioclase, observed in small areas within occasional grains; inclusions are common but not abundant, and include quartz, sphene, biotite and zircon.

7 Biotite Anhedral to subhedral grains up to 3 mm; pleochroic in light to dark brown and olive green; many grains are raggy and replace plagioclase and hornblende; partial riming of plagioclase by interstitial biotite occasionally observed; some bending of plates observed.

6 Hornblende Subhedral to anhedral grains up to 4

		mm, rarely to 5 mm; pleochroic in light brown, olive green, to blue-green; larger grains are prismatic and commonly have a few small sphene inclusions; some small grains interstitial to biotite and plagioclase; occasional partial replacement by biotite and to a lesser extent by epidote.
0.5	Sphene	Euhedral to subhedral grains up to 2 mm, rare small anhedral grains; inclusions of opaque common, some of which are wormy.
tr	Epidote	Generally anhedral grains up to .5 mm, rare grains are subhedral up to 1.5 mm; found as a replacement mineral associated with biotite, and less commonly with hornblende and plagioclase.
tr	Apatite	Euhedral prisms up to .2 mm; mostly in hornblende and biotite.
tr	Opaque	Subhedral to anhedral grains up to .5 mm; commonly associated with mafic grains, particularly sphene.
tr	Zircon	Small euhedral prisms.

Approximate mode determined by visual estimation.

Sample: 23 Location: From Woodson Mountain, 7 km NE of Poway, San Diego Co., California

Rock Type: Leucogranodiorite

General description: A medium-coarse grained hornblende-biotite leucogranodiorite from the type locality of the Woodson Mountain granodiorite (Larsen, 1948). In hand specimen, large, equant quartz grains are prominent, and finer grained mafics are uniformly distributed. The low mafic and high quartz content of this rock is characteristic of this volumetrically important western rock type.

Mineralogy and Textures

42 Plagioclase Subhedral grains up to 5 mm, but most less than 3 mm; euhedral oscillatory zoning and normal zoning, cores An$_{23-25}$, rims to roughly An$_{15-20}$; some patchy zoning superimposed on many grains; strong sausseritic alteration of infrequent interiors, indicating more calcic cores and zones were originally present; otherwise, slight dusting of grains.

35 Quartz Anhedral equant grains, some up to 6 mm; slight undulatory extinction; several of the larger grains have crude but recognizable hexagonal and bipyramidal shapes.

12 Alkali feldspar Anhedral to subhedral grains up to 3 mm; incipient microcline twinning; interstitial to plagioclase and quartz, occasionally partly rimming or entirely surrounding subhedral plagioclase; grains are well dusted with fine grained alteration.

7 Biotite Anhedral grains up to 2 mm; pleochroic in light to dark brown and olive brown; often ragged, grouped in small aggregates, with limited irregular penetration of feldspar; poikilitic with abundant inclusions of small, equant, anhedral sphene grains, which are occasionally concentrated at grain boundaries; zircon, opaque and apatite are less common; in occasional grains many small overlapping polygonal-like biotite flakes occur at the margin or even in an interior, apparently late growth, replacement, or recrystallization.

3 Hornblende Subhedral, partly raggy grains up to

		1 mm; often twinned; pleochroic in light brown to green to blue-green; hornblende is locally replaced by biotite; rare interiors with abundant sphene and biotite; opaque is a common inclusion.
<1	Opaque	Subhedral grains up to .5 mm; mostly in hornblende and biotite, and often has a granular rim of sphene.
<1	Sphene	Anhedral grains less than .5 mm; found only with mafic minerals; partially surrounds some opaque and biotite grains, suggesting granular exsolution.
tr	Apatite	Euhedral to subhedral grains less than .1 mm; mostly in biotite.
tr	Zircon	Euhedral grains less than .1 mm, in biotite.

Approximate mode determined by visual estimation.

Sample: 29

Rock Type: Tonalite

Location: From In-Ko-Pah Gorge, 6 km WSW of Ocotillo, Imperial Co., California

General description: A medium-coarse grained hornblende-biotite tonalite. Seriate size distribution with most grains less than 6 mm; rare grains greater than 1 cm. An abundance of independent sphene grains is notable in this rock, and a characteristic feature of several lithologies on the east side of the batholith.

Mineralogy and Textures

61.8	Plagioclase	Subhedral to anhedral, most grains having a tabular form; seriate size distribution with most grains less than 6 mm, but rare crystals greater than 1 cm; weak normal and oscillatory zoning; cores An$_{35-40}$, rims to about An$_{30}$; zoning is euhedral, but some mottling is present due to replacement and/or partial internal reequilibration.
22.7	Quartz	Equant to subequant anhedral single grains and aggregates up to 4 mm; undulatory extinction; some grains interstitial to plagioclase.
9.5	Biotite	Anhedral plates up to 4 mm; pleochroic pale to dark brown; irregular but only occasionally ragged grain boundaries; small proportion of grains have an unusually large number of minute apatite prisms; to a minor extent, biotite replaces hornblende and penetrates plagioclase.
5.4	Hornblende	Subhedral to anhedral, up to 4 mm; pleochroic green, olive green, light brown; rarer blue-green irregularly distributed and with sphene; commonly twinned; sometimes interstitial to plagioclase; small apatite and opaque inclusions common.
.55	Sphene	Subhedral to anhedral independent grains up to 4 mm; polysynthetic twinning observed; poikilitic, partly to completely enclosing quartz and subhedral plagioclase up to .5 mm in size; overall shape usually suggestive of euhedral diamond cross section, prismatic form.

tr	Alkali feldspar	Rare small interstitial grains; clear.
tr	Opaque	Small equant grains up to a few tenths of a mm; mostly in hornblende.
tr	Zircon	Squat equant grains less than .1 mm; mostly in hornblende and biotite.
tr	Chlorite	Small anhedral grains as interlayers in biotite.
tr	Apatite	Euhedral prisms up to .2 mm; mostly as inclusions in mafic minerals.
tr	Epidote	Small anhedral grains; rims allanite and occurs as a secondary mineral on feldspar.
tr	Allanite	Subhedral-anhedral; one .5 mm grain observed; pleochroic in tan to reddish brown (α); 2V large (-).

Mode determined on 1 thin section, 1833 points.

Sample: 32
Rock Type: Low-K_2O Granodiorite

Location: On Mexico Highway 2, approximately 4 km east of Santa Isabel, Baja California del Norte, Mexico

General description: A medium grained hornblende-biotite granodiorite. Nonfoliated with a uniform texture characterized by evenly distributed, equant books of subhedral biotite and subhedral prisms of hornblende. This rock is from one of the largest plutons of the batholith, the La Posta pluton of Miller (1935), which is striking in its uniformity. It is an excellent example of the low-K_2O granodiorites of the central region of the batholith.

Mineralogy and Textures

Plagioclase Subhedral tabular grains up to 4 mm; euhedral oscillatory zoned interiors, normally zoned rims; typically cores An_{35}, rims An_{25}; occasional calcic zones and cores have moderate fine grained alteration; myrmekite developed along some borders with alkali feldspar.

Quartz Anhedral grains up to 4 mm; generally interstitial to plagioclase, with occasional inclusions of plagioclase.

Biotite Sub- to anhedral equant grains up to 3 mm; pleochroic light to dark brown; interstitial to but sometimes slightly penetrating plagioclase; altered to chlorite along occasional cleavages, sometimes accompanied by sphene, less commonly by opaque and epidote; rare grains have high concentration of small apatite grains; others have opaque exsolved (?) along cleavages.

Hornblende Subhedral grains up to 3 mm; commonly twinned; pleochroic light brown to green, but occasionally to blue-green when adjacent to sphene; rare partial replacement by chlorite.

Alkali feldspar Anhedral grains up to several mm, but forms poikilitic grains greater than 1 cm; interstitial; clear.

Sphene Anhedral grains up to .3 mm; found in hornblende and biotite, associated with chlorite.

Chlorite	Anhedral grains mostly less than .5 mm, but rare grains up to 1 mm; found along cleavage in biotite, less commonly in and about hornblende.
Myrmekite	Small wormy plagioclase-quartz intergrowths at some alkali feldspar-plagioclase contacts.
Apatite	Small euhedral prisms up to .2 mm; mostly in mafic grains.
Opaque	Small sub- to anhedral grains up to .1 mm; mostly in hornblende and biotite; also as exsolved(?) plates along cleavages in occasional biotite grains.
Epidote	Anhedral grains up to .5 mm; pleochroic from clear to pale yellow; as alteration in biotite, associated with chlorite.
Zircon	Small, rare euhedral prisms up to .1 mm.

Sample: 43

Location: 4.5 km NE of San Telmo,
Baja California del Norte,
Mexico.

Rock Type: Leucocratic Tonalite

General description: Medium grained biotite-hornblende leucocratic tonalite, fairly typical of the San Telmo pluton. Oscillatory zoned plagioclase and abundant large equant quartz characterize this rock and the pluton. Distinguished from an average leucotonalite of San Telmo by the presence of biotite and subhedral rather than interstitial hornblende, and by a higher abundance of alkali feldspar.

Mineralogy and Textures

51.0 Plagioclase Euhedral to subhedral, seriate distribution up to 4 mm; very fine oscillatory zoning of a few An units superimposed on normal zoning of An_{35} cores to An_{25} rims; this euhedral form is overgrown by anhedral to subhedral rind of An_{20} or slightly less, which constitutes 10-20% of the diameter of most grains; occasional grains have slight dusting of fine grained alteration, particularly in calcic zones; some limited replacement of plagioclase by fine grained biotite.

35.5 Quartz Unstrained equant grains up to 4 mm; mostly in polygonal aggregates to 7 mm, with an interlocking texture.

6.2 Hornblende Subhedral grains up to 1 mm; $2V \approx 70^{\circ}$ (-); pleochroic in pale brown, green, and blue-green, the latter mottled and irregularly distributed in some grains; some interior patches of biotite in occasional grains, apparently replacing hornblende; other grains in mixtures of anhedral hornblende, biotite, and opaque.

Pale amphibole Fibrous masses likely to be pseudomorphs after a prismatic mineral (pyroxene?); slightly pleochroic from clear to very pale brown and green; 2V very large; associated with opaque and minor chlorite, biotite, and hornblende, which may be replacing the pale amphibole; aggregates of anhedral hornblende, biotite, and opaques may be the result of this process carried further.

2.0	Biotite	As anhedral plates up to 2 mm, and as fine shreddy replacements of other mafics; anhedral plates are often in aggregates, which also contain opaques.
4.0	Alkali feldspar	Small anhedral grains generally not larger than a few tenths of a mm; most are interstitial, commonly along plagioclase margins; considerable dusting with fine grained alteration.
tr	Apatite	Rare small prisms in mafics.
tr	Zircon	Euhedral to subhedral prisms up to .2 mm; mostly in biotite.
tr	Opaque	Anhedral, up to 1 mm; strongly associated with other mafic minerals.

Mode determined on 1 thin section, 1600 points, by Don DePaolo.

Sample: 50

Rock Type: Tonalite

Location: 2.2 km NW of the Meling Ranch, in the western foothills of the Sierra San Pedro Martir, Baja California del Norte, Mexico.

General description: A medium grained biotite-hornblende tonalite. Weakly foliated, with rare 1 cm-sized plagioclase crystals. This sample is located within a gradational zone between the prismatic hornblende and seriate porphyritic tonalite units of the San Jose pluton (Murray, 1978). The mode and parts of this description are after Murray (1978).

Mineralogy and Textures

63.0	Plagioclase	Generally subhedral tabular to equant grains up to 4 mm; rare larger grains to approximately 1 cm; euhedral oscillatory and normal zoning, but occasionally some patchiness is developed; interiors typically An$_{36-40}$, rims to An$_{20-25}$; rare patches of sausseritic alteration, otherwise slight to moderate dusting of grains.
18.2	Quartz	Anhedral grains up to 4 mm; generally interstitial to plagioclase; larger grains usually composed of several domains; uniform or slightly undulose extinction.
8.6	Hornblende	Subhedral to anhedral grains up to 3 to 4 mm; generally prismatic; pleochroic pale tan, light olive green, and light green; occasional penetration and replacement by biotite and chlorite; hornblende commonly in aggregates of several grains, often accompanied by biotite.
5.8	Biotite	Anhedral to subhedral grains mostly 1 mm or less, but occasional grains up to 4 mm; pleochroic light brown to slightly reddish brown; poikilitic with inclusions of plagioclase, often penetrating plagioclase.
1.4	Opaque	Generally anhedral grains up to .5 mm; mostly associated with mafic minerals, but many found as scattered independent grains; sometimes poikilitic.
0.3	Alkali feldspar	Anhedral grains up to .3 mm; mostly interstitial.

0.2	Sphene	Anhedral grains up to .3 mm; found associated with opaque and with chlorite as alteration of biotite.
0.3	Epidote	Anhedral grains up to .4 mm; pleochroic yellow to clear; associated with hornblende and to a lesser extent biotite; also found replacing plagioclase.
0.6	Chlorite	Anhedral grains up to .3 mm; mostly as alteration of biotite along cleavage.
0.4	White mica	Anhedral grains up to .5 mm, mostly as alteration of plagioclase.
0.2	Apatite	Subhedral to euhedral grains up to .2 mm; commonly associated with opaque and other mafics.
tr	Zircon	Generally euhedral prisms up to .1 mm.
1.0	Clay?	Unidentified very fine grained alteration of plagioclase.

CHEMICAL ANALYSES OF PETROGRAPHICALLY DESCRIBED ROCKS

Sample	1 Tonalite	5 Tonalite	9 Quartz Gabbro	16 Calcic Gabbro	19 Granodiorite	20 Low-K_2O Granodiorite
SiO_2	63.11	62.23	58.05	45.05	66.45	63.93
TiO_2	.73	0.89	.97	.20	.65	0.80
Al_2O_3	16.50	17.87	16.77	22.10	15.89	16.46
Fe_2O_3	1.09	.70	1.36	1.05	1.08	2.12
FeO	4.28	3.86	5.83	5.74	2.67	2.57
MnO	.08	.07	.12	.12	.06	.08
MgO	2.48	1.95	4.20	8.57	1.56	1.56
CaO	5.72	6.04	7.70	14.47	4.15	4.72
Na_2O	3.42	3.95	3.19	.93	3.34	3.72
K_2O	1.69	1.36	1.04	.12	3.15	2.76
P_2O_5	.15	.20	.16	.03	0.17	0.26
H_2O^+	.81	.75	.46	1.33	.64	.64
TOTAL	100.06	99.87	99.85	99.71	99.81	99.62

CHEMICAL ANALYSES . . . (continued)

Sample	23 Leuco- Granodiorite	29 Tonalite	32 Low-K_2O Granodiorite	43 Leucocratic Tonalite	50 Tonalite
SiO_2	73.17	61.07	67.95	74.45	62.54
TiO_2	.33	.99	.55	.39	.70
Al_2O_3	13.60	17.51	15.68	13.01	17.66
Fe_2O_3	.74	1.45	0.76	1.21	2.36
FeO	1.84	3.44	2.30	1.85	2.16
MnO	.06	.05	.06	.06	.08
MgO	.57	2.16	1.20	.60	2.29
CaO	2.38	5.66	3.86	2.74	5.87
Na_2O	4.08	3.89	4.03	4.10	4.90
K_2O	2.61	1.75	2.16	1.11	.82
P_2O_5	.07	.26	.13	.05	.19
H_2O^+	.41	1.02	.75	.37	.54
TOTAL	99.86	99.25	99.43	99.94	100.11

Sources: Samples 1,9,29,50 - L.T. Silver, Unpublished Data.

Samples 5,16,19,20,23,32,43 - B.W. Chappell & L.T. Silver, Unpublished Data

APPENDIX II

ELECTRON MICROPROBE ANALYSES OF SELECTED MINERALS IN A GRANODIORITE NEAR INDIO, CALIFORNIA

The following analyses were determined on a MAC-5-SA3 electron microprobe interfaced to a PDP-8/L computer for control and online data processing. Oxide weight percents were calculated by the procedure of Bence and Albee (1968), using modified correction factors from Albee and Ray (1970). The analytical procedure followed is discussed in detail by Chodos, Albee, Gancarz, and Laird (1973). For determinations of some of the REE, peak and background locations provided by Ms. Mary Johnson were used.

Analyses for plagioclase, alkali feldspar, hornblende, biotite, allanite, epidote, and sphene are presented. Core and rim compositions of one grain each of allanite and sphene give an indication of the range of REE and Th concentrations in these two minerals.

REFERENCES

Albee, A.L., and Ray. L., 1970, Correction factors for electron probe microanalysis of silicates, oxides, carbonates, phosphates and sulphates: Anal. Chem. 42, 1408-1415.

Bence, A.E., and Albee, A.L., 1968, Empirical correction factors for the electron microanalysis of silicates and oxides: J. Geol. 76, 382-403.

Chodos, A.A., Albee, A.L., Gancarz, A.J., and Laird, J., 1973, Optimization of computer-controlled quantitative analysis of minerals: Proc. 8th Nat. Conf. on Electron Probe Analyses, New Orleans, La.

MINERAL	PLAGIOCLASE	ALKALI FELDSPAR	HORNBLENDE	BIOTITE
WEIGHT %				
SiO_2	58.95	65.22	42.10	36.91
TiO_2	0.00	0.03	1.37	2.62
Al_2O_3	25.11	18.50	11.66	15.35
FeO_t	0.13	0.05	20.39	20.41
MnO	---	---	0.42	0.25
MgO	0.00	0.00	8.12	10.45
CaO	8.45	0.00	11.96	0.00
Na_2O	7.16	1.28	1.56	0.09
K_2O	0.26	13.44	1.55	9.35
BaO	0.00	0.58	---	---
F	---	---	0.12	0.21
Cl	---	---	0.13	0.06
TOTAL	100.06	99.10	99.30	95.60
FORMULA PROPORTIONS				
# Molecules	4	4	15	7
Si	2.659	2.996	6.365	2.874
Ti			0.156	0.154
Al	1.336	1.002	2.079	1.409
Fe	0.005	0.002	2.579	1.329
Mn			0.053	0.017
Mg			1.830	1.213
Ca	0.409		1.938	
Na	0.627	0.114	0.457	0.013
K	0.014	0.778	0.299	0.928
Ba		0.010		
F			0.057	0.052
Cl			0.034	0.008
Ab	59.7	12.7		
An	38.9	0.0		
Or	1.4	87.3		
Excess Oxygen(%)			0.0	4.30
Positive Charge			45.88	22.40
Al IV			1.635	1.126
Al VI			0.444	0.282
Mg/(Mg+Fe)			0.415	0.477

MINERAL	ALLANITE		EPIDOTE	SPHENE	
WEIGHT %	RIM	CORE		RIM	CORE
SiO_2	31.34	31.37	37.84	29.92	29.82
TiO_2	0.35*	0.34*	0.08	36.51*	36.48*
Al_2O_3	15.10	15.18	22.06	1.62	1.44
FeO_t	12.29	14.36	13.02	0.98*	0.98*
MnO	0.20*	0.20*	0.19	0.17*	0.17*
MgO	0.71	1.22	0.03	0.00	0.01
CaO	13.27	11.93	23.70	27.81	27.31
La_2O_3	4.65	6.10	0.00	---	---
Ce_2O_3	7.51	10.15	0.03	0.27	0.48
Nd_2O_3	1.82	2.64	0.00	---	---
Dy_2O_3	---	---	---	0.15	0.14
Y_2O_3	0.04	0.02	0.00	0.08	0.16
ThO_2	3.57	1.64	0.00	0.02	0.05
TOTAL	90.83	95.16	96.94	97.51	97.03

FORMULA PROPORTIONS

# Molecules	8	8	8	3	3
Si	3.094	3.002	3.015	0.994	0.999
Ti	0.026	0.025	0.005	0.913	0.919
Al	1.757	1.713	2.072	0.063	0.057
Fe	1.014	1.150	0.868	0.027	0.027
Mn	0.016	0.016	0.013	0.005	0.005
Mg	0.104	0.174	0.003	0.000	0.000
Ca	1.404	1.223	2.024	0.990	0.980
La	0.169	0.215			
Ce	0.271	0.356	0.001	0.003	0.006
Nd	0.064	0.090			
Dy				0.002	0.002
Y	0.002	0.001		0.001	0.003
Th	0.078	0.035		0.000	0.000

*These values not determined on the same grain.

APPENDIX III

PURITY OF MINERAL SEPARATES

The purity of the mineral separates were checked by visual examination of grain mounts in Lakeside or Aerochlor cement. Each mount was point counted, and the results reported below.

Plagioclase	1440 grains counted; all of these were either plagioclase or quartz, but 40 grains carried small inclusions of apatite 4 grains had inclusions of hornblende or biotite 2 grains had inclusions of hematite (?) Determination of the proportion of quartz at 27% was by relative intensities of X-ray diffraction peaks, calibrated by making mixtures of the sample with pure quartz. (Note: This material was not analyzed directly. This material was ground finely and acid washed to remove apatite.)
Alkali feldspar	1091 grains counted; of these 2 grains were bimineralic with biotite 1 grain was biotite 1 grain was quartz
Fine biotite	2130 grains counted; of these 30 grains were hornblende 3 grains were a compacted aggregate 1 grain was epidote
Hornblende	558 grains counted; of these 1 grain was bimineralic with epidote 1 grain was a fibrous mineral (chlorite?) Several hornblende grains had apatite inclusions.
Apatite	1014 grains counted; of these 4 grains were sphene

	85 zircon inclusions were noted. Since apatite was dissolved in dilute HNO_3, these impurities should not have contributed to the analysis.
Epidote	548 grains counted; of these 8 grains were bimineralic with hornblende 3 grains were bimineralic with biotite 1 grain was allanite 1 grain had a small allanite core
Sphene	2319 grains counted; of these 1 grain was a compacted aggregate Several dozen of the sphene grains had small opaque inclusions.

Grain mounts for coarse biotite, allanite, and zircon were not prepared. The coarse biotite material was not appropriate for making such a mount, the allanite was hand picked so purity was assured, and the zircon was used as received from L.T. Silver. Petrographic observations indicate that the more common inclusion in biotite is apatite.

APPENDIX IV

ANALYTICAL PROCEDURES

PREPARATION OF STANDARD AND SPIKE SOLUTIONS

Standard and spike materials for 8 REE were obtained in oxide form from Johnson Matthey Chemicals Limited and Oak Ridge National Laboratory, respectively. The preparation of solutions of accurately known concentrations from REE oxide powders by gravimetry is difficult due to the absorption of volatiles from air, particularly for several of the light REE oxides. Careful work by Masuda et al. (1972) has shown that ignition of the oxides can remove nearly if not all of the volatiles (mostly H_2O). In addition, ignition should correct any nonstoichiometry in the oxides as received from the suppliers. Duval (1963) lists the thermal stability limits of REE compounds as determined by thermogravimetric analysis. Above 820°C, all REE dealt with in this investigation are present as sesqueoxides with the exception of Ce, which is as the dioxide.

Each oxide was placed in a nested quartz beaker assembly that had previously been baked out to 820°C, to constant weight. This assembly was placed in a furnace and heated to 820°C, held at this temperature for 2 days, cooled to 400°C over 5-6 hours in the furnace, then transferred to a ported dessicator containing P_2O_5. During the initial 10 minutes, the dessicator was vented often to relieve excess pressure. When fully cool, the dessicator was opened (now under significant vacuum) and the beaker with oxide quickly weighed. Nd_2O_3 was the only oxide observed to gain weight over the duration of the weighing process, so its weight was recorded as a function of time and extrapolated to the time of exposure to air. This heating procedure was repeated with heating periods of 6 hours or more until a constant weight was obtained. Nd_2O_3 and Sm_2O_3 lost the most weight of all the oxides, 14% and 13%, respectively. The CeO_2 standard lost less than 0.1%.

All oxides except CeO_2 were dissolved in a very small amount of concentrated HNO_3 with heat, then diluted with 1N HNO_3. CeO_2 resisted several attempts to dissolve it in hot concentrated HNO_3, even when H_2O_2 was used as a reducing agent. A H_2SO_4-HBr dissolution procedure was used as described by Potts, Early, and Herrmann (1973). This attack produces $Ce_2(SO_4)_3$, which is soluble in H_2O and dilute acids.

The solutions were prepared both gravimetrically and volumetrically. Precision in gravimetry is better than 0.1%, as is the precision of volumetric flasks scribed and calibrated in this laboratory. Except for the Ce solutions, the density of the solutions are known by measurement and by calculation from the components of the solution. Agreement is within 0.1% in all cases.

The resulting solutions were mixed and diluted to form working solutions and stored in polypropylene bottles.

Considering the errors associated with gravimetric and volumetric measurements, stated purity of the oxides, non-stoichiometry, and various laboratory procedures, calibration of the solutions are believed accurate to within ±1%. This is supported by a comparison of standard concentrations determined by gravimetric-volumetric techniques and by isotope dilution using the prepared spike solutions (Table A).

LABORATORY PROCEDURES

Sample Dissolution

200 to 300 mg of finely powdered sample is weighed out in a 30 ml Teflon (TFE) beaker, wetted to eliminate problems with static, and spiked. After evaporation to bring the liquid fraction to about 1 ml, 5 ml of concentrated HF is added while swirling the contents of the beaker. Digestion is allowed to proceed without heating for 6 hours or overnight. Roughly 0.5 ml 70% $HClO_4$ is added and evaporated to dryness at a low heat. The residue is dissolved in a few ml of 2N HNO_3 and allowed to sit for a few hours. This solution is centrifuged to separate a small amount of white residue (a Ti compound?).

Ion Exchange Column Procedure for REE Separation

A two column technique supplied by Dr. J.G. Arth (personal communication) with some modification was used for the separation of REE. After centrifuging, the supernate is pipeted onto a 20 X 1 cm column of AG50W-8X 100-200 mesh resin equilibrated to 2N HNO_3, and eluted as follows:

TABLE A

COMPARISON OF GRAVIMETRIC-VOLUMETRIC AND ISOTOPE DILUTION CALIBRATIONS

ISOTOPE	GRAVIMETRIC-VOLUMETRIC	ISOTOPE DILUTION	DIFFERENCE
	Micromoles/10^{-3}ml		
^{142}Ce	4.2838E-4	4.2472E-4	-0.85%
^{145}Nd	4.2022E-4	4.2041E-4	+0.05%
^{149}Sm	9.5023E-5	9.5353E-5	+0.35%
^{153}Eu	7.8422E-5	7.8089E-5	-0.42%
^{155}Gd	8.2050E-5	8.1880E-5	-0.21%
^{161}Dy	1.0436E-4	1.0500E-4	+0.61%
^{167}Er	5.5869E-5	5.5483E-5	-0.69%
^{171}Yb	6.7148E-5	6.6298E-5	-1.27%

5 ml	2N HNO_3	column wash (discard)
5 ml	2N HNO_3	column wash (discard)
130 ml	2N HNO_3	most major elements (discard)
10 ml	6N HNO_3	
25 ml	6N HNO_3	collect for REE (yield ≃ 90%)

This column separates the REE as a group from most of the sample. A few drops of 70% $HClO_4$ are added to the REE fraction and it is evaporated to dryness. The residue is dissolved in a few ml of 3N HCl and loaded on a 19 X 0.5 cm column of AG50W-X8 100-200 mesh resin equilibrated to 3N HCl and eluted as follows:

1 ml	3N HCl	column wash (discard)
1 ml	3N HCl	column wash (discard)
9.5 ml	3N HCl	column wash (discard)
9.5 ml	3N HCl	collect heavy REE fraction
17.5 ml	3N HCl	collect light REE fraction (rarely run)
5.0 ml	3N HCl	collect Ce fraction

This procedure does not produce sharp separations among adjoining REE, nor is it desired that it do so. The heavy REE fraction contains sufficient quantities of the light REE so that they may be measured during the same run but not so abundant as to interfere with measurement of the heavy REE. Therefore, the light REE fraction is rarely run. However, when analyzing samples with light REE enriched patterns, it is necessary to adjust the cuts on the column to insure against excessive light REE in the heavy REE fraction. Typically, the volume of the heavy REE cut is adjusted to 9.0 or 8.5 ml and the light REE cut increased accordingly. A few drops of 70% $HClO_4$ are added to the fractions and they are evaporated to dryness.

Mass Spectrometry

All analyses were performed on a 12 inch radius of curvature solid source mass spectrometer with magnetic peak switching and online data processing. Ion currents were collected in a Faraday cup. All the REE were run as metal species. No corrections for mass fractionation were made.

The heavy REE fraction is picked up with dilute HNO_3 and dried onto the side filaments of a Re triple filament (1x30 mil Re ribbon). With the center filament at 2000°C, the side filament currents are raised slowly. Typically in the range from 1.0 to 2.5 amperes, the elements usually appear in the following sequence: Eu-Yb-Dy-Sm-Nd-Er-Gd. Signals are stable even when sustaining rapid growth or decline. In most cases, signal strength changed only moderately with time during the period data collection was appropriate.

Measured pairs for the isotope dilution calculations have been chosen to avoid interferences. It is not possible to completely eliminate mass interferences caused by oxide species of lighter REE overlying metal species of heavier REE, and caution must be exercised during a run to ascertain that such interferences are negligible. The most serious problem is with ^{161}Dy and ^{157}Gd, which have mass overlaps with $^{145}Nd^{16}O$

and $^{141}Pr^{16}O$, respectively. In most runs these oxide species are present, and run conditions must be adjusted to insure that they are insignificant. Center filament temperatures of 2000°C cause most of the REE to have metal/oxide ratios of 100 or greater. Increasing the filament temperature to 2050°C or higher results in further suppression of oxides. Observations in this laboratory indicate that when the metal species of the offending oxide does not exceed the intensity of the metal which is to be measured, the temperatures above are sufficient to limit interferences to less than 1%. This is insignificant for this study if the spike to sample ratio is within a commonly achieved favorable range.

The Ce fraction is loaded, often on the same filament used for heavy REE. With the center filament at 1950° to 2000°C and side filaments at 1.0 to 1.2 amperes, Ce is measured after ascertaining that ^{142}Nd is not contributing to the ^{142}Ce peak.

The data obtained are usually precise to 0.2% or better, and the concentrations calculated are believed to be accurate to 2%, considering all sources of error in the chemical procedure and mass spectrometry.

BLANK DETERMINATIONS

The blank levels for REE during this study have been low, and in no rock or mineral analysis is a blank correction warranted. In the worst cases, the plagioclase determinations, sample/blank ratios always exceed 50; in whole rock analyses, values well in excess of 100 are the rule.

Three blank determinations, two near the beginning of this study and one near the end, are presented below. The first two determinations suffer from very unfavorable spike/blank ratios in the middle to heavy REE, and conservative upper limits are given. The comparatively high blank levels for Ce and Nd are fairly well determined. The high Ce values may have resulted from general laboratory contamination during the repeated efforts to dissolve the CeO_2 standard powder; a similar explanation can be offered for Nd. The last blank determination was made with more favorable spike/blank ratios, and it indicates lower blank levels for all REE.

TABLE B

BLANK DETERMINATIONS IN NANOGRAMS

	11/75	2/76	7/78
Ce	3.5	4.7	0.8
Nd	4.5	4.8	0.5
Sm	---	0.25	0.04
Eu	---	0.1	0.006
Gd	<0.4	<0.3	---
Dy	<0.1	<0.2	0.05
Er	<0.2	<0.1	0.05
Yb	<0.04	<0.2	0.05

APPENDIX V

LISTING OF ROCKS ANALYZED BY SAMPLE NUMBER, FIELD NUMBER, AND ROCK TYPE

Sample Number	Field Number	Rock Type
1	SCB-19	TONALITE
2	SCB-101	TONALITE
3	SCB-10	LEUCOGRANODIORITE
4	SCB-801	TONALITE
5	HEM-12	TONALITE
6	VLL-13	LEUCOGRANODIORITE
7	PEC-33	HORNBLENDE GABBRO
8	SCB-14	PYROXENE GABBRO
9	SCB-32D	QUARTZ GABBRO
10	SCB-702	LOW-K_2O LEUCOGRANODIORITE
11	RRM-21	TONALITE
12	PMO-27	QUARTZ GABBRO
13	PMO-23	LOW-K_2O GRANODIORITE
14	SCB-316	TONALITE
15	SCB-923	TONALITE
16	IDY-30	CALCIC GABBRO
17	IDY-18	LOW-K_2O GRANODIORITE
18	SCB-403	GRANODIORITE
19	SCB-925	GRANODIORITE
20	VAL-16	LOW-K_2O GRANODIORITE
21	SCB-803	TONALITE
22	SCB-313	TONALITE
23	SCB-300	LEUCOGRANODIORITE
24	SCB-602	QUARTZ GABBRO
25	SCB-402	LEUCOGRANODIORITE
26	SCB-401	TONALITE
27	Los Pinos #Ia	HORNBLENDE GABBRO
28	SCB-932	LOW-K_2O GRANODIORITE
29	SCB-606	TONALITE

APPENDIX V, Continued

Sample Number	Field Number	Rock Type
30	BC-I-950	LEUCOGRANODIORITE
31	BC-I-714	TONALITE
32	BC-I-410	LOW-K_2O GRANODIORITE
33	BC-I-713	HORNBLENDE GABBRO
34	BC-I-986	LOW-K_2O GRANODIORITE
35	BC-I-411	TONALITE
36	BC-I-711	GRANODIORITE
37	BC-I-969	TONALITE GNEISS
38	BC-I-968	ADAMELLITE
39	BC-I-238	LEUCOGRANODIORITE
40	BC-I-19	TONALITE
41	BC-I-808	LEUCOGRANODIORITE
42	BC-I-230	LEUCOCRATIC TONALITE
43	BC-I-800	LEUCOCRATIC TONALITE
44	BC-I-852	LEUCOCRATIC TONALITE
45	BC-I-858	LEUCOCRATIC TONALITE
46	BC-I-859	LEUCOCRATIC TONALITE
47	BC-I-900	QUARTZ GABBRO
48	BC-I-851	LOW-K_2O GRANODIORITE
49	BC-I-908B	PYROXENE GABBRO
50	BC-I-12	TONALITE
51	BC-I-402	TONALITE
52	BC-I-505	TONALITE
53	BC-I-501	TONALITE
54	BC-I-503	GRANODIORITE