THE PETROLOGY AND EMPLACEMENT OF THE MORENA RESERVOIR PLUTON, SAN DIEGO COUNTY, CALIFORNIA

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"Plutonic holism" - a view proposed here that a pluton or plutonic complex has a tectonic reality independent of and greater than the sum of its rock units and internal structure.

Morena Reservoir, "Heart" of the Morena Reservoir pluton.

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CHAPTER I

INTRODUCTION

Regional Setting

The Morena Reservoir pluton is a part of the Mesozoic Peninsular Ranges batholith of western North America. The Peninsular Ranges batholith (PRB) extends from south of the Transverse Ranges in southern California (34° N) to the southern end of Baja California Norte, Mexico (28° N) (Figure 1). The batholith, which is roughly 1000 km long and about 70 to 140 km wide, is a deeply eroded segment of the Cordilleran magmatic arc that was active along the margin of western North America during the middle to late Mesozoic. The batholith comprises several hundred 1 and 50 km-wide plutons along with a wide range of smaller dikes emplaced within variably metamorphosed sedimentary and volcanic country rocks (Larsen, 1948; Gastil and others, 1975; Gastil and others, 1983; Todd and Shaw, 1979; Todd and others, 1988, Walawender and others, 1991). The Peninsular Ranges batholith appears to have been emplaced near the edge of the North American craton as at least two roughly continuous sequences of magmatic activity (G. Girty and D. Kimbrough, oral communications, 1994).

Regionally consistent longitudinal regularities and systematic transverse variations between the western and eastern parts of the Cretaceous batholith (Figure 1) have been used to define distinct plutonic belts (Gastil, 1975; Silver and Chappell, 1988).

Figure 1. Locality Map of the Field Area and schematic chemical variations across the Peninsular Ranges Batholith (from D.B. Clarke, 1992).

These zones trend north-northwest and are characterized by various criteria that include pluton composition, size, age, structure, and depth of emplacement, prebatholithic rock type and metamorphic grade, style of deformation, isotopic and REE signatures, magnetic susceptibility, mineralization, gravity and magnetics profiles, and Pn and Pg velocities (Gastil and others, 1975; Silver and others, 1979, Silver and Chappell, 1988; Todd and Shaw, 1979; Todd and others, 1988; DePaolo, 1981; Oliver, 1982; Erskine, 1986; Hearn and Clayton, 1986; Grommet, 1987; Gastil and others, 1990; Walawender and others, 1990; and Grove, M., 1994;). Several of the above parameters are geographically illustrated on page 4, Figure 2. A summary of the major aspects of the Peninsular Ranges batholith is found in Table 1 on page 5.

Pre- and Synbatholithic Rocks and Regional Metamorphism

The host rocks of the western Peninsular Ranges batholith are largely volcanic-volcaniclastic rocks of locally greenschist facies that make up the Early Cretaceous Santiago Peak Volcanics (Larsen, 1948). Kimbrough and Herzig (oral communication, 1994) consider the Santiago Peak Volcanics to be the coeval volcanic edifice of the associated intrusives of the western magmatic arc. The gently-dipping and undeformed Santiago

Figure 2. Regional Geologic Boundaries of the Northern Peninsular Ranges Batholith (after Todd and others, 1990)

Table 1.

Summary of Information for the Cretaceous Peninsular Ranges Batholith (after Clarke, 1992)

Geologic setting:

Number of intrusions: hundreds.
Sizes of intrusives: diameters 1–50 km.
Nature of intrusives: multi-pluton complexes (west); large, single, zoned plutons (east).
Country rocks: related metavolcanics(west), metasedimentary(east).
Regional metamorphism: sub-greenschist facies (west);upper amphibolite facies (east).
Relationship to orogeny: syntectonic and post-tectonic.

Tectonic setting: subduction-related fringing/continental magmatic arc.

Petrology:

Principal granitoid lithologies: hornblende-biotite tonalite and low-K granodiorite.
Granitoid abundances: metaluminous>>peraluminous (peralkaline = 0); range of A/CNK 0.8–1.6 (most plutons 0.9–1.2).
Associated rocks: gabbro, quartz gabbro, quartz diorite, granite, leucogranite, two-mica granite (latter associated only with pelitic roof pendants, and edge of craton).
Absolute ages: west = 140–105 Ma (static); central = 150–190 Ma (older arc); east 105–80 Ma (transgressive eastward).
Opaque mineral variations: dominantly magnetite (west); minor ilmenite, no magnetite (east.

Geochemistry:

Chemical variations: genetically important systematic chemical variations from west to east, as follows: generally increasing A/CNK, Sr, REE's abundance, $(La/Yb)_N$, Eu/Eu* (0.8–1.0); decreasing transition elements.
Isotope geochemistry: genetically important systematic chemical variations from west to east, as follows: increasing delta^{18}O (+6–+20 per mil), $^{87}Sr/^{86}Sr_i$ (0.7030–0.7154), $^{207}Pb/^{204}Pb$ (18.71–18.87 (exceptionally 19.5)); decreasing (epsilon)Nd (+8–-6).

Economic geology: Au, Sn, Ni and granite pegmatite gems.

Peak Volcanics nonconformably overlie Tithonian-age, argillaceous volcanigenic metasedimentary strata that are overturned, steeply-dipping, with well-developed slaty cleavage (C.T. Herzig, oral communication, 1994).In the central part of the batholith across southern California, the host rocks are metamorphosed flysch-type clastic rocks whose protoliths included greywacke, arkosic sandstone, and argillite. These rocks are termed the Julian Schist (Hudson, 1922) and were metamorphosed to amphibolite-grade metamorphism. The prebatholithic rocks in the eastern part of the batholith are miogeoclinal in character and the protolith included quartz arenite, shale and limestone (Gastil and Miller, 1984). Some stratigraphic interfingering between the arc clastic rocks in the west and the continental apron clastic rocks in the central zone has been reported (Todd and Shaw, 1979; and Todd and others, 1988). Todd and Shaw (1979) draw a prebatholithic boundary denoting the transition from volcaniclastic to flysch-type country rocks, (Figure 2). This transition may be due mostly to the amount of uplift and depth of erosion of the batholith, considering that Jurassic to Cretaceous volcanigenic strata crop out in the east across Sonora, Mexico (G. Gastil, written communication, 1994).

The grade of regional metamorphism increases eastward reaching the highest intensity of metamorphism near the central-eastern zone boundary that coincides with the eastern physiographic escarpment (at the latitude of San Diego) (Todd and Shaw, 1979; and Grove, M., 1994). Here an eastward step in increasing metamorphic grade and decreasing pluton cooling ages suggests a major juxtaposition of hotter, deeper rocks in the east against shallower rocks on the west along mostly unrecognized faults near the batholith's eastern escarpment (Grove, 1994). However, in the far eastern part of the southern segment of the batholith (seen along the eastern side of the Baja peninsula and the western part of Sonora, Mexico), the metamorphic grade drops to greenschist facies (Gastil and others, 1975).

Systematically deeper levels of the batholith are exposed eastward from San Diego. This is evidenced by the occurrence of coeval assemblages of hypabyssal plutonic and volcanic rocks in the west (epizonal), the increase in size and zonation of plutons to the east (mesozonal), the eastward increase in metamorphic grade, and the increasing divergence of hornblende and biotite K-Ar cooling ages to the east (Krummenacher and others, 1973; Walawender and others, 1990; Grove, 1994). The zoning of increasing metamorphic grade and decreasing intrusive cooling ages appear to be "stepped" eastward and separated by mostly unrecognized shear or fault zones that parallel the long axis of the batholith (Grove, M.; Morton, D. and Todd, V., personal communications, 1994).

Transverse Variations in the Peninsular Ranges Batholith

Plutonic Variations

The plutons in the western zone are typically small (<15 km in diameter), irregularly shaped and reversely-zoned, and tend to form roughly concentric, nested, structurally concordant, multiple-pluton complexes (Gastil and others, 1991; Todd and others, 1994). Most of these rocks are magnetite-series (after Ishihara, 1974). The western plutons can be described as broadly variable in composition, shape, foliation, and distribution patterns of various rock types. The most abundant rock type in the western zone is tonalite, with lesser amounts of granodiorite. Gabbroic plutons and dark enclave-rich plutons are common in the western zone; true granites are uncommon.

In the central to eastern zones, plutons are larger (up to 50 km in diameter), display well developed compositional zonation with gradational internal contacts. Emplacement depths for these plutons are estimated to be 11 to 15 km by Grove (1994) and 9 to 11 km by Clinkenbeard (1987). The plutons in the central zone of the batholith include relatively old, small, elongate, heavily contaminated, and foliated Late Jurassic Harper Creek and Cuyamaca Reservoir type plutons (Todd and Shaw, 1985; Girty and others, 1994; and Thomson and others, 1994). These Late Jurassic granitoids are termed S-types by Todd and Shaw (1985), or heavily crustal-contaminated I-types by Thompson and others (1994).

In contrast to the plutons of the western zone, the eastern zone plutons have less lithologic diversity with a larger number of granodiorite plutons that are comagmatically associated with, but dominated by, tonalite. Gabbroic plutons are much less common in the eastern zone of the batholith compared to the western zone, and are very rare in the far eastern part of the batholith; monzogranites are more numerous. The mid-Cretaceous granitoids of La Posta-type are metaluminous to peraluminous that have been slightly contaminated by earlier crustal rocks. The rocks of the central and eastern zones are virtually all ilmenite-series (after Ishihara, 1974).

Based on the distribution of gabbroic rocks across the batholith, Gastil (1975) divides the batholith along a "gabbro line", (Figure 2). Todd and others (1988) note that S-type granitoids are restricted to the eastern side of the batholith and they define an I-S line,(Figure 1).

A narrow, transitional zone where western- and eastern- type plutons overlap, straddles the boundary between the western and central zones of the batholith; this coincides with the irregular boundary between older western-type and younger eastern-type magmatism. The eastern edge of the western zone is marked by both a tectonic solid-state foliation and the Cuyamaca-Laguna Mountain Shear Zone (Todd and Shaw, 1979; M. Girty and others, 1994; and Thompson and others, 1994).

The Cuyamaca-Laguna Mountain Shear Zone is a northwest-striking, steeply northeast-dipping, polygenetic structure that is characterized by SC mylonites (G. Girty and others, 1994). This shear zone is located near the eastern boundary of the western magmatic zone. Most of the deformation is in the central zone Late Jurassic metagranitoids (Figure 2). Early Cretaceous contraction occurred along this zone before and during 118 to 114 Ma. Normal-sense shearing overprints the contractional fabric, and cross-cutting relations suggest that active down-to-the-east extension along the normal-sense shear zone post-dates 105 Ma (age of the shear-deformed Las Blancas pluton) and pre-dates 94 Ma (age of the undeformed and shear zone-truncating La Posta pluton; Thomson and others, 1994; and G. Girty and others, 1994). Contractional shearing can be tied to a major surge in voluminous western-type magmatism and the onset of the proposed Cretaceous mid-proto-Pacific superplume. The extensional shearing can be correlated to the onset of eastern, La Posta-type magmatism when strong dextral-sense oblique subduction occurred between the North American and

Farallon plates.

Local Structure and Rock-type Classification

An extensive, detailed field mapping transect across the batholith at the latitude of San Diego, California suggests that many of the western plutons are arranged into much larger, structurally concentric/concordant, multi-pluton complexes that typically include rocks from gabbro to leucomonzogranite (Todd and others, 1988; Todd, oral communication, 1994). Each complex appears to be a variation of a single general plutonic-tectonic theme that typically includes several generally definable patterns in lithology, foliation, structure, and relative timing of emplacement of the plutonic rock types. The Morena Reservoir pluton is a part of one of these complexes, informally named here by the author the Hauser Complex. The numerous plutonic facies that make up the batholith, as defined by Todd and others (1988), e.g., Alpine (K_{a}), Granite Mountain (K_{gm}), Chiquito Peak (K_{cp}), Pine Valley (K_{pv}), Corte Madera (K_{cm}), La Posta (K_{lp}), Las Bancas (K_{lb}), Japatul Valley (K_{jv}), Cuyamaca (K_{c}), Mother Grundy Peak (K_{mgp}), etc., have a broad range of magmatic compositions, textures, and structures that probably originated from a number of different source regions and emplacement conditions.

Zircon Ages

The batholithic rocks have been divided into an older syn-tectonic intrusive sequence in the western zone, and a younger, late- to post-tectonic sequence in the eastern zone (Silver and others, 1979; Silver and others, 1988; Todd and Shaw, 1979). Uranium-lead zircon ages of plutonic rocks range from about 140 to 105 Ma in the western zone.

Uranium-lead zircon ages of the Santiago Peak Volcanics range from 130 to 120 Ma (Herzig and Kimbrough, 1994; Kimbrough and Herzig, 1994).

The eastern half of the Peninsular Ranges batholith has yielded zircon ages ranging from 98 to 80 Ma, (Silver and others, 1979). The youngest ages are found in garnet-bearing leucogranodiorites that intrude the La Posta pluton (Walawender and others, 1990). Zircon uranium-lead ages in the highly foliated, elongate plutons in the axial part of the batholith are reported to be as old as Late Jurassic with ages between 140 and 190 Ma (M. Girty, 1994). A transverse profile of pluton ages led Silver and others (1979) to conclude that the western zone represents a static magmatic arc and the eastern zone a migrating arc. However, Walawender and others (1990) suggest that the western and eastern zones are parts of a single arc that was affected by a major change in subduction dynamics around 105 Ma.

Isotopic and Rare-Earth-Element Profiles

Regional isotopic and rare-earth-element (REE) data also have regular spatial patterns that define the west to east zonations in the northern 600 km of the batholith (Taylor and Silver, 1978; Silver and others, 1979, Silver and others, 1988; Gromet, 1987; DePaolo, 1981). Silver and Chappell (1988) observed that the batholithic zones delineated by various geochemical parameters appear to be almost independent of pluton lithology. These relationships are probably related to both the nature of the source regions from which the various magmas of the batholith were derived and the associated wall rocks through which the magmas moved.

Initial strontium ratios are low in the western part of the

Peninsular Ranges batholith (0.701 to 0.704) and increase systematically eastward to as high as 0.713 (Morton, written communication, 1994) along the batholith's eastern margin, with a value of 0.704 dividing the western and eastern zones (Silver and others, 1975). These trends in initial strontium are regional generalizations do not reflect variations in specific plutons.

Oxygen isotope data ($\delta^{18}O$) show a similar systematic increase from west to east with a pronounced steep gradient or "step" between the western zone (6 to 8 per mil) and eastern zone (9 to 12 per mil), (Figure 2; Silver and others, 1979). Rocks from an east-west transect from the eastern edge of the La Posta pluton to the eastern edge of the Morena Reservoir pluton were analyzed for oxygen isotope systematics by Walawender and others (1990). They report a value of 8.0 mil from the Morena Reservoir pluton sample, and a value of 10.2 mil from the adjacent hornblende-biotite facies of the La Posta pluton. This reflects the $\delta_{18}O$ step illustrated in Figure 2.

A single profile of neodymium isotopic compositions across the batholith in southern California has a systematic west to east range that correlates with strontium and oxygen isotopic variations (De Paolo, 1981).

The REE patterns vary systematically across the Peninsular Ranges batholith, and can be divided into three parallel longitudinal zones (Gromet and Silver, 1987). Variation across each zone is largely independent of rock type, and a sharp break in REE pattern occurs between the western and central zones. Tonalite and granodiorite rocks in the western part of the batholith have REE profiles that are flat through the middle to heavy REE's and negative-sloped in the light REE's. These profiles are slightly concave-up with negative (tonalite and granodiorite) europium anomalies. The gabbroic rocks of the western zone have roughly flat concave-down REE profiles with positive europium anomalies.

The tonalite and granodiorite rocks of the central and eastern zones have negative, steeply sloping REE patterns with little to no europium anomalies. A gradational eastward transition towards high LREE enrichment occurs from the central to eastern zones. Gromet and Silver (1987) attribute these changing REE patterns to eastward deepening of the source region with associated phase changes that define gabbroic to amphibolitic sources in the west, with a transition to eclogitic sources in the central to eastern provinces.

Magnetic Susceptibility

The Peninsular Ranges batholith is zoned transversely in magnetite content. Most western-zone plutons are magnetite-series rocks). The eastern-zone plutons are virtually magnetite-free ilmenite-series rocks. Based upon both aeromagnetic surveys and ground-based magnetic susceptibility measurements along the axis of the batholith, a regional magnetite boundary has been drawn (Figure 2; Erskine, 1986; and Gastil and others, 1990). This boundary generally coincides with the eastern edge of the western zone and with prebatholithic, lithologic/ magmatic, isotopic, and REE pattern boundaries. Straddling this boundary are several zoned plutons that contain both magnetite- and ilmenite-bearing rocks (Gastil, 1990; Rector, 1988).

Geophysical Features

First order Bouguer gravity anomaly patterns of the Peninsular

Ranges batholith have a minor positive anomaly for the western half, and a large negative anomaly for the eastern half of the batholith (Oliver, 1982). The eastern anomaly is interpreted as reflecting the batholithic root, whereas the western anomaly is considered to represent a shallow mafic crustal floor with the absence of a root.

Heat flow of the western side of the batholith is unusually low, whereas, the eastern portion has higher, more normal continental heat flow values (Hearn and Clayton, 1986).

The geophysical data suggests a fundamental difference in crustal character between the western and eastern halves of the batholith. The above coincidental boundaries along the central axis of the batholith have prompted workers to suggest a suture zone between two arc terranes, one marginal oceanic, the other continental margin (Gastil, 1975; Todd and Shaw, 1979). The Morena Reservoir pluton is located along the proposed suture zone. However, convincing evidence now supports a single evolving arc model, and that the coincidental boundaries simply mark the change from oceanic to continental crust (Wooden and Stacy, 1987; Ague and Brimhall, 1988; Walawender and others, 1990). The Morena Reservoir pluton's composition, texture, structure, and spacial and temporal position define it as a transitional-type pluton, i.e., magmatism emplaced into the western zone, but characterized by a combination of western- and eastern-style properties.

Western Zone Profile of Magmatism

As described above, the western zone of the Peninsular Ranges batholith contains a 140 to 105 Ma syntectonic plutonic sequence of magnetite-bearing, epizonal to mesozonal, metaluminous, rocks ranging in composition from peridotite to granite. Larsen (1948) and other workers developed a simple model for the petrogenetic evolution of the western Peninsular Ranges batholith. They hypothesized that the various granitoid types form a time-related, spatially static magmatic differentiation sequence derived from a single primary melt. In this sequence, gabbro is the oldest plutonic rock, followed in order by tonalite, granodiorite, granite, and finally pegmatite. This model was supported by local cross-cutting field relationships. However, recent field studies and new isotopic age and cooling data indicate that the model is far too simple - tonalites locally cut granodiorites and monzogranites; some gabbros are relatively young; and there is a problem of differentiating between nested groups of small individual plutons versus large comagmatically zoned plutonic complexes (V.R. Todd, oral communication, 1990; D.L. Kimbrough, oral communication, 1993).

Field relationships in the western Peninsular Ranges batholith can be difficult to interpret (V.R. Todd, oral communication, 1990). For example, many contacts between granitoid bodies show no reliable cross-cutting features or contact effects. In many places in the western Peninsular Ranges batholith, the contacts between different granitoid types are gradational, suggesting comagmatic zonation of a pluton, or equilibration between adjacent part liquid plutons.

A post-magmatic solid-state deformation overprint on the eastern edge of the western zone of the Peninsular Ranges batholith appears to be temporally related to the onset of La Posta-type magmatism and coeval ductile shearing along the Cuyamaca-Laguna Mountain shear zone

(Todd and Shaw, 1979; G.H. Girty, oral communication, 1993). Several "satellitic" La Posta-like plutons intrude the eastern part of the western zone (V.R. Todd, oral communication, 1990). These zoned "satellitic" plutons have characteristics of both the western- and La Posta-type plutons, and appear to represent transitional-type plutons in the Peninsular Ranges batholith.

CHAPTER II

PROJECT PURPOSE, DESIGN, AND GOALS

Purpose

The purpose of this investigation is to answer the following questions: 1) are the compositionally zoned rocks of the Morena Reservoir pluton comagmatic, or do they represent two or more noncomagmatic "nested" tonalite bodies?; 2) based on the field relations, petrography, and geochemistry, what were the magmatic processes and emplacement mechanisms that took place in the growing pluton?; 3) what was the source region or regions for these rocks?; 4) what is the crystallization age of the pluton?; 5) what is the cooling history of the pluton?; 6) at what depth was the pluton emplaced?; and 7) do the rocks record syn- and post-crystallization tectonics?

The 200 km^2 study area (area within thesis geology map) is located in the northern Peninsular Ranges batholith at the boundary between the western and eastern zones near the western edge of the La Posta pluton (Figures 1-3).

The Morena Reservoir pluton was mapped and sampled in detail with approximately 5 localities per km^2 (Plate I). A series of samples were selected for transects across the pluton for petrographic and major, trace and REE geochemical analyses. Zircon U-Pb geochronology was determined for one sample, and hornblende Ar-Ar geochronology on determined for two samples.

Figure 3. Location Map of the Morena Reservoir Pluton(after Walawender and others, 1990).

Geographic Data

The 65 km^2 Morena Reservoir pluton is located along the central axis of the northern Peninsular Ranges batholith in southern California, forty five miles east of San Diego near the United States-Mexico border (Figures 1-3). The study area is located between the northern latitudes of 32°42' and 32°35', and between the western longitudes of 116°40' and 116°30'. Geographic features that surround the pluton include Los Pinos Peak on the northwest, the Narrows on the northeast, the towns of Cameron Corners and Campo on the east, the U.S.-Mexico border on the south, Morena Butte on the west. The east-west trending Hauser Canyon lineament dissects the southern part of the pluton. Morena Reservoir occupies the central part of the pluton. The Morena Reservoir pluton has a total topographic exposure of about two thousand feet.

The terrain consists of wide, flat, grassy alluvial valleys situated between brush-covered rugged highlands. The vegetation is dominantly chaparral, with scrub oaks outlining the deeper drainage patterns. Most of the pluton is within parts of the Cleveland National Forest, Hauser Wilderness Area and BLM lands. The pluton can be reached by four paved roads and scores of dirt roads.

The town of Morena Village, located near the center of the Morena Reservoir pluton, can be reached from San Diego, by travelling east for 50 miles on Highway 94 to Cameron Corners, left on Buckman Springs Road (1 mile north), and finally right on Morena Road (2.5 miles north); or by taking Interstate 8 east for 35 miles to Buckman Springs Road, right on Buckman Springs Road (9 miles south), to the Morena Village turnoff, and finally, right on Morena Road (west 1 mile).

The climate of the study area is mostly dry and sunny with cold nights and warm to hot days, punctuated by intermittent winter cyclones or summer thunderstorms.

Field Methods and Data Sets

Field work on the Morena Reservoir pluton was carried out between the fall of 1990 and the winter of 1993. The method of field mapping included an outcrop-to-outcrop examination of the pluton along many traverses. Data stations were chosen on outcrops that afforded the best exposure and rock freshness for the immediate area. Data station density for the 200-km^2 field area averages about four stations per km^2, 5 stations per km^2 for the Morena Reservoir pluton. The data collected at each station include the following: 1) mineral modes; 2) mineral size and habit; 3) strike and dip of foliation plus the character and estimated intensity of foliation; 4) general structural elements that include schlieren, contacts, and shear zones; 5) magnetic susceptibility of the host rock, mafic enclaves and dikes, including the range and average; 6) character, size range and average, morphology range and average (long axes versus short axes ratios), and abundance distribution of mafic enclaves; 7) character and orientation of dikes; and 8) collection of samples from roughly one out of every five stationed outcrops. Outcrop to outcrop examination was carried out between data stations to ensure station-to-station fabric continuity, especially for magnetic susceptibility. Fresh 3 to 6 kg samples from roughly evenly-spaced, km-wide distances across the pluton.

Field data were collected from 260 data points within the 65 km^2 Morena Reservoir pluton, and 50 rock samples were collected for laboratory work (Plate I). Neighboring plutons and prebatholithic wall rocks and their contacts with the Morena Reservoir pluton were also examined in this study.

The IUGS rock classification scheme is used for naming the plutonic rocks (Streckeisen, 1975)

CHAPTER III

ROCKS OF THE MORENA RESERVOIR PLUTON

Outline of the Pluton's Geology

Data for mineralogy, texture, foliation, structure, magnetic susceptibility, dikes, and mafic enclave abundance, size, and morphology were collected from two hundred and sixty localities within the Morena Reservoir pluton (Plate I). Samples from forty localities were stained, fifteen samples were thin-sectioned and polished, eleven samples were analyzed for major and trace geochemistry, one sample was processed for U-Pb zircon geochronology, and two samples were processed for hornblende and biotite Ar-Ar geochronology. Two hundred localities were also mapped within the surrounding plutonic wall rocks.

The 65-km^2 Morena Reservoir tonalite pluton has an irregular "Y" shaped outline and is elongated along a northwest trend (Figures 4 and 5, Plate II). The northern part of the pluton branches into two distinct northeast and northwest appendages. The pluton's internal lithologies and structure are generally concentric to its geographic center and are generally elongate northwest-southeast.

Figure 4. Geology map of the Morena Reservoir pluton. Rock units are in Figure 5 on following page. Full-size geology map is in Plate II located in back pocket.

Figure 5. Rock units of geology map illustrated in Figure 4 on preceding page.

The Morena Reservoir pluton is moderately zoned from mafic biotite-hornblende quartz diorite along its margins adjoining the Los Pinos and Morena Reservoir gabbro rocks, grading to more siliceous biotite-hornblende tonalite elsewhere, particularly adjacent to granodiorite plutons and metasedimentary wall rocks. A pod-like body of hornblende gabbronorite (K_{lop}) is located in the pluton's core (Figures 4 and 5, Plate II).

Several structural discontinuities in the Morena Reservoir pluton occur around parts of the core region and along the Hauser Canyon lineament. These boundaries are defined by abrupt changes in lithology, texture, foliation, and mafic-enclave character, and they suggest magmatic and tectonic dislocations.

Roughly concentric color index zones are generally concordant to the pluton's margins and centered about the its core (Plate III). Color indices grades from 16 to 27. Variation of the color index appears to be locally related to the adjoining country rock lithology; highest values occur adjacent to gabbroic rocks and lowest values are in rocks adjoining granodiorite and metasedimentary rocks. Color index patterns appear to be truncated and offset left-laterally along the Hauser Canyon lineament.

The rocks are medium-grained, subhedral equigranular to subhedral seriate with small variation in grain size and habit. The rocks are characterized by medium- to coarse-grained, euhedral to subhedral plagioclase and hornblende, medium- to coarse-grained, euhedral to anhedral biotite, medium-grained, anhedral quartz, and fine-grained, interstitial potassium feldspar.

Plate III. Color Index Map of the Morena Reservoir Pluton.

The Morena Reservoir pluton is zoned in ground-measured magnetic susceptibility. Volume-percent magnetite distribution is inferred from the magnetic susceptibility data (Plate IV). The magnetic susceptibility patterns are roughly concentric to the margins and center of the pluton. Relatively high susceptibility (abundant magnetite) occurs along the western and northeastern margins of the pluton. Rocks with low magnetic susceptibility (low magnetite content) occur across most of the pluton, particularly in the central and southern parts. Broad zones of high, moderate, or low magnetic susceptibility values are separated by narrow zones of abrupt changes. The most notable changes from high to low magnetic susceptibility in the Morena Reservoir pluton are found along the eastern side of the pluton. This boundary corresponds to the regional magnetite line of Gastil and others (1989).

Mafic enclaves occur throughout the pluton and are zoned in abundance, morphology, size, and spatial distribution. Mafic enclave abundance and morphology patterns are roughly concentric to the margins and center of the pluton (Plates V and VI). Mafic enclaves decrease in abundance from the pluton's margins to the core,

Plate IV. Magnetic Susceptibility Map of the Morenab Reservoir Pluton.

Plate V. Mafic-Enclave Distribution Map of the Morena Reservoir Pluton

Plate VI. Mafic-Enclave Morphology Map of the Morena Reservoir Pluton

and are typically flattened with "discus-like" to "pancake-like" forms. Flattening is greatest near the pluton's margins and systematically lessens towards the core.

The pluton has a pervasive but variable, magmatic foliation that is generally northwest-striking and steeply-dipping (Plate VII). The foliation is defined by the preferred orientation of magmatic plagioclase, hornblende, and biotite, and to a lesser degree, flattened quartz. Foliation is concordant to the pluton's external and internal margins, but is locally deflected around the contact with the pluton's gabbroic core body where the foliation dips steeply inward around the gabbroic core. Based upon foliation attitude, the pluton consists of a southwest zone of inward-dipping, moderately steep northeast dips, and a northeast zone of steep, inward-dipping, southwest dips. A roughly linear northwest-southeast inflection boundary separates these zones along the northern edge of Morena Reservoir. Dips along this boundary are vertical or subvertical.

The foliation is best-developed near the pluton's margins and least-developed in the core. A crenulated foliation defined by folding of the primary foliation, occurs in parts of the pluton's margin, particularly adjoining the gabbro and granodiorite bodies, and along several internal, curvilinear structural boundaries. Lineations are rare and only occur along the pluton's boundary with the granodiorite of Corte Madera. Flattened mafic enclaves are subparallel to the foliation.

Plate VII. Foliation Map of the Morena Reservoir Pluton.

Schlieren are scarce and mainly occur at, or near, the pluton's margin, particularly adjacent to the Long Potrero pluton. They exhibit centimeter- to meter-scale layering that closely conform to the general foliation pattern. Local textural variations occur where schlieren and/or mafic enclaves are abundant, and the intensity of foliation is moderate to intense.

Lithology patterns, texture, foliation, magnetic susceptibility, and mafic enclave abundance and morphology variations form distinct patterns that are generally concordant to the margins and are elongated along a northwest trend. Abrupt pattern changes define structural boundaries that divide the pluton into at least two pulse units.

The structural discontinuity along the Hauser Canyon lineament appears to show oblique left-lateral displacement (Figures 4 and 5; Plate II). This is based on map pattern offsets, and offset fabric markers noted in outcrops of a N70W trending shear zone in the bottom of Hauser Canyon. The maximum age of displacement postdates the emplacement of the Morena Reservoir pluton's southern part.

Internal Lithologic Variations

The Morena Reservoir pluton consists of a relatively narrow range of medium-grained, subhedral equigranular to seriate, plagioclase-rich, potassium feldspar-poor rocks that grade from biotite-hornblende quartz diorite to biotite-hornblende tonalite (Figures 6-11). Fresh rock is white with black speckling of hornblende and biotite (Figures 6, 8, and 10). Textural variation in the rocks across the pluton is mostly due to the variation in abundance, size and habit of the hornblende and biotite, and to the foliation intensity. Hornblende and biotite reach sizes of up to 2 cm in length, with biotite occurring either as discrete thin tablets, or as larger, irregular, multi-grain clusters.

Sixteen samples were slabbed, stained, with a minimum of 700 points counted on each. Polished thin sections were also made from the same samples for petrographic analyses, in part to determine the accessory minerals.

Petrography

Introduction

Polished thin sections from eighteen rock samples that represent both the compositional and geographical zones of the pluton were studied using flat stage petrographic techniques (see Plate I for sample localities). Thin sections were studied for mineralogy, phase relations, deformation, and alteration.

Figure 6. Unstained slabs from the western part of the Morena Reservoir pluton.

Figure 7. Stained slabs from the western part of the Morena Reservoir pluton.

Figure 8. Unstained slabs from the central part of the Morena Reservoir pluton.

Figure 9. Stained slabs from the central part of the Morena Reservoir pluton.

Figure 10. Unstained slabs from the eastern part of the Morena Reservoir pluton.

Figure 11. Stained slabs from the eastern part of the Morena Reservoir pluton.

Figure 12. Sample localities of Morena Reservoir pluton rock slabs illustrated in figures 6-11.

Rock fabric

The texture and grain size the Morena Reservoir tonalite varies only slightly across most of the pluton. The tonalite is medium-grained (0.5 to 4 mm), equigranular to seriate with significant strain features noted in the phenocrysts. Brittle and crystal-plastic stain features include undulatory extinction and subgrain boundary development in the quartz, undulatory extinction and warped cleavage planes in biotite, patchy albite twinning and bent and fractured plagioclase laths, and fractured hornblende prisms. The fabric consists of a plagioclase-quartz-biotite-hornblende groundmass (0.25 to 2.5 mm) with medium- to coarse-grained (3 to 8 mm), euhedral to subhedral plagioclase and hornblende phenocrysts, and medium- to coarse-grained (2 to 6 mm), subhedral to anhedral biotite and quartz phenocrysts (Figure 13). Accessory minerals consist of fine-grained (0.05 to 0.2 mm) opaque minerals, apatite, sphene, zircon, and monazite, typically as inclusions in the major minerals, except for sphene and monazite. Chlorite, epidote (clinozoisite), hematite, and sericite are also identified.

Variably developed microscopic foliation is defined by the preferred orientations of tabular, phenocrystic plagioclase and hornblende, and elongate megacrystic quartz. All studied samples show crystal-plastic strain in the quartz and biotite, and brittle deformation in the plagioclase, biotite, and hornblende; the intensity of the strain features vary with locality. The groundmass crystal matrices appear to have a dominantly igneous texture as evidenced by irregular, granular quartz and albite-twinned plagioclase.

Figure 13. Photomicrograph of the Morena Reservoir tonalite depicting typical microscopic texture (2.5 mm-wide image).

The abundant, yet nonpervasive, ductile and brittle strain features seen in the phenocrysts, and the low amounts of strain features in the groundmass suggests that these textures developed during submagmatic conditions. The foliation and deformation microfabrics of the Morena Reservoir pluton are described and analyzed in detail in Chapter IV.

Mineralogy and Phase Relations

Major modal analyses were made from stained slab samples. Accessory phase trace modes were made from thin section, and integrated into the slab modes (Appendix A). Descriptions of each mineral are presented below.

Plagioclase. Albite-twinned plagioclase constitutes 50% of the slide area, and appears to be the earliest-formed major mineral. Plagioclase has the best developed grain boundaries, and it is the most abundant phenocryst phase. Moderate to intense oscillatory zoning in the interiors of most plagioclase crystals indicates a complex crystallization history (Figure 14).

Plagioclase grains range in size from 0.5 to 7 mm. Synneusis textures are common with grain mosaics up to 1 cm in diameter, and consist of 4 to 8 subgrains (Figure 15). Plagioclase phenocrysts up to 12 mm in length are seen in outcrop. Crystals range from euhedral (25%? of grains) to subhedral (65%?) to anhedral (10%?), with equant to tabular prismatic form. Habit characteristics are generally similar for both phenocrysts and groundmass grains. Roughly 70% to 90% of the plagioclase crystals are albite-twinned, and about 50% have variably developed oscillatory zoning. Oscillatory zoning is better developed in the rocks of the western and southern parts of the pluton.

Plagioclase crystals are euhedral where adjoined by potassium feldspar, and less so by quartz, biotite, and hornblende, particularly hornblende. Subhedral crystals are typically surrounded by hornblende, quartz, in the other plagioclase, and to a lesser degree, by biotite. Xenomorphic plagioclase crystals are mostly found surrounded by hornblende and other plagioclase. In several of the thin sections, a significant number of plagioclase phenocrysts are recessed and filled with coarse-grained, "ragged-edged", xenomorphic biotite platelets; these plagioclase laths have otherwise idiomorphic grain boundaries with quartz and hornblende and zoned interiors (Figure 16). This phenomena is noted across most of the pluton except for the northeastern part.

Figure 14. Photomicrograph of oscillatory-zoned plagioclase in the Morena Reservoir tonalite (2.5 mm-wide image).

Figure 15. Photomicrograph of synneusis texture of plagioclase in the Morena Reservoir tonalite (2.5 mm-wide image).

Figure 16. Photomicrograph of anomalous plagioclase-biotite textural relations in the Morena Reservoir tonalite (2.5 mm-wide image).

Primary inclusions in plagioclase includes euhedral hornblende (in only two samples) and opaque minerals. Clinozoisite occurs as a common alteration of plagioclase. Mottling of the cores of strongly zoned plagioclase phenocrysts is common, yet the percentage of mottled laths varies from thin section to thin section. Deuteric alteration of plagioclase is greatest in the west-central and northwestern parts of the pluton and least developed near the pluton's core.

The majority of the plagioclase crystals, particularly the phenocrysts, appear to be aligned (Figure 17). The degree of alignment varies, and reflects the intensity of the field-measured macroscopic foliation. A more detailed analysis of the microfoliation character of the plagioclase, and the other major minerals, is found in Chapter IV.

A small percentage of the medium to large phenocrysts have patchy and noncontinuous, twinning and oscillatory zoning, and "warped" and/or fractured crystal laths (Figure 18). Small zones of mermikitic intergrowths of quartz and albite form between some plagioclase crystals and interstitial potassium feldspar. The intensity of strained plagioclase is greatest in the marginal parts of the pluton with most intense deformation in the eastern region and particularly in the northeastern part.

Figure 17. Photomicrograph of preferred orientations of plagioclase, hornblende and biotite in the Morena Reservoir tonalite (2.5 mm-wide image).

Figure 18. Photomicrograph of strained plagioclase and quartz in the Morena Reservoir tonalite. (2.5 mm-wide image).

Hornblende. Olive green hornblende is ubiquitous and has a highly variable habit in terms grain boundary development, from euhedral prismatic to anhedral "swiss cheese"-textured (Figure 19).

Next to plagioclase, hornblende is the second most abundant phenocryst, and it also occurs in the groundmass. Hornblende occurs as euhedral to anhedral, equant to elongate (width:length=1:4), prismatic crystals that range in size from 0.25 mm to 5 mm. The two-dimensional shapes of the grains have the typical amphibole diamond-shape. Phenocrystic hornblende is mostly subhedral and rarely euhedral. Groundmass hornblende crystals are typically anhedral and rarely subhedral. About half the hornblende crystals have simply twinning. Pleochroism ranges from greenish straw yellow to bluish forest green. The negative 2V ranges from 65° to 75°, with 2°

yellowish olive green birefringence. There is little or no "wrap-around" mantling of hornblende by biotite. Distinct crystals of biotite and hornblende are typically found side-by-side within mafic mineral-rich zones (Figure 19).

Primary inclusions within hornblende crystals, particularly in the phenocrysts, include euhedral plagioclase, apatite and monazite; and euhedral to subhedral opaque minerals, biotite, and sphene.

Alteration and replacement of hornblende appears to be of two fundamental types. These types are considered to have occurred at different stages of the rocks history: 1) quartz and biotite (magmatic?/submagmatic?); and 2) chlorite and clinozoisite (post-magmatic? deuteric-style). In all samples, rims and cores of hornblende are variously resorbed, particularly in the phenocrysts; quartz occurs in the resorption sites. Hornblende crystals with the most resorbed features are found around the core region and in the northeastern part of the pluton. The least resorbed hornblendes occur in the western and southern parts of the pluton. However, the resorption character of the hornblende is considerably heterogeneous in all samples, with some crystals largely resorbed and others largely pristine and untouched. The resorbed regions of the rims and interiors have rather smooth, curved/rounded boundaries, particularly where quartz filled in the zone of resorption. Resorbed hornblende grain boundaries produce a anhedral crystals, but with a "ghost" euhedral form. Adjacent grains such appear to dominate the grain boundary.

The resorbed nature of hornblende makes the interpretation of the crystallization sequence more tenuous. However, in rocks where resorption is minor, hornblende phenocrysts have euhedral grain boundaries adjacent to some plagioclase and all quartz, biotite, and potassium feldspar. Hypidiomorphic hornblende phenocrysts occur where they are surrounded by plagioclase. Primary hypidiomorphic to xenomorphic hornblende crystals occur as a groundmass phase and may be surrounded by plagioclase, quartz, and biotite.

Elongate hornblende prisms display various degrees of preferred alignment. This alignment is considered to be part of the magmatic foliation also exhibited by elongate crystals of plagioclase and quartz, and somewhat by biotite (Figure 19). The degree of preferred alignment varies from thin section to thin section and appears to be directly related to the measured intensity of the foliation measured in the field.

Other types of hornblende alteration include deuteric alteration of hornblende to chlorite and clinozoisite. The habit of the secondary chlorite or clinozoisite is anhedral and irregular.

The variable, resorbed textures of hornblende indicate a complex crystallization history; these textures record significant changes in crystal-melt conditions such as pressure, temperature, and oxygen, water, and silica fugacities.

Figure 19. Photomicrograph of hornblende in the Morena Reservoir tonalite (2.5 mm-wide image).

Biotite. Ubiquitous chestnut brown biotite appears to be a late magmatic phase, similar to that of quartz. Biotite occurs both as phenocrysts and in the groundmass. Pleochroism ranges from straw-yellow to greenish dark chestnut brown. Biotite negative 2V range from roughly 5° to 15° with low 2° birefringence colors.

Biotite is 0.5 to 5 mm in size with individual grains subhedral to anhedral (Figure 20). Biotite also forms multi-grain, "patchwork" clusters made up of numerous, fine-grained, euhedral to subhedral crystals up to 10 mm sizes. Grain clusters up to 20 mm long were observed in outcrop. Clustered biotite habit is best developed in the external margins of the pluton and around the gabbroic core, typically where mafic content, hornblende resorption, and strain features are most intense.

Inclusion-free, anhedral biotite platelets have anomalous, ragged, grain boundaries that appear to recess into otherwise idiomorphic plagioclase and hornblende phenocrysts (Figure 16). This characteristic is noted in rocks over most of the pluton, with the best development found across the pluton's central zone, but it appears to be absent in the northeastern part of the pluton where macroscopic deformation is most intense.

Minor amounts of primary inclusions in biotite include opaque minerals, apatite, and zircon. Biotite is the most frequent host to minute zircon, and the second most frequent host to opaque minerals, a distant second to hornblende. In all rock samples, some biotite is partly altered to chlorite. In a few rock samples, epidote is an alteration product of biotite. Chlorite alteration is most pervasive in rocks from the northwestern part of the pluton, and least in the central core area.

Figure 20. Photomicrograph of biotite in the Morena Reservoir tonalite (2.5 mm-wide image).

Quartz. Strained, granular, anhedral, inclusion-free quartz crystals are generally equigranular to elongate (width:length up to 1:3) (Figure 21). Quartz dominantly occurs as discrete grains and rarely as an interstitial phase. Quartz typically ranges in size from 0.25 to 4 mm with some 4 to 8 mm-size megacrysts. Quartz megacryst abundance varies proportionally to the quartz content. Rocks with quartz megacrysts are located in the northwestern, northeastern, and eastern parts of the pluton, and along its margins where it contacts tonalite or granodiorite.

Quartz crystal shapes are typically controlled by plagioclase and hornblende grain boundaries. Quartz grain boundaries typically dominate adjacent biotite grain boundaries (Figure 21). These interphase relations are evidenced by the abundance of idiomorphic plagioclase and hornblende grain corners, i.e., "salients", into otherwise
equidimensional granular quartz grains.

Crystal-plastic deformation includes undulatory extinction and development of low- to high-angle subgrain boundaries (Figure 22). The intensity of crystal strain appears to be proportional to the intensity of the foliation measured in the field. Typically, elongate quartz grains typically are aligned concordant with oriented plagioclase, hornblende, and biotite. Slight to moderate quartz recrystallization fabrics are characterized by 0.5 to 2 mm-size band- or blob-like assemblages of fine-grained, generally polygonal, undeformed quartz grains. These fabrics are most common in the pluton's margins.

Figure 21. Photomicrograph of quartz grains in the Morena Reservoir tonalite (2.5 mm-wide image).

Figure 22. Photomicrograph of strained quartz in the Morena Reservoir tonalite (2.5 mm-wide image).

Potassium feldspar. Little to no potassium feldspar occurs in the Morena Reservoir rocks. It is invariably interstitial between plagioclase grains or between plagioclase and hornblende crystals (Figure 23). Orthoclase is the dominant potassium feldspar and ranges in size from 0.1 to 3 mm with larger grains having poikilitic textures. Stained slabs show larger potassium feldspar oikocrysts up to 10 mm in diameter. Interstitial zones of mermikite are surrounded by potassium feldspar, plagioclase, and quartz (Figure 23). Potassium feldspar is most abundant in rocks from the central portion of the pluton south of the gabbroic core unit.

Figure 23. Photomicrograph of interstitial potassium feldspar and mermekite in the Morena Reservoir tonalite (2.5 mm-wide image).

Accessory minerals. Fine-grained, euhedral to anhedral opaque minerals (dominantly ilmenite, lesser magnetite, and scarce hematite) are characteristically found within or adjacent to hornblende phenocrysts, and to a much lesser

degree in biotite (Figure 24). Opaque mineral abundance varies proportionally with the hornblende content. Typically, equant opaque minerals range in grain size from 0.05 to 0.5 mm. Opaque minerals in polished thin sections consist of crystals with exsolved textures that give them a sort of "swiss cheese" appearance, i.e., translucent dark gray with small, irregular, jet black blobs and strings (Figure 24). The highly variable crystal shapes of opaque minerals range from equigranular cubic, to anhedral/needle-like, even within one hornblende crystal. The widely variable habits of the opaque mineral grains suggest that the opaque minerals either grew (exsolved from, or replaced, resorbed hornblende) throughout much of hornblende's crystallization history, or they all crystallized early and were variably resorbed during the growth or resorption of hornblende.

Opaque mineral abundance across most of the pluton is relatively low (less then 0.05 volume%). A few rock samples have significantly more opaque minerals (up to 0.5 volume%). These rocks are typically the most mafic-rich and magnetically susceptible, and are exclusively located close to the gabbroic plutons.

Opaque minerals with irregular shapes are typically associated with anomalous mafic mineral clusters and hornblende resorbed textures (Figure 24). Here, opaque minerals have partial or complete sphene rims, along with locally abundant fine-grained apatite and quartz grains. Many of these sphene-rimmed opaque mineral crystals also have highly elongate habits that are aligned with the host hornblende or biotite cleavage planes.

Figure 24. Photomicrograph of opaque minerals in the Morena Reservoir tonalite (2.5 mm-wide image).

Trace amounts of euhedral to subhedral, equant triangular-, parallelogram-, or elongate "coffin"-shaped, prismatic sphene are most commonly adjacent to or within hornblende, and less commonly between biotite and plagioclase grains (Figure 25). Crystals of amber-colored primary sphene range in size from 0.2 to 3 mm. The abundance of sphene appears to vary systematically in the pluton, with highest amounts noted in the more marginal parts of the pluton and the smallest amounts in the more mafic-rich rocks surrounding the gabbroic core body. Sphene also forms partial or complete rims around opaque mineral grains in rocks containing anomalous mafic-rich aggregates of fine-grained hornblende and biotite. The habit of sphene suggests that it nucleated and grew throughout much of the rock's crystallization history, well into the later stages of solidification, much like the opaque minerals.

Figure 25. Photomicrograph of primary sphene in the Morena Reservoir tonalite (2.5 mm-wide image).

Apatite occurs in relatively trace amounts as rounded, equant to elongate, euhedral prisms in all rocks examined. Apatite occurs as inclusions predominantly in plagioclase, hornblende and biotite, and also between these phases. Apatite crystals range in size from 0.02 to 0.3 mm. Anomalously abundant numbers of apatite crystals are found around concentrations of opaque crystals in hornblende.

Euhedral zircon prisms occur within biotite, and less commonly within hornblende. Most zircon crystals are bipyramidal, equant to elongate prisms and range in size from 0.02 to 0.1 mm. Crystals are clear with no observed coloration in the cores.

Equant, euhedral to subhedral monazite prisms were found in all but one of the thin sections. Monazite occurs adjacent to the edges of biotite and hornblende crystals. Equant monazite crystals range in size from 0.05 0.2 mm with one highly unusual euhedral crystal measuring 1 mm. Some monazite crystals contain minute zircon grains.

Secondary minerals. Secondary minerals are found in all thin sections. Secondary minerals include quartz, biotite, epidote (clinozoisite), chlorite, sphene, hematite, and sericite. The type and abundance of the secondary minerals vary both, across the pluton as seen in the different thin sections, and within a single thin section. Two types of alteration are identified and probably are either late-magmatic (quartz, biotite, sphene, and hematite), or as post-solidus deuteric alteration (chlorite, epidote, and sericite). Replacement by epidote
of plagioclase and hornblende, and chlorite of biotite and hornblende, are most common; replacement by sphene of hornblende and biotite, quartz and biotite of hornblende, and hematite of dark opaque minerals, are less common; and alteration of plagioclase to sericite is rare.

Irregular masses and rarely equant subhedral to anhedral grains of epidote have partially replaced rims and interiors of plagioclase, biotite and resorbed regions of hornblende. Trace amounts of chlorite and sphene have
replaced parts of biotite and hornblende grains, and chlorite occurs as pseudomorphs after biotite. Very fine-grained anhedral crystal aggregates of sphene occur between biotite crystals and within resorbed parts of hornblende prisms. Quartz and lesser biotite is

found within resorbed parts of hornblende crystals. Translucent red hematite? is found as separate, minute grains within hornblende, or as replaced? parts of opaque grains.

Discussion

Defining and using textures to interpret crystal-lization histories and crystal-melt processes is an important part in modeling a plutonic rock's petrogenesis. However, recent experimental research on magma-like crystallizing systems indicate that classic magmatic textures that have been confidently assigned to specific crystal-melt processes or relations may need to be interpreted more cautiously (Means and Park, 1994). Means and Park (1994) have shown that textural metamorphism may begin in magmatic and submagmatic systems.

These rocks are considered to have magmatic textures, yet there are several types of strain fabrics that could be described as submagmatic to high temperature subsolidus. Defining magmatic, submagmatic, and subsolidus (foliation) textures and their genetic processes has been attempted with some success (Paterson and others, 1989). A general crystallization sequence is interpreted from the above described mineral textures, based upon the degree of grain boundary development. The crystallization sequence of the major minerals of the Morena Reservoir rocks is plagio-clase/hornblende - quartz/biotite - potassium feldspar. Plagioclase and hornblende dominated the early stages of crystallization, whereas quartz and biotite crystallized in the later stages. The accessory minerals zircon, apatite, opaque minerals and monazite formed early in the crystallizing magma. Sphene appears to have formed in a latter stage of rock formation, after plagioclase and hornblende but before quartz and biotite.

Resorption textures of hornblende and possibly opaque and sphene minerals, oscillatory zoning in plagioclase, and mermikite textures indicate disequilibrium crystallization in the tonalite of the Morena Reservoir pluton. Disequilibrium crystallization is likely attributed to both open-system crystal-melt magmatic processes and critical timing of temperature-pressure-rheology changes of the pluton during emplacement and crystallization.

Hornblende and the associated opaque and sphene minerals, appear to have experienced crystal-melt disequilibrium during the crystallization of these rocks, based on variable resorption/crystallization textures in the hornblende and its opaque and sphene inclusions. This is particularly evident in the more mafic-rich rocks where large, anomalous aggregates of fine-grained hornblende crystals (and possibly the opaque inclusions) were resorbed, and followed by crystallization of opaque minerals, quartz, and biotite, +/- sphene in the zones of resorption.

Plagioclase also shows multiple episodes of crystal-melt disequilibrium throughout much of the crystals' growth history based on oscillatory zoning in most of the grain interiors. Embayments of the rims of some plagioclase laths and the associated anomalous relation of anhedral , "ragged-edged" biotite recessing into otherwise euhedral plagioclase phenocrysts indicates a late-magmatic? crystal-melt disequilibrium.

The above inferred crystal-melt disequilibriums were initiated and sustained by some combination of complexly changing melt geochemistry

and one or more fundamental physical parameters such as critical changes in pressure, temperature, and rheology related to magma displacement and crystallization of the Morena Reservoir rocks. Ascent of the magma during crystallization necessitates a systematically decreasing pressure-temperature regime.
Changing pressure or water fugacity during isothermal conditions in the magma can move the crystal-melt system out of the hornblende stability field.

The oscillatory nature of the plagioclase requires a growth mechanism that is characterized by the punctuated alternation in the ratio of Ca and Na taken into the growing crystals. Systematic fluctuations of the "feeder" melt geochemistry adjacent to the growing plagioclase laths, i.e., changes such as Na-Ca ratio, silica content, water, etc., could account for the oscillatory character. Fluctuating changes in the crystal-melt system within the Morena Reservoir pluton recorded by the oscillatory zoning in plagioclase could be attributed to an open system process, i.e., multiple repetitions of local melt replaced by fresh melt driven by convective forces. Alternately, a closed-system model for the formation of oscillatory zoning in plagioclase would have to involve the cyclic modifi-cation or evolution of either, the local interstitial melt composition by the ongoing crystallization of plagioclase, i.e., the dynamic changes in the Ca and Na concentrations in the interstitial melt associated with the growing plagioclase crystals, or a physical parameter such as temperature or pressure.

The crystal-plastic and brittle strains exhibited by the megacrystic quartz and phenocrystic biotite, and plagioclase within a groundmass that is dominantly magmatic and undeformed suggests deformation of the rock in a submagmatic state, i.e., 10%? to 30%? melt (Paterson and others, 1989). The relation between the submagmatic deformation and the development of the magmatic foliation and the above described disequilibrium textures is unclear, however, most if not all the strain features in these rocks appear to be synplutonic. The preferred alignment of the plagioclase and hornblende phenocrysts and elongate quartz megacrysts can be attributed to several mechanisms, such as alignment of suspended elongate crystals in a laminar flow, or preferred crystallization attitude on growth surfaces, or alignment by crystal rotation while the rock was under a flattening stress (ballooning?) during submagmatic conditions. The first two mechanisms are generally disfavored because they can not account for the deformed nature of the rocks, i.e. deformed/strained phenocrysts in a typically undeformed groundmass. The third mechanism is favored because it is compatible with the deformation fabrics, i.e. alignment of a crystal-rich mush by compression where phenocrysts impinge on each other and undergo various degrees of synchronous alignment and strain in a low melt fraction. A progressive flattening stress would impart the observed deformation (strain and foliation) features. Progressive flattening forces might initiate the mobilization of interstitial melt from between grains and fractures. The above mechanism could effectively affect the petrogenesis and emplacement of the pluton by modifying the late-stage crystal-melt geochemical composition and rheology of the magma, i.e. melt migration and crystal compaction.

The submagmatic deformation by compressive forces is most likely related to an emplacement-related "space" problem, either from internal magma pulse diapirism and ballooning, or from local-field or far-field tectonic compression or transpression.

The secondary subsolidus alteration of the major magmatic mineral phases, particularly the replacement of plagioclase and hornblende by respective epidote and chlorite, indicates that fluids interacted with these rocks subsequent to crystallization. The timing of secondary alteration is unknown.

Geochemistry

Introduction

A reconnaissance sampling and analyses of the Morena Reservoir tonalite for major, minor, and trace elements was undertaken to both, delineate pluton-scale compositional zoning and for interpreting the pluton's petrogenesis. The notably homogeneous small- and intermediate-scale mineralogic variation makes a small geochemical study of the Morena Reservoir's gabbro, quartz diorite, and tonalites useful to delineate any large-scale zonations in chemical composition. Twelve samples were chosen from a sample selection of forty that best represents each region of the pluton (Plate I). No samples of mafic enclaves or dikes were analyzed.

Sample Collection and Processing Procedures

Sampling of the Morena Reservoir pluton was made across the pluton in a roughly 1 km^2 grid pattern for the freshest rocks. The freshest portions of 6 to 10 kg samples were sent to Washington State University GeoAnalytical Laboratory for analyses of major, minor, and trace elements (XRF), and rare earth elements (ICP-MS).

Analytical Results

The results of the geochemical analyses are given in Appendix B. Some duplication for several rare earth elements occurs between the tables because certain elements were analyzed by both methods. The ICP-MS method results for those elements are considered to be more accurate. Major and minor element data are expressed as weight percent (wt%) oxides, and are normalized on a volatile-free basis. The total iron is expressed as FeO*. The trace and rare earth element data are in parts per million (ppm). The symbol "#" denotes values >120% of the lab's highest standard. There are some significant differences between unnormalized and normalized results, particularly for silica. The difference in silica weight percent between unnormalized and normalized may be as much as 1%. This appears to be a reflection of the low unnormalized weight percent totals in most samples, i.e., weight percent totals of 99.80 or less.

The geochemical results have been organized into various graphs and diagrams that include Harker diagrams (Figures 29-34), REE and spider diagrams (Figure 35), and ternary classification and tectonic environment diagrams (Figures 35 and 36).

The geochemistry of the Morena Reservoir pluton rocks shows modest yet significant variations. The systematic variations are both large enough and spatially defined to imply pluton-scale differentiation. Therefore, a system of comparing geochemical patterns and relationships is necessary for defining and understanding the processes and mechanisms of differentiation that may have operated in the Morena Reservoir pluton before and during crystallization.

An element with wide systematic variation makes a good index against which chemical differentiation can be evaluated. Harker diagrams compare SiO_2 against other elements of the sample suite. Relatively tight linear to curvilinear geochemical trends of rocks that are spatially and temporally associated indicate a cogenetic suite, i.e., a single pluton. These trends would thus represent an evolving geochemical relationship between the different parts of the pluton, in terms of partial melting, crystallization differentiation, assimilation, or hybridization processes. The Morena Reservoir tonalite appears to satisfy the above relations.

Major and Minor Element Geochemistry

SiO_2 has the greatest absolute variation for the major oxides in the Morena Reservoir pluton, from 55 wt% to 68 wt% for the quartz diorite and tonalite suite, a SiO_2 range of 13 wt% (Appendix B). A 5 wt% gap in silica exists between the analyzed gabbro samples suite from the core and the pluton's sampled suite of quartz diorites and tonalites; this gap could result from the limited sampling, or it could be real. Al_2O_3 ranges in the tonalites and quartz diorites from roughly 15.5 wt% to 18.0 wt%. FeO* ranges from 4 wt% to 8 wt%, and MgO varies from 1 wt% to 5 wt%. CaO ranges from 4.0 wt% to 7.5 wt%. Na_2O and K_2O have wt% variations from roughly 3.5 to 4.7 Na_2O and 1 to 2 K_2O. P_2O_5 and MnO have wt% ranges of 0.15 to 0.23 P_2O_5 and 0.08 to 0.13 MnO.

SiO_2 variation in the Morena Reservoir pluton rock suite can be used as an index of differentiation. Other elements will be compared to SiO_2 in Harker diagrams for determining each element's behavior in the rock suite.

Harker diagrams of the major and minor elements are shown in Figures 26 and 27. The diagrams show tight linear to curvilinear trends for the combined quartz diorite and tonalite suite, with some discrepancies, particularly for the two samples with respectively highest and lowest silica content (MR-R761 and MR-R768). TiO_2, Al_2O_3, FeO*, MgO, CaO, MnO, and P_2O_5 have negative slopes, and K_2O has a positive

slopes; Na_2O appears to have no slope. Sample MR-R761 has relatively high Na_2O, MnO, and P_2O_5, and relatively low K_2O. Sample MR-R768 has relatively low FeO*. The two gabbroic samples plot in line with the quartz diorite/tonalite suite for major and minor elements, however separated by the 5 wt% silica gap; this is particularly true for TiO_2, Al_2O_3, FeO*, Na_2O, and K_2O. The gabbroic sample MR-R790 plots off trend in MgO, MnO, and P_2O_5.

Except for the Na_2O trend and the samples MR-R761 and MR-R768 in specific diagrams, the major and minor elements behave predictably. Each element is systematically enriched or depleted relative to modeled and experimental crystallization differentiation schemes for a common granitoid magma that produces a normal suite of tonalite rocks. The trends found in the major and minor element Harker diagrams of the Morena Reservoir pluton samples suggest a comagmatic suite of rocks. This can be supported or refuted by examining the trace and rare earth element data. There are two general reasons why Na_2O concentrations are nearly equal across the pluton: 1. analytical error/insensitivity; and 2. the Na_2O trend is real but of unknown origin.

Mineral norms were calculated from the geochemistry. The Morena Reservoir pluton rocks fall within the tonalite and quartz diorite fields (Figure 28). The albite-quartz-orthoclase (Ab-Qz-Or) and albite-anorthite-orthoclase (Ab-An-Or) ternary diagrams show little to no normative orthoclase, and a fairly limited Ab:An range of $Ab_{70}An_{30}$ to $Ab_{80}An_{20}$ (Figure 28).

Figure 28. Mesonormative mineral diagrams of the Morena Reservoir suite. Q and Qz = quartz; P = plagioclase; A = alkali feldspar; AN = anorthite; OR = orthoclase; and AB = albite. Symbols: box = tonalite; x = gabbro).

AFM and AFC diagrams are shown in Figure 29. Excluding two samples in the AFM diagram, all samples including the gabbros fall into a limited area. These diagrams indicate very little variation in the three-way proportion between these elemental groups, even though the silica content changes significantly.

The Peacock diagram for the Morena Reservoir plutonic suite (Figure 29) places these rocks into the calcic category which is comparable to the general trend in the northern Peninsular Ranges batholith (Silver, 1979). A diagram defining a rock as metaluminous, peraluminous, or peralkaline has been devised by Maniar and Piccoli (1989). In this diagram (Figure 29), the Morena Reservoir plutonic suite falls within the metaluminous field along a steep, negatively sloped trend that is fairly close to the vertical metaluminous-peraluminous line.

Trace Element Geochemistry

Trace element geochemistry may help define source region character, melting processes, and potential magmatic processes, particularly crystal fractionation (Green, 1980). The Morena Reservoir plutonic suite has several trace elements with curvilinear trends in Harker diagrams,
suggestive of participation in a differentiation process. Sr, V, Zn, and Cr show negative slopes (Figure 30), and Rb, Ba, Pb, and Cs have positive slopes (Figure 31). These elements act compatibly during differentiation, and fractionate into plagioclase (Sr), hornblende/biotite (Rb and Cs), and the iron opaque minerals (Cr and V). Zr can be used as an index of differentiation in evolved granitoid suites, but has a scattered Harker diagram pattern in the Morena Reservoir pluton rocks. Although zircon (where virtually all Zr is concentrated) is detected in thin sections, Zr appears to have not participated in any differentiation process (Figures 32). Other trace elements that do not show any systematic variations with silica are Ni, Sc, Cu, Nb, Y, Th, and Sn (Figures 32 and 33).

Several trace element diagrams designed to discriminate tectonic environment are illustrated in Figures 34 and 35. These diagrams characterize the Morena Reservoir plutonic suite as a medium-K, orogenic, calc-alkaline, andesitic, arc granitoids formed in a subduction-related, magmatic arc.

Rare-Earth-Element Geochemistry

Introduction. Rare-earth elements substitute for the major and minor elements in the common and accessory minerals (Le Marchand and others, 1987). Most of the REE's reside in the accessory minerals. However, hornblende, pyroxene, plagioclase, and garnet can also appreciably fractionate specific REE's: garnet, amphibole, and pyroxene prefer the HREE's, whereas apatite and sphene prefer the LREE's. Hence, specific minerals fractionate the REE's such that a given mineral can produce a unique REE signature. A problem in deciphering a REE pattern is the potential complication and overlapping of fractionating mineral systems that are active during the partial melting episode(s) that generate the parental magma(s), and the crystal-melt differentiation processes active during the ascent, emplacement and crystallization of the pluton.

Figure 34. Tectonic environment discrimination diagrams of the Morena Reservoir suite. Upper diagram: data lie within the medium-K, orogenic andesite field. Lower diagram: data lie within the pre-plate collision field. Symbols: box = tonalite; x = gabbro.

Figure 35. Tectonic environment discrimination diagrams of the Morena Reservoir suite. WPG = within plate granites; ORG = ocean ridge granite; VAG = volcanic arc granites; syn-COLG = syn-collisional granites. Symbols: box = tonalite; x = gabbro.

Rare-Earth-Element Analyses. Eleven samples were analyzed for rare-earth elements and chondrite-normalized (Figure 36). Chondrite values are from Boynton (1984). The REE patterns for the quartz diorite and tonalite suite are slightly curved, concave-up profiles that are generally subparallel. They have modest LREE enrichment with either, no Eu anomaly, or a slight negative Eu anomaly. The subparallel HREE to MREE patterns suggest a common history. The gabbroic core sample MR-R790 has an enriched, mildly-curved, convexupward, profile, with a small positive Eu anomaly. This sample approximates the heavy to middle REE profiles of the quartz diorite and tonalite suite, but has substantially lower LREE abundances relative to the granitoid samples. The analysis of each phase, common and accessory, for REE's from each sample might clarify the partitioning of REE's. The REE patterns appear to reflect several mineral melt fractionating systems that may relate to both the partial melting episode(s) of the source region and the crystal fractionation of the parental emplacing melt(s). The concave-up, negative-sloped REE patterns, with highly consistent La/Yb ratios of 3 to 4, suggests involvement of amphibole or possibly clinopyroxene in the absence of garnet. The small negative Eu anomalies further suggest the modest participation of plagioclase. These REE crystal-melt fractionating systems may have operated during the partial melting of the source region, or during the early stages of crystal fractionation of the parental melt. The tonalite and quartz diorite patterns closely correspond to a reflection of an amphibole mineral/liquid distribution coefficient (Figure 36). This suggests amphibole in the source region. Alternately, the patterns could be produced by the crystallization and removal of amphibole. Hornblende mode varies from 7 to 15%, roughly a factor of two. Pattern variation of La is 21x to 47x, roughly a factor of two. This relationship suggests that the subparallel patterns, not the abundance, are a feature of the parent melt, and that REE abundance is attributed to the amount of amphibole in the rock.

In contrast to the tonalite and quartz diorite, the gabbro pattern generally resembles a "negative" reflection of a plagioclase mineral/liquid distribution coefficient (Figure 36). This most likely suggests the systematic removal of plagioclase from the crystallizing melt (crystal fractionation via crystal floating?).

Geochemistry Discussion and Conclusion

The compositional variation in the Morena Reservoir tonalite suite is attributed mainly to the crystallization differentiation of a single parental melt. This is suggested by curvilinear trends in major and trace element Harker diagrams, the modest range in silica oxide, and the subparallel REE profiles. Minor assimilation of the wall rocks is considered to have produced the geochemical anomalies in samples from the margins of the pluton. The ubiquitous deuteric alteration of the rocks is believed to have resulted from migrating late-stage, magmatic hydrothermal fluids.

Major element Harker diagrams, except for Al_2O_3 and Na_2O, have smooth, curvilinear differentiation trends, as do the trace elements Cr, V, Cs, Sr, and Rb. These elements behave in ways that closely correspond to a crystallization differentiation process.

The Harker diagram for Al_2O_3, and possibly CaO, have nearly linear trends. Linear trends are typically associated with mixing processes. Na_2O, Ba, and Zr do not increase in abundance with differentiation as defined by increasing SiO_2, but are relatively constant. These elements normally behave incompatibly during the early stages of magmatic differentiation because early forming minerals do not incorporate these elements. However, early-forming zircon would take Zr; and all samples have prismatic zircon. Geochemical conditions may have favored crystal-melt distribution coefficients of near unity for

Na_2O and Ba. Alternately, hydrothermal fluids might have altered the Na_2O and Ba abundances. However, other relatively mobile elements such as K and Rb should have been equally affected. It therefore seems most likely that the geochemical abundance patterns reflect systematic differences in either the initial upwelling parental melt composition, or in element-specific homogenizing effects of a convecting magma within the crystallizing chamber, or in the element-melt distribution coefficients.

The geochemical variation patterns could be produced if the pluton was connected to a much larger parental magma body at depth. Large-scale, dynamic, magma (crystal-melt) interactions, or zone refining might have taken place between the pluton and the much larger parental magma source. This process would eliminate the need for exposure-level crystallization fractionation to account for the compositional variation in the rocks. The variations would then be attributed to compositional differences in a succession of unique pulses, assuming that the rocks closely reflect near-parental melt compositions.

Sample MR-R761 has numerous relatively anomalous values that suggest an effect from outside the pluton such as magma mixing or wall rock assimilation.

REE patterns are consistent for the granitoid suite, both in La/Yb ratio and in total REE abundance. This suggests crystallization differentiation from a single parental source melt.

The gabbroic core rocks may have a remote genetic connection with the quartz diorite- tonalite suite, but very unlikely at the level of emplacement.

Geochronology

Introduction

Radiometric dating of the Peninsular Ranges batholith is mostly reconnaissance-style (see Silver and others 1979; 1988). However, in progress is U-Pb work by D. Kimbrough and others at San Diego State University, and Ar-Ar work by A. Ortega at Queens University, Ontario. These two projects will attempt to constrain the timing of emplacement and the cooling history of many of the plutons. The Morena Reservoir pluton is included in both projects. A single coarse zircon fraction from the tonalite at locality MR-R756 of the Morena Reservoir pluton was analyzed for U-Pb systematics. Two samples from localities MR-R591 and MR-R756 were analyzed for hornblende ^{40}Ar-^{40}Ar systematics.

Sampling and Processing Methods

One 50 kg sample was collected for zircon U-Pb dating. The sample was collected near the northern margin of the pluton at sample locality MR-R756 (Plate I). Also from locality MR-R756 a separate 10 kg sample was later collected and processed for hornblende ^{40}Ar-^{40}Ar dating. Another 10 kg sample was taken from locality MR-R591 and processed for hornblende ^{40}Ar-^{40}Ar dating (Plate I). All three samples are medium-grained, moderately foliated, medium-grained, biotite-hornblende tonalites. Macroscopic and microscopic examination of these rocks revealed modest amounts of deuteric alteration of plagioclase, hornblende, and biotite to epidote and chlorite.

The samples MR-R756 and MR-R591 were processed for zircon and/or hornblende concentrates by the mineral usual separation methods. The zircon dissolved by methods discussed in Parrish (1987). U and Pb were isolated by techniques of Krogh (1973).

Uranium-Lead Zircon Analyses

Certain assumptions must be made in the U-Pb zircon analysis. These include: 1) zircons remained closed to U and Pb, and all intermediate daughters throughout its history; 2) correct values are used for the initial Pb isotope ratios; 3) the decay constants of 235U and 238U are known and accurate; 4) the isotopic composition of U has not been modified; and 5) all analytical results and errors are accurate (Faure, 1977).

Only a single coarse-sized zircon fraction was analyzed from the Morena Reservoir pluton for a crystallization age. The analytical results are compiled in Table 2.

Table 2.

~~U-Pb Data - Morena Reservoir Pluton~~

Fraction	weight (mg)	Pb (ppm)	U (ppm)
coarse	0.0052	8.4314	519.39

Pb Isotopic Compositions - Pb-Unspiked Aliquot, Corrected

208/206	207/206	207/206
0.12933	0.053717	0.000314

Radiogenic Ratios

Age (m.y.)

206/238	207/235	207/206	7/6 error +/-
0.01586	0.107346	0.04910333	
101.4	103.5	152.8	2

The Pb-Pb age for sample MR-R756 is roughly 152.8 +/-2 Ma, and this is about 50 Ma older than the U-Pb ages. The discordancies between the concordant $^{206}Pb^*/^{238}U$ and $^{207}Pb^*/^{235}U$ ages and the Pb-Pb age for this sample can be due to 1) recent Pb loss by either episodic release (metamorphic event), steady-state diffusion, or weathering; or 2) the inheritance of older zircons. Recent Pb loss would decrease the Pb-U ages, and inheritance of very old radiogenic Pb would increase the Pb-Pb ages.

The low concentration of radiogenic ^{207}Pb relative to radiogenic ^{206}Pb in these samples is important in assigning an age. The nearly 1 x 10^2 ppm difference in concentrations makes the $^{207}Pb^*/^{235}U$ ratio relatively less accurate in terms of number of significant figures. Therefore, the Pb-Pb age is less reliable, and the $^{206}Pb^*/^{238}U$ age is preferred over the $^{207}Pb^*/^{235}U$ age.

Two zircon samples from the Long Potrero pluton have been processed for U-Pb systematics from zircon. Of the two samples, sample LP-R18 has been analyzed. The preferred U-Pb age for the hornblende-biotite facies of the Long Potrero pluton is 103 Ma +/-2 Ma (Kimbrough, written communication, 1990). An U-Pb zircon age of 94 Ma +/-2 Ma was determined for the La Posta pluton's western hornblende-biotite facies (Clinkenbeard, 1986).

Geochronology Discussion

The U-Pb zircon ages conform to the interplutonic field relations between the Long Potrero, Morena Reservoir, and La Posta plutons. Field relations discussed in Chapter V, suggest that the Long Potrero pluton's northeastern flank was intruded by the Morena Reservoir pluton, and that the Morena Reservoir pluton's eastern flank was intruded by the La Posta pluton.

The 103 Ma zircon U-Pb age for the Long Potrero pluton, the 101 Ma zircon U-Pb age for the Morena Reservoir pluton, and the 94 Ma U-PB age for the La Posta pluton, are in agreement with the field relations. Considering the +/- 2 Ma on these dates, the Long Potrero and Morena Reservoir plutons may be contemporaneous.

Argon-Argon Cooling Ages

Two samples from the Morena Reservoir plutons have been analyzed for ^{40}Ar-^{39}Ar hornblende cooling ages by A. Ortega as part of her Doctoral Dissertation (in progress) at Queens University, Ontario. The preliminary hornblende cooling ages for samples MR-R756 and MR-R591 are 95 and 101 Ma +/- 2 Ma, respectively (Amabel Ortega, oral and written communication, 1994). These ages are close to and identical with the preferred U-Pb zircon age of sample MR-R756. The hornblende age indicates rapid cooling of the pluton after initial crystallization. Alternately, the 95 Ma cooling age could be a reset age from La Posta plutonism. Sample locality MR-R756 is 6 km from the La Posta pluton; it is unknown whether a 6 km radius is within the La Posta pluton's thermal aureole.

A sample from the Long Potrero pluton's intermediate facies was analyzed for ^{40}Ar-^{39}Ar hornblende cooling age by A. Ortega. This sample (LP-R18) had an integrated age of 101 Ma +/- 1.4 Ma. This age is only 2 Ma younger than the 103 Ma +/- 2 Ma U-Pb zircon age from the same locality. This relation is nearly identical to those in the Morena Reservoir pluton, and adds further weight to a model of rapid subsolidus cooling for these plutons.

Magnetic Susceptibility

Introduction

Magnetic susceptibility or MS, is the acquired magnetization of a substance in the presence of an external magnetic field. MS is defined as the ratio of the intensity of induced magnetization over the applied field. As a vector equation, it is defined as: $J = kH$, where J is the intensity of magnetization, H is the applied field, and k is the magnetic susceptibility. The magnitude of k is measured in dimensionless SI units, and can be converted to the c.g.s. system by the relationship: k(SI) = 4 k(c.g.s.).

The magnetic susceptibility is one component contributing to the magnetization of a rock; other components are the various types of remnant magnetization. The factors controlling the magnetic susceptibility in a rock include composition and abundance of mafic minerals, grain size and shape, crystallographic anisotropies, and grain distribution.

Magnetite and ilmenite are the two main opaque iron- oxide minerals in granitoid rocks. Magnetite is the major contributor to magnetic susceptibility and is roughly three orders of magnitude higher than ilmenite. Ishihara (1977 and 1979) recognized magnetite- and ilmenite-series granitoid rocks as follows:

magnetite-series	ilmenite-series
1. magnetite-rich	1. ilmenite; magnetite-poor
2. high m.s ($>130\times10^{-5}$ SI)	2. low m.s ($<130\times10^{-5}$ SI)
3. high bulk Fe_2O_3/FeO	3. low bulk Fe_2O_3/FeO

A magnetic susceptibility "step" occurs in the Morena Reservoir pluton, and separates outcrops which have values below 60 $x10^{-5}$ SI (ilmenite-series), or above $200x10^{-5}$ SI (magnetite-series).

Magnetite or ilmenite may solely crystallize, or both may form as an intergrowth in a crystallizing melt.
The quantity and nature of the magnetite and ilmenite are a function of the rock's geochemical history. The phase relations of magnetite and ilmenite are a product of the evolving conditions in the magma and the resultant rock. Conditions that effect the crystallization and change of magnetite and ilmenite from the initial melt generation to postmagmatic deuteric alteration include FeO*, TiO_2 , MgO, SiO_2, Al_2O_3, organic C and ammonia concentrations, O_2, H_2O, H_2S, Cl_2 fugacities, temperature and pressure, and postplutonic magmatic/hydrothermal fluid effects (Shimazu, 1961; Sack and others, 1980). Later conditions most likely mask earlier conditions. Crustal carbon from the source region and wall rock has been proposed as a major controlling factor in opaque phase preference in magmas, but this does not appear to be true in the PRB (Gastil, oral communication, 1991). Instead, either the depth of opaque mineral crystallization, or depth of emplacement, or depth of melt generation, probably is the major control on magnetite and ilmenite formation (Gastil, oral and written communications, 1993). If the depth of emplacement is the primary control on magnetite crystallization, then the magnetite zoning in a pluton can provide information on the magma dynamics, e.g., magnetite-rich zones are from the shallow parts of the chamber, and ilmenite-rich zones are from deeper parts of the chamber. The zoning at a given erosional depth might suggest magma convection in the pluton's magma chamber

prior to solidification, i.e., synchronous upwelling and downwelling of magma; this could apply to the Morena Reservoir pluton.

Magnetic susceptibility was measured at over 300 localities within the Morena Reservoir pluton, using a hand-held magnetic susceptibility meter. Twenty to fifty magnetic susceptibility measurements were recorded at each locality in order to establish a detailed range of variation. Magnetic susceptibility readings were taken between successive numbered data stations in order to insure that the location-to-location MS variability is consistent.

Magnetic Susceptibility Data

The magnetic susceptibility variations in the Morena Reservoir pluton (Plate IV) form a concentric pattern that define a heterogeneous yet distinctive distribution of magnetite. The MS contours are divided into ten intervals (Plate IV): violet (0 to 39)/(0 to 0.03); lavender (40 to 129)/(0.03 to 0.07); blue (130 to 299)/(0.07 to 0.14); dark green (300 to 499)/(0.14 to 0.21); light green (500 to 799)/(0.21 to 0.31); yellow (800 to 1299)/(0.31 to 0.46); orange (1300 to 1999)/(0.46 to 0.65); orange-red (2000 to 3499)/(.65 to 1.10); red (3500 to 4999)/(1.10 to 1.40); and magenta (5000 and up)/(1.4 and up); all values ($x10^{-5}$ SI) and (volume percent), respectively. Magnetic susceptibility intervals are divided at the value $130x10^{-5}$ SI following the division made by Ishihara (1977) for the magnetite-series/ilmenite-series boundary. The other contour divisions were chosen to most fully develop the magnetic susceptibility/magnetite data.

Contouring Results

The contoured magnetic susceptibility and inferred volume percent magnetite data of the Morena Reservoir pluton is displayed over topography and geology in Figure 6 and Plate IV. It shows that the pluton consists of a narrow, variably magnetic margin, and a large, nonmagnetic intermediate and core region. These patterns are generally concentric to the pluton's form, and appear to have local magnetic correspondence with the wall rocks.

The largest zone of high magnetic susceptibility and inferred volume percent magnetite is found in the northeastern part of the pluton, between moderately to strongly magnetic granodiorite wall rocks. The southeastern end of this zone appears to be truncated at the La Posta pluton contact. Other smaller zones of high values are found along narrow regions of the pluton's outer and inner margins. Notable areas are located adjacent to the moderately magnetic Morena Butte granodiorite, the highly magnetic Los Pinos gabbro, and along the northern edge of the pluton's moderately magnetic gabbroic core body. The small, highly magnetic zone within the Morena Reservoir pluton that partially wraps around the gabbroic core appears to be truncated along their common contact.

A large, continuous, northwest-trending nonmagnetic zone makes up virtually the entire central part of the pluton, and extends into the La Posta pluton, a totally nonmagnetic pluton. The outline of this zone forms the local segment of the regional magnetite-ilmenite boundary of Gastil and others (1990). The north and south continuations of this segment of the boundary runs along the La Posta pluton's western margin.

Magnetic Susceptibility Discussion

The Morena Reservoir pluton's magnetite distribution pattern appears to be influenced by the regional magnetite-ilmenite boundary, local external contact effects, and its internal petrogenetics.

The outline of the largest MS feature, the nonmagnetic zone that appears to extend from the magnetite-free zone of the La Posta pluton, forms a small part of the regional magnetite boundary that extends along the entire length of the batholith. This feature generally corresponds to a suggested tectonic boundary that divides the western and eastern parts of the PRB. The western magnetite-bearing rocks versus the eastern ilmenite-bearing rocks most likely relate to a depth-dependent factor that controls the redox conditions in the magmas, such as depth of the subducting slab and zone of melting, or depth of ponding of parental melts. In the case of the Morena Reservoir pluton, its petrogenesis appears to be partly similar to that of the La Posta pluton in terms of opaque mineral crystallization.

The magnetite-rich zones of the pluton's margin occur where wall rocks are also magnetite-rich, and magnetite-poor (ilmenite-rich?) zones are found next to magnetite-poor (ilmenite-rich?) wall rocks. This would implies that the wall rocks apparently control, to some degree, the redox conditions of the marginal parts of the emplacing magma and the formation of magnetite and ilmenite.

Variations in magnetite content at all scales in the pluton not attributed to contact effects are probably related to conditions of magma emplacement. On a large scale, zones of high- and low-magnetite abundance or zones of high and low total opaque abundance most likely correlate to the pluton's temporal (early- versus late-forming parts of the pluton) and spacial (paleodepth of the pluton's magma chamber) character, as defined by its petrogenesis.

Mafic Enclaves

Introduction

Mafic enclaves found in the Morena Reservoir pluton typically consist of mafic-rich, tonalite/quartz diorite, and have spheroidal to discoidal shapes. They most commonly consist of plagioclase and hornblende with lesser amounts of quartz and biotite. Mafic enclaves have both igneous and metamorphic textures and are found in outcrop as isolated, well-defined bodies in random patterns or as well-defined sheet-like clusters.

They have variations in form and size that are observable at all scales of the pluton. Form is most significant at the pluton scale whereas size becomes more variable at the outcrop scale.

Mafic enclaves of similar form, composition and texture are ubiquitous in the local tonalites (Hurlbut, 1935; Lombardi, 1991; Rector, 1993; V. Todd; oral communication, 1994).

Mafic enclaves are typically oblate to some extent and commonly have preferred orientations concordant with the host-rock foliation. The degree of enclave flattening has been used to measure shortening in other plutons that has been attributed to either diapiric ballooning or regional tectonic contraction (Tobisch and others, 1993; and John and Bundy, 1994).

Field data on mafic enclaves in the Morena Reservoir pluton include lithology, texture, abundance, size, shape, orientation, and magnetic susceptibility. Field data and measurement methods are found in Appendix D.

Data for mafic enclave shape of the Morena Reservoir pluton have been contoured and overlaid on topography and geology (Plate VI).

Mineralogy and Texture

Compositionally, mafic enclaves of the Morena Reservoir pluton are porphyritic, medium-grained, biotite-hornblende diorite, quartz diorite, and tonalite. Mineralogy in order of decreasing abundance is plagioclase, hornblende, biotite, quartz, epidote, opaque minerals, chlorite, apatite, and zircon. Mafic enclave texture is porphyritic, medium-grained (0.5 to 3 mm) and moderately recrystallized with variably strained, coarse-grained (5 to 7mm), euhedral to subhedral plagioclase, hornblende, and biotite phenocrysts. Modes were not made on mafic enclave samples.

The structure of mafic enclaves is covered in Chapter V.

Mafic Enclave Abundance – Variations in Outcrops

Variation of mafic enclave abundance is greatest at the outcrop scale. Large outcrops may show local variations from a few mafic enclaves per 9 m^2 of outcrop to several 100 per 9 m^2 of outcrop. Mafic enclaves typically occur in randomly arranged patterns. However, in outcrops were mafic enclave abundance is high, groups of densely spaced mafic enclaves form "matrix-supported" (individually separated by host tonalite), sheet-like clusters. These clusters vary in width (10 cm to 10 m) and density (10 to 80 volume percent of host-rock/enclave rock). The mafic enclave sizes in the sheet-like clusters vary from 3 cm to 3 m). Mafic enclave sheet-like clusters are typically occur around the margin of the pluton and near intraplutonic structural discontinuities.

It appears that different sheet-like clusters of mafic enclaves are at different stages of having been dissected and dispersed by the host tonalite rock, i.e., from synplutonic(?) "dike-like" mega-enclaves (massive, tabular bodies 10's of meters long and meters wide, with only mm-wide tonalite injection veinlets), to sheet-like clusters of widely dispersed mafic enclaves (10's of cm-wide separations between mafic enclaves). The best localities exhibiting this phenomena are along the western and eastern sides of the pluton. Mafic enclaves outside the sheet-like clusters appear to have a random arrangement.

Contour Patterns of Mafic Enclave Abundance

The mafic enclave data of the Morena Reservoir pluton have been contoured for variation in averaged abundance, based upon the number of enclaves per unit area and overlaid on topography and geology (Plate V). The mafic enclave abundance map is divided into ten units. Each unit defines a range of abundance (average count per $9m^2$): violet = less than 1; lavender = 1 to 4; dark blue = 5 to 14; light blue = 15 to 29; green = 3 to 59; yellow = 60 to 99; orange = 100 to 199; orange red = 200 to 299; red = 300 to 399; magenta = 400 and up.

The pluton-scale variations of contoured mafic enclave abundance reveal patterns that are generally elongated subparallel to the pluton's northwest trend, and mostly concordant to the contacts with the wall rocks. These patterns do not appear to be centered about the core of the pluton, but are broken up into several distinctive zones of relative high and low abundances.

Zones of highest abundance are found in the pluton's marginal rocks. Isolated zones of high abundance are also found in several localities in the pluton's intermediate regions, particularly between the Los Pinos pluton and the Morena Reservoir pluton's gabbroic core. Zones of lowest mafic enclave abundance are found in the east-central part of the pluton, adjacent to Morena Butte, between the Los Pinos pluton and the gabbroic core, and near its southern contact with the Long Potrero pluton.

Abrupt changes in continuous patterns are found around the pluton's gabbroic core and along the Hauser Canyon lineament.

Discussion on Mafic Enclave Origin and Abundance Patterns

The distribution pattern of mafic enclaves in the Morena Reservoir pluton has great tectonic-plutonic significance if the origin and history of the mafic enclaves can be determined. Field observations indicate that over 95% of examined mafic enclaves have similar megascopic composition, texture; and form. They are foliated, medium-grained, porphyritic, mafic-rich quartz diorite with euhedral plagioclase that form submeter-size, flattened spheroids and discoids. Megascopic mafic-enclave foliation is typically concordant with the host foliation.

Field observations have also shown a definite, close connection between synplutonic mafic diorite dikes/tabular megaxenoliths and mafic enclaves. Tonalite has intruded the mafic rock and dispersed it into sheet-like clusters of mafic enclaves. Further dispersion of the sheet-like clusters most likely formed the more common random arrangements of mafic enclaves. When the sub-meter-sized mafic enclaves were separated from the multi-meter-sized parent enclave body by the invading tonalite magma, the injecting host tonalite appears to have been very fluid. And the invaded mega-enclaves appear to have been ductile, yet rheologically distinct. This phenomena is greatest near the margin of the pluton, but also occurs along concentric arcuate zones within the pluton, where mafic enclaves, synplutonic mafic dikes, tabular, mafic megaxenoliths, and schlieren are plentiful.

Contour Patterns of Mafic Enclave Morphology

The Morena Reservoir pluton is contoured for shape variation of mafic enclaves, and is illustrated in Plate VI. Shapes of mafic enclaves are expressed in terms of a subhorizontally measured length to width ratio, l:w; height is not expressed, but it can be considered essentially equal to length. The mafic enclave morphology map is divided into seven unequal intervals or units, and are defined by a length:width ratio (l:w) as follows: violet = 1 to 1½; blue = 2 to 2½; green = 3 to 3½; yellow = 4 to 5½; orange = 6 to 8½; red = 9 to 12; magenta = 12½ and up. Flattening ratios from each data locality used for contouring are averages. Each data locality has a distinctive range of shapes that may include morphologic variations that span across two to three contour units. However, in the vast majority of outcrops, roughly 80% of the mafic enclaves have a width:length ratio that falls into a single contour interval.

Dikes

Synplutonic and postplutonic dikes occur throughout much of the Morena Reservoir pluton. The dikes are distributed in zones and are subparallel to host rock lithologic and structural patterns. Several general types of dikes were distinguished in the Morena Reservoir pluton that include leucocratic, fine-grained tonalite; mesocratic, fine-grained tonalite; fine- to medium-grained melanocratic tonalite to diorite; aplite-textured granodiorite to granite; and granite pegmatite dikes. A number of the leucocratic to melanocratic dikes with diorite to tonalite compositions have sinuous, disrupted forms that can have diffusive contacts with the host rock, and locally deflect, but do not truncate the magmatic foliation. The above character suggests that these dikes have a synplutonic origin. Most of the aplitic and granite pegmatite dikes are tabular with sharp contacts. They truncate the host rock's foliation, typically at high angles. The aplite-textured and pegmatitic granite dikes are considered to be postplutonic.

Synplutonic, fine- to medium-grained leuco- to melanocratic dikes are located mostly in the marginal parts of the pluton, and typically occur in rocks that contain schlieren. These dikes typically are oriented subparallel or at low angles to foliation. In some outcrops, contacts between synplutonic dikes and schlieren are gradational, where the dikes have irregular, sinuous form and diffuse contact between the dikes and the host rock.

Leucocratic and melanocratic dikes rarely occur at the same scale, and appear to have distinctive zones of injection within the marginal part of the pluton. Pegmatite and aplite dikes are associated with either leucocratic or melanocratic dikes. The melanocratic dikes are concentrated close to the pluton's external contact, and the more leucocratic dikes in the interior parts of the pluton. Large, meter-scale, melanocratic dikes are associated with tabular zones containing abundant mafic enclaves, most notably near the pluton's northeastern margin.

Granite pegmatite and aplite dikes are found along the pluton's outer external contact and in and around the pluton's gabbroic core. Multiple-sheet, meter-scale dikes and pods of granite pegmatite and aplite, with northwest strikes and very low southerly dips, crosscut both the tonalite and gabbro core. The dikes appear to be much wider and numerous in the gabbro core and along the external contact of the gabbro core.

The intermediate part of the pluton is virtually devoid of dikes.

CHAPTER IV
STRUCTURE OF THE MORENA RESERVOIR PLUTON

Pluton Form

The Morena Reservoir pluton has a Y"-shaped outline elongated NW with three narrow appendages (Figures 4 and 5; Plate II). The western side of the pluton has a smooth, regular edge with a sharp contact with the Long Potrero pluton and the granodiorite body of Morena Butte. The western side of the pluton dips steeply inward, as inferred by dip of contact and foliation. The eastern side of the pluton is smooth and regular, and has a sharp contact with the granodiorite body of The Narrows and the La Posta pluton. The southern part of the eastern side of the pluton dips steeply outward, and the northern part dips steeply inward. The northwestern corner of the pluton has a somewhat irregular shape adjacent to the Morena metasedimentary screen, the granodiorite of Corte Madera, and the Los Pinos gabbroic pluton. The northern side of the pluton is generally dips steeply outward. Based on the local foliation, the contact with the gabbro core body dips steeply inward toward the core's center.

Interplutonic Field Relations

Most of the pluton's external contacts are sharp and concordant with the wall rock foliation, except for the gabbro and metasedimentary rocks. These contacts have variable character that are primarily defined by the type of wall rock. The external contacts are submeter-wide and may show either an abrupt "razor-sharp" meeting of two lithologies, e.g., tonalite and granodiorite, or may display a more transitional, intermingled character, e.g., tonalite and tonalite or tonalite and gabbro. Sharp boundaries suggest the meeting of materials that have significantly different temperatures, rheologies, or compositions, whereas the more transitional sharp contacts indicate smaller differences in the above criteria.

A more detailed description and interpretation of the Morena Reservoir pluton's external contact and adjoining wall rocks is found in Chapter V.

Mafic Enclaves as Magmatic Strain Markers

The magnitude and variation of flattening in mafic enclaves are important to a deformation analysis of the Morena Reservoir pluton in respect to diapirism, ballooning, emplacement, and syn- and post-tectonic regional deformation. As a general rule, the degree of mafic enclave flattening in the pluton and its wall rocks appears to be directly related to the intensity of foliation. The orientations of flattened mafic enclaves across the pluton are nearly identical to that of their host rock's magmatic foliation. The above relations suggests a shared structural origin for foliation and mafic enclave flattening and orientation in the Morena Reservoir pluton.

The foliation map of the pluton (Plate VII) can substitute for a map depicting the orientation patterns of flattened mafic enclaves. Flattened mafic enclaves generally have northwest strikes and inward dips. Mafic enclave orientations are concordant to the pluton's margins and locally wrap around the pluton's core with the foliation.

The mapped variation patterns of zones of equivalent flattening of mafic enclaves are generally concentric to pluton margins and center, and appear to roughly correspond with the first- and second-order variations seen in magnetic susceptibility, enclave abundance, and foliation (Plates IV, V, and VII). Areas of greatest flattening are mostly around the pluton's margins. Zones of minimal flattening are mostly around the core. Contoured morphologic zones appear to be discordant and truncated by the pluton's gabbroic core, the Los Pinos gabbro pluton, and the Hauser Canyon lineament.

Discussion on Mafic Enclave Flattening and Orientations

The structural significance of mafic-enclave flattening is dependent upon the shaping mechanism. If the degree of flattening of mafic enclaves was caused by synplutonic deformation, then the mafic-enclave flattening can be used as strain markers to qualitatively and quantitatively measure emplacement-related shortening of the rocks. The amount and orientation of flattening would reflect the anisotropic stress patterns in the pluton during ascent and emplacement of individual surges or pulses. Evidence for post-plutonic, mafic-enclave flattening would be reflected in solid-state deformation of the host rocks; petrographic analyses generally refute this. The amount of strain recorded by the mafic enclaves indicates mass displacement during magma diapirism (ballooning).

If the variations in mafic-enclave flattening is solely or mostly developed during the period of separation or "calving" from a larger parent xenolith/magma proto-enclave material by intrusion, dissection and dispersion mechanisms of the host tonalite under magmatic flow conditions, then the flattened character is a genetic quality of mafic enclaves and would not be a good shortening indicator. It is possible that the mafic enclave flattening was an evolved combined product of original elongated form and subsequent progressive flattening during ballooning.

If the variation in mafic enclave flattening is the result of forces that acted upon the mafic enclaves after their initial formation (assuming that mafic enclaves form as essentially equant bodies or blobs by a process such as mafic enclave "calving" described above), then the degree of flattening would directly correspond to the magnitude of shortening deformation of the mafic enclave-host tonalite system by hydraulic ballooning forces within the pluton during active pulse injections. In other words, the mafic enclaves were originally roughly equant and were subsequently flattened during submagmatic flow of the mafic enclave-host tonalite system by bulk shortening forces attributed to ballooning of magma. In this model, the amount of flattening can be used as a measure of bulk shortening during ballooning and expansion of the diapir. The mass transfer would be accomplished by radial shortening in the more marginal parts of the pluton during active ballooning of a magmatic pulse into the center of the pluton. Ultimately, mafic enclave flattening information can be applied to the quantitative analysis of mass transfer within the Morena Reservoir pluton and outside the pluton in the wall rocks. The assumption of initially equant mafic enclave bodies prior to deformation is probably poor considering the above discussion on mafic enclave origin. Also, the rheology of the mafic enclaves is most likely different from that of the host magma. Likewise, the amount of flattening in the mafic enclaves would be different from that of the host magma. Assuming the host magma rheology would have lower viscosity and rigidity than the mafic enclaves, the amount of strain recorded in the mafic enclaves would be a minimum value.

The parallelism between host-rock foliation and mafic- enclave orientations, and between host-rock foliation intensity and flattening intensity favors syn-foliation genesis for the development of flattening in mafic enclaves. Therefore, mafic enclave flattening could be used in the general determination of shortening in the Morena Reservoir pluton, shortening that may directly relate to mass transfer mechanisms that facilitated the emplacement of the pluton.

Foliation and Schlieren

Introduction

Foliation is defined by the preferred coplanar orientation of a rock's platy and tabular minerals. The degree of preferred coplanar orientation is the criteria of foliation intensity. Magmatic foliation results from crystal alignments in a "suspended" magmatic laminar flow regime where the long axes of suspended crystals align with the flow vector (least resistance between crystal and flowing liquid), or possibly by preferred orientation of in- situ crystal growth on the walls of magma chamber. Submagmatic foliation is similar to magmatic, except that the magma contains less than 30% melt, i.e., a "matrix supported" magmatic flow regime (Paterson and others, 1989). It may also include some primary mineral deformation, i.e, brittle-strained feldspars, kinked biotite, and crystal-plastic-strained quartz, but with a primary magmatic interstitial assemblage. Magmatic foliation may have a magmatic crenulation (folding of the foliation) that probably originated from magma flow turbulence or disruption (Paterson and others, 1989).

Tectonic foliation is formed by solid-state flow that includes mineral reorientation by crystal-plastic or brittle deformation, and/or recrystallization of a preexisting igneous fabric; most types of crenulation are typically included under tectonic foliation as a supplemental overprint to a preexisting foliation. Solid-state flow in granitoids is commonly illustrated by recrystallized elongate aggregates of quartz and biotite surrounding deformed phenocrysts of plagioclase, hornblende, and quartz (Paterson and others, 1989).

The transition from submagmatic to high temperature solid-state foliation is characterized by more abundant microscopic evidence of crystal-plastic deformation and particularly recrystallization in the interstitial zones. This is typically associated with 1) grain size reduction and elongated aggregate/ribbon forms around deformed phenocrysts; 2) double foliation as either conjugate sets or "S-C" planes; and 3) the presence of late- to postmagmatic mineral overgrowths that appear to erase parts of the original fabric (Paterson and others, 1989).

A crystallizing magma can undergo a progressive foliation development that extends from magmatic to solid-state flow. The systematic growth of "melt-suspended" randomly-oriented crystals may progressively give way to more preferred orientations via magma flow regimes, such as laminar or convergent flow. At some point in time, a crystallizing magma attains a "matrix supported" threshold (30% to 10% melt?) and it becomes "semi-rigid". External forces, i.e., magma surges and pulses, tectonic shortening forces, etc., may act upon the semi-rigid body causing reorganization of crystals into more preferred orientations. During shortening deformation, an interstitial melt might be partly mobilized, or evacuated, i.e., a filter-pressing process. If the temperature in the crystallized part of the magma chamber is near the solidus, the rock can continue to deform via high temperature solid-state flow by crystal-plastic deformation, and a further drop in temperature or pressure into the subsolidus range might include significant brittle deformation.

Plutonic rock foliation may be generated by 1) pure magmatic flow during ascent; 2) synplutonic shortening during diapirism and ballooning; 3) tectonic shortening or shearing during syn-emplacement regional deformation; 4) tectonic shortening during post-emplacement regional deformation; or 5) a combination of the above (Paterson and others, 1989). Foliation processes at work during pluton emplacement are dominantly magmatic flow to submagmatic flow with the possibility of late-stage high temperature solid-state flow. Post-plutonic tectonic foliation processes are dominantly solid-state flow.

A time/space progressive, superposed continuum of the above foliation processes can operate in an evolving diapiric pluton. The changing conditions during the emplacement, and crystallization of the pluton would suggest greater foliation overprinting with increasing distance from the core towards the margins. Older pulses/surges should have more time to acquire subsequent synplutonic overprints in relation to younger ones. This assumes that the oldest magma is found along the pluton's margins, and that post-emplacement deformational events are most intense at the margin.

The ballooning of diapiric plutons is defined here as the upward vertical surging and pulsing of magma from a (relatively narrow?) conduit into a radially dilating middle- to upper-magma chamber, accompanied by radial bulk shortening in the pluton and/or wall rocks. Ballooning diapirism could be considered as one of the principle mechanisms for the generation of zoned foliation fabrics that include an apparent continuation from magmatic flow to submagmatic flow to high temperature solid-state flow in individual rocks.

Episodic pulsing during ballooning may produce concentric zoning of foliation in plutons. Areas of coolest temperatures, greatest crystallization, and highest bulk shortening would correspond to the wall rocks and outer margins of an intruding pulse, where rigid body impingement of crystals against one another is critical. Foliation develops more readily by mechanical compaction of crystals in the presence of a minor melt fraction. Areas of weak magmatic foliation and low bulk shortening would correspond to the interiors of pulses, where the development of foliation is inhibited by higher temperatures, few crystals in the melt, and lack of rigid body crystal-crystal impingement, i.e., hydrostatic conditions during ballooning. If the above mechanism works, then the progressive, concentric zoned development of magmatic to high temperature solid-state foliation in a pluton indicates intraplutonic mass transfer parallel to the plane of the foliation, i.e., horizontal and vertical transfer, respectively. This represents an internal, partial solution to a pluton's evolving "space problem", i.e., making emplacement space via mass transfer within and around a growing diapiric pluton.

Mass transfer within a pluton is locally made available normal to the plane of foliation through shortening. However, the displaced mass must go sideways into the wall rocks or upwards into the roof. This means that either uplift, or outward shortening, or vertical diapir "shadow replacement" downwelling external mass transfer of the local wall rock envelope, or far-field extensional tectonics, must be called upon to accommodate a pluton's diapiric "ballooning" growth (Paterson, 1989).

Megascopic Foliation

The Morena Reservoir pluton has a pervasive magmatic foliation defined in outcrop by the preferred orientations of biotite and hornblende crystals. Elongate plagioclase laths and flattened quartz megacrysts typically show preferred orientations concordant to the aligned mafic crystals. The first-order, pluton-wide, foliation pattern is consistently northwest and parallel to the long axis of the pluton (Plate VII). Foliation is deflected around the gabbroic core and is concordant to the contact with the core body and dips steeply inward. The magmatic foliation dips constitute two elongate, opposing zones that parallel the strike pattern. The boundary between these zones is oriented northwest along the northern margin of Morena Reservoir. The foliation of the southwest zone has moderately steep inward dips. The northeast zone has foliations that dip steeply inward. The magmatic foliation of the external margins of the pluton are concordant with the magmatic foliation of the adjacent margins of neighboring plutons.

The inward-dipping foliation zones, particularly along the pluton's margins, would suggest that the Morena Reservoir pluton has a flattened funnel-shape. Further, the inward dips of the deflected foliation around the core may indicate that the gabbroic core is also funnel-shaped.

The intensity of the magmatic foliation of the pluton appears to have a fundamental zonation pattern that generally matches the lithologic and body design of the pluton (Plate VII).

The intensity of the magmatic foliation is a subjective quantity. However, if a rigorous visual criteria for estimating the intensity is firmly established for an individual observer, then the usefulness for this feature is important in interpreting magmatic flow and deformation in and around the pluton during emplacement and post-emplacement.

The best development of foliation is located near the margins of the Morena Reservoir pluton in narrow zones adjacent the Long Potrero and Los Pinos plutons and in a much broader zone in the northeastern appendage of the pluton. The degree of development of the foliation decreases, with the weakest development found around much of the gabbroic core body. The basic pattern of foliation development is concentric to the body and is zoned inward with highest intensities at the outer margins and lowest intensities in the core.

Parts of the pluton have a magmatic foliation that has been pervasively kinked or folded along one or more planar orientations. Crenulation of the magmatic foliation may be late-magmatic to post-magmatic. Crenulated fabrics occur across a broad region in the eastern and northeastern parts of the pluton. Smaller, isolated regions of the pluton that contain a crenulated foliation occur within narrow zones in the marginal parts of the pluton, and along the eastern side of the gabbroic core body.

Orientations of the long axes of flattened mafic enclaves are generally concordant with the magmatic foliation. The degree of foliation development appears to be directly proportional to both the degree of coplanar orientation of the mafic enclaves, and the degree of flattening of the mafic enclaves. In other terms, the more developed the foliation, the greater coplanar alignment of the mafic enclaves to the foliation, and the greater the flattening ratio of the mafic enclaves.

Microscopic Foliation and Strain Features

Microscopic foliation is inferred to have been predominantly formed by magmatic flow that evolved into a superposed concordant, submagmatic flow with local solid-state flow (Paterson and others, 1989). The crystal-plastic and brittle solid-state deformation observed throughout the pluton cannot be unequivocally characterized as either submagmatic or post-magmatic. However, the general lack of solid-state deformation features in the groundmass suggests submagmatic flow conditions during solid-state deformation of igneous microstructures.

Rolling extinction and fracturing is found in quartz, plagioclase and biotite phenocrysts throughout the pluton, with brittle fracturing in plagioclase, and abundant crystal-plastic strain in quartz and biotite. In-filling of plagioclase fractures with quartz or biotite is found in several rock samples. Generally, interstitial plagioclase, quartz and biotite have little to no indication of strain; there is little to no patchy zoning of either, twin lamellae or oscillatory zoning in the interstitial plagioclase, and little to no undulatory extinction in quartz and biotite.

A post-magmatic, high-temperature solid-state foliation defines a megascopic crenulation and is found in much of the northeastern and eastern marginal rocks of the pluton. The crenulated fabrics consist of semi-pervasive, coplanar, ductile, shear sets, defined principally by recrystallized, elongate, biotite aggregates and phenocryst-size grains that cross the primary magmatic fabric at high angles.

Schlieren

Schlieren are centimeter- to decimeter-scale, planar layering of alternating felsic- and mafic-rich mineral zones of host rock that form larger, typically submeter- to multimeter-sized lensoidal structures that have contacts with the unzoned host rock that range from gradational to sharp. Schlieren may have layering and crosscutting relations seen in clastic sedimentary rocks. Schlieren are typically parallel to subparallel with the host rock's magmatic foliation.

Schlieren are mostly found in the marginal parts of the Morena Reservoir pluton, but also occur around parts of the core. A few isolated occurrences of schlieren are noted in the intermediate parts where foliation intensity is unusually high. The pluton has so few schlieren that it could be characterized as schlieren-free.

Schlieren are a local phenomena and typically associated with well developed magmatic foliation and synplutonic dikes. Two major zones of schlieren occur along the pluton's western margin, and in the pluton's northeastern appendage, particularly between the granodiorite bodies near the south end of Cottonwood Creek. Some schlieren occur around the gabbro core body.

Schlieren appear to constitute heterogeneous zones of that parallel the magmatic foliation. Magnetic susceptibility of schlieren is typically high in the mafic layers and low in the felsic layers. The magnetic susceptibility profile of schlieren tend to be higher than the homogeneous host rock.

The Foliation – Mafic Enclave Relationship

As described in Chapter V, the host rock's magmatic foliation is parallel to flattened mafic enclaves. The magnitude of enclave flattening is typically directly proportional to the intensity of the magmatic foliation. Foliation within mafic enclaves is invariably concordant with the host magmatic foliation.

Foliation and schlieren locally are deflected around the mafic enclaves. Foliation between mafic enclaves that are parts of larger mafic enclave sheet clusters may appear irregular, folded, and disrupted.

Flattening of mafic enclaves appears to be generated by outward radial shortening deformation associated with plutonic ballooning. The fact that mafic enclaves in nearly massive tonalite are close to spheroidal (l:h:w = 1:1:1 to 1½:1½:1), and that progressively more foliated tonalite contains mafic enclaves that are progressively more flattened, suggests that foliation is developed by the same mechanism that flattened the mafic enclaves. The close correspondence between host foliation and the orientation and foliation of the mafic enclaves also support this premise.

It is also possible that part or all of the mafic enclave flattening is mostly original and that mafic preferred orientations of mafic enclaves is due solely to magmatic laminar flow without significant flattening deformation. Alternately, the mafic enclaves' nature (foliation-parallel mafic enclaves that have flattening proportional to the intensity of the host magmatic foliation) may have evolved by a an initial range in mafic enclave shapes that underwent preferred orientation during purely magmatic flow. This was followed by progressively increasing flattening by submagmatic flow under bulk shortening conditions as the more marginal host tonalite-mafic enclave system was systematically compressed by upwelling and ballooning of a magmatic pulse into the pluton's core.

Structural Discontinuities

Introduction

The pluton has at least two major linear to curvilinear structural discontinuities that separate tonalites with distinctly different fabrics. One discontinuity appears to be synplutonic and internal in origin, and the other appears to be produced by a postplutonic, regionally tectonic phenomena.

Internal Structural Boundaries

A structural boundary is mapped discontinuously around the intermediate region of the pluton. The boundary is defined by tonalites with notably different textures, foliations, and mafic enclaves. It is geomorphologically defined by curvilinear valleys found on the northeast and west sides of the gabbro core body. The tonalite outboard of the boundary is finer grained with larger amounts of idiomorphic hornblende phenocrysts, and is generally schlieren-free with foliation that is locally well developed and crenulated. The tonalite inboard the boundary is characterized by centimeter-size aggregates of biotite, moderate to poorly developed noncrenulated foliation, and larger overall grain size.

The Hauser Canyon Lineament

The Hauser Canyon lineament is a regional topographic feature that extends 40 km from the La Posta pluton westward down Hauser Canyon, across Barrett Lake, through Lyon's Valley, across Skyline Truck Trail Drive, and further westward.

The Hauser Canyon lineament coincides with a semi-continuous west-northwest-striking and steeply dipping shear zone, the acute offsets of pluton boundaries, the sharp, and curvilinear discontinuities of contoured field data (Figures 4 and 5; Plates II-VII). These features appear to represent sinistral offset between the southern segment of the Morena Reservoir pluton and its larger northern portion, and between the northern segment of the Long Potrero pluton and its larger southern portion. The La Posta pluton's western margin has a major, left-lateral curve at the eastern terminus of the Hauser canyon lineament, and this may indicate that the Hauser Canyon shear zone was activated to help accommodate the emplacement of the La Posta pluton.

The Hauser Canyon Shear Zone. A curvilinear, 10 to 50 meter-wide, brittle/ductile shear zone is mapped along the bottom of Hauser Canyon through the Morena Reservoir and Long Potrero plutons, and the granodiorite of Barrett Lake (Figures 4 and 5; Plate II). The shear zone is characterized by two sets of anastomosing faults that strike N70W 85SW in the eastern portion of the canyon, and N80W 85SW in the western half. Typically, there are two anastomosed sets of shear planes, with average orientations of N75W 85SW (dominate set), and N20W 80SW (secondary set). Individual faults and close pairs of faults in the shear zone truncate mafic inclusions, and also locally bend and intensely crenulate host foliation, both, with centimeter- to meter-scale left-lateral displacements. Parts of the shear zone have hydrothermal alteration; alteration minerals include potassium feldspar, chlorite and epidote.

The trace of the shear zone is tangential to the northern edge of the Long Potrero pluton core facies. The shear zone extends directly into the contact between the biotite and hornblende-biotite facies, where evidence of the shear zone ends (Figures 4 and 5; Plate II). Here the biotite facies in the contact zone is intensely foliated, variable in texture, and contains abundant schlieren. The shear zone is also located on the west side of the core facies in the hornblende-biotite facies.

Inter- and Intraplutonic Offsets. Sinistral 2 km offsets of the external boundaries of the Morena Reservoir and Long Potrero plutons were mapped in the bottom of Hauser Canyon (Figures 4 and 5; Plate II). The offsets are near the southern terminus of Morena Butte where the Morena Reservoir and Long Potrero plutons meet, and near the southern end of Boneyard Canyon where the Long Potrero tonalite and granodiorite of Barrett Lake are in contact.

There are sinistral offsets in the continuity of the contoured field data of the Morena Reservoir and Long Potrero plutons across the Hauser Canyon lineament. (Figures 4 and 5; Plates II-VI). Foliation intensities and dips do not differ across the lineament, however, poorly to intensely developed crenulation of the magmatic foliation occurs in the rocks adjacent to, or within, the shear zone (Plate VII).

All magnetic susceptibility patterns are either offset across, or terminated at, the lineament. However, the offset patterns do not exactly match up when palinspastically realigned (Plate IV). Several distinctive second-order patterns best define the offset, such as the truncation of the narrow band of very high magnetic susceptibility along the Long Potrero pluton's eastern margin on the south side of the lineament.

A very similar truncation pattern exists for the abrupt termination of mafic-enclave abundance patterns. This is also best seen along the eastern side of the Long Potrero pluton where a narrow, NW-trending band of high mafic- enclave abundance is terminated at the lineament with no recognizable offset pattern match on the other side (Plate V). Patterns of variations in mafic-enclave elongation are also disrupted along the lineament, and show left-lateral offset of distinct patterns (Plate VI). The truncated patterns on each side of the lineament do not match when the left-lateral displacement is removed.

In summary, all noted internal features, especially lithology and the contoured data, provide evidence for left-lateral displacement along the Hauser Canyon lineament. However, the general lack of pattern matches across the lineament (after the left-lateral displacement is removed) and the presence of steeply-plunging lineations in the shear zone indicate that there must also be appreciable dip-slip displacement. The displacement cannot be entirely to dip-slip because the contacts between the plutons are vertical, and yet they are offset 1.5 km. Therefore, the displacement is oblique. The question as to whether the dip-slip is north-side-down or -up remains to be answered.

La Posta Pluton's Western Edge Anomaly. The undeformed western margin of the La Posta pluton has a left-laterally- curved boundary that coincides with the extrapolated eastward continuance of the Hauser Canyon lineament (Figures 4 and 5; Plate II). Pure coincidence for the distinctive 2 km left-lateral curving jog in the pluton's otherwise smooth arcuate western edge at this location seems unlikely.

The La Posta pluton has no evidence of deformation in either its marginal or interior rocks. If displacement associated with the Hauser Canyon lineament did occur in the La Posta pluton, it was then still fluid. This places an uppermost age limit on displacement at no younger than the La Posta's 94 Ma U-Pb zircon age.

Discussion

The offsets within the Long Potrero and Morena Reservoir plutons places a general lower age of displacement at 101 Ma +\- 2 Ma. No evidence is found for synplutonic deformation, and in fact, the concentric foliation pattern in the Long Potrero pluton, as seen on opposite sides of Hauser Canyon, do not appear deflected into the strike of the Hauser Canyon lineament. The local, consistent northwest-striking foliation strikes of the Morena Reservoir pluton is not disturbed nor deflected across the mouth of Hauser Canyon. Therefore, the oldest age for initial displacement would be at or following the time at which the involved outer portions of the Long Potrero and Morena Reservoir plutons were fully crystallized, probably after 101 Ma. However, the apparent lack of evidence for shearing through the extreme northern edge of the Long Potrero pluton's biotite facies might suggest the possibility for late-plutonic fault/magma motions, in respect to the emplacement of the Long Potrero's core facies (apparently younger than the 101 Ma U-Pb age for the Morena Reservoir tonalite).

In conclusion, the apparent displacement along this section of the Hauser Canyon lineament appears to be well constrained between 101 Ma and 94 Ma, based upon U-Pb zircon ages, and the appearance of sharp, offsetting structural discontinuities in the outer portions of the Long Potrero and Morena Reservoir plutons. There exists the possibility that displacement was concurrent with the magmatic pulse of the Long Potrero pluton's silicic core.

The timing of displacement and the regional scale of the Hauser Canyon shear zone suggests that this fault zone helped accommodate the emplacement of the 1400 km^2 La Posta pluton. The Cuyamaca-Laguna Mountain Shear Zone (CLMSZ) has documented 118-114 Ma contractional deformation and post-105 Ma/pre-94 Ma extensional deformation (Girty and others, 1994; Thompson and others, 1994). The timing of the extensional deformation in the CLMSZ closely matches that of the displacement in the Hauser Canyon shear zone. Both structures are truncated by the La Posta pluton.

CHAPTER V

METAMORPHIC AND PLUTONIC ROCKS SURROUNDING THE MORENA RESERVOIR PLUTON

Introduction

The Morena Reservoir pluton is mostly surrounded by plutons that range in composition from gabbro to granodiorite. Along a small part of its northern margin is metasedimentary rock. Several of the plutons, including the Morena Reservoir pluton, appear to be a part of a large, concentrically nested complex that is informally termed by the author as the Hauser complex, after Hauser Canyon which cuts the middle of the complex). The plutons that make up the Hauser complex include the gabbro of Los Pinos Peak, Potrero Peak, Lawson Peak, and include the gabbro in the center of Morena Reservoir pluton; the tonalite of Morena Reservoir and Long Potrero; and the granodiorite of Corte Madera and Mother Grundy Peak, which include the rocks that underlie Morena Butte and The Narrows. The core of the Hauser complex is dominated by the tonalite composition plutons. The tonalite core is mantled by relatively small, discreet gabbro bodies. Forming the outermost part of the complex is extensive, massive granodiorite.

The structural relations between the various units of the complex are generally concordant, in terms of pluton elongation, foliation and mafic enclave orientations and contact character (Figures 4 and 5; Plate II). The first order structure pattern forms concentric shells from the margins to the center of the complex. This results from the nested arrangement of the plutons and much of their foliation patterns. The second-order structure pattern forms a northwest trend. This is formed by the pluton shape and internal units, and foliation. The La Posta pluton appears to truncate the eastern side of the Hauser complex.

The intrusive sequence within the Hauser complex appears to be gabbro-tonalite-granodiorite. However, there are exceptions to this order, particularly between tonalite and granodiorite.

The Morena Metasedimentary Screen

The Morena metasedimentary screen (TJ_{ms}) crops out north of and adjacent to the Morena Reservoir pluton (Figures 4 and 5; Plate II). This unit mainly consists of amphibolite grade quartzo-feldspathic schist that is very similar in appearance to the extensive, Triassic(?)/ Jurassic Julian Schist found further north near Julian.

For a more detailed description of this unit refer to (Bergreen and Walawender, 1977).

Geology of the Gabbros

The Gabbro Suite of Los Pinos

The gabbro suite of the Los Pinos pluton, exposed over 10 km_2 (K_{lop}) is in contact with the northwestern margin of the Morena Reservoir pluton (Figures 4 and 5; Plate II). The elliptical-shaped Los Pinos pluton is west-northwest elongated. It consists mainly of fine- to medium-grained, hypidiomorphic granular olivine-pyroxene gabbronorite and amphibole gabbro, with lesser amounts of anorthositic gabbro and hornblende pegmatite (Walawender, 1976).

The Gabbroic Suite of Morena Reservoir

A kidney-shaped, 3 km^2, lithologically heterogeneous, moderately magnetic gabbroic body occurs within the core of the Morena Reservoir pluton (Figures 4 and 5; Plate II). The lithologies and structure of the gabbro of Morena Reservoir are very similar to those of the gabbroic Los Pinos pluton (K_{lop}); these two bodies most likely share a common history. The gabbroic core body consists of northwest-trending multi-intrusions of fine- to medium-grained, hypidiomorphic granular anorthositic gabbro and hornblende gabbronorite, hornblende gabbro pegmatite. The above rocks are intruded by abundant granite pegmatite dikes. Foliations strike northwest and dip 80° to the south. Much of the hornblende gabbronorite appears to have undergone hydrothermal alteration. Abundant tourmaline-bearing, granite pegmatite dikes and pods intrude the contact between the gabbro and the surrounding tonalite. Although granite pegmatite dikes crosscut both rock types, the dikes appear to be larger and more abundant in the gabbro. Two samples from this unit have been analyzed for geochemistry and are discussed in the geochemistry section.

The Gabbro of Morena Butte

Several small gabbro bodies (K_{1op}) are located between the Long Potrero pluton and the granodiorite of Morena Butte (Figures 4 and 5; Plate II). These bodies have similar compositions and fabrics to that of the gabbro suites of Los Pinos and Morena Reservoir, and consist of poorly- to moderately foliated, medium- and fine-grained, magnetite-rich, hornblende pyroxene gabbro. Rock textures in single outcrops are highly heterogeneous and this suggests multiple-pulse intrusions of magmas emplaced successively into one another. Near the southern end of the Morena Butte body, a large, 10- to 40 meter-wide, northeast-striking granite pegmatite dike crosscuts the more southern gabbro body, the Morena Butte body, and the Long Potrero pluton. The dike is widest in the gabbro body and it appears to pinch out in the Morena Butte body to the north.

Geology of the Tonalites

The Long Potrero Pluton

The 103 Ma Long Potrero pluton is a transitional-type 140-km^2 pluton with an irregular elongate elliptical shaped, zoned, possibly composite, tonalite body. The pluton's form and internal zoned lithologies are concentric to its center and elongated along a northwest trend (Figures 4 and 5; Plate II). Previous studies on the Long Potrero have been made by Hoppler (1983) and Rector (1989 and 1993). Most of the following information on the Long Potrero pluton is from research conducted by the author, and includes over 800 mapped localities with detailed field data similar to that collected for the Morena Reservoir pluton.

Lithologies. The Long Potrero pluton is gradationally zoned inward from a mafic biotite-hornblende tonalite in the margins, through a hornblende-biotite tonalite in the intermediate part, into a biotite leucotonalite in the southwestern part of the core, and into a muscovite-biotite leucotrondhjemite in the greater central part of the core, and a garnet-muscovite-biotite leucotrondhjemite in the north-eastern side of the core. Pods of comagmatic early-plutonic mafic pyroxene-biotite-hornblende quartz diorite and late-plutonic aplitic muscovite-biotite leucotrond- jhemite have sharp contacts. The mafic pods occur near the pluton's margins, and the aplitic pods crop out along a northwestern trend within the core. Potassium feldspar occurs in all the rock types with stained-slab modes of 1% to 3%, except for the potassium feldspar-free, pyroxene-bearing unit.

Hoppler (1983) mapped similar units with similar boundaries and petrographic descriptions in the Long Potrero pluton, but he did not recognize the pyroxene-bearing and garnet-bearing units. However, Hoppler (1983) does mention the presence of pyroxene in the biotite-hornblende tonalite.

Color Index. The contoured color index of the Long Potrero pluton consist of concentric zones that are concordant to the margins and centered about the core. Color index grades from 36% at the margins to 5% in the core (Figure 38). Anomalous changes in the variation of the color index are noted within the biotite-hornblende unit, along the biotite-hornblende/hornblende-biotite boundary, and at the hornblende-biotite/biotite contact. Color index patterns appear to be offset along Hauser Canyon.

Figure 37. Contour map depicting the color index of the Long Potrero pluton. The ten grey-scale color contours each have a color index interval of 3, and systematically darken from interval 5-8 (white) to interval 32-35 (black).

Textural Variations. The texture of the tonalite is medium- to coarse-grained with variably foliated fabrics, and is concentrically zoned in crystal size. It systema- tically grades inward from fine-grained and medium-grained in the margins and medium-grained in the intermediate part of the pluton, to a coarse-grained core, except for the fine-grained aplitic unit (Figure 37). The textural patterns are concentric about the center of the pluton and are concordant to the lithologic zoning. However, there are several concentric boundaries within the pluton that are defined by rather abrupt changes (tightly gradational; not sharp) in texture are noted within the biotite-hornblende unit, between the biotite-hornblende and hornblende-biotite units, along much of the pluton's core margin, around the pyroxene-biotite-hornblende and aplitic These structural boundaries are co-defined by other field data such as lithology, foliation, magnetic susceptibility, and mafic enclave character.

Figure 38. Contour map defining the textural variation of the Long Potrero pluton. Blues to greens to yellow = progressive coarsening of medium-grained rocks; orange to reds = progressive coarsening of coarse-grained rocks.

Foliation. The pluton has a pervasive megascopic magmatic foliation concordant to its margins and concentric about its center (Plates II and VII and Figure 39). South of the Hauser Canyon lineament, the foliation dips moderately to steeply to the northeast. The foliation dips units, and along the Hauser Canyon lineament. inward in the southwest half of the pluton and outward-dipping in the northeast half. Vertical dips occur along an inflection line separating the northeast and southwest halves. North of the Hauser Canyon lineament the foliation dips moderately to steeply south in much of the biotite-hornblende unit, and moderately to steeply north in the innermost part of the biotite-hornblende unit and the hornblende-biotite unit.

The foliation intensity is greatest near its margins and least in its core. A second-order scale of foliation intensity consists of a zone of intensity around the core and concentric zones of high and low intensities within the biotite-hornblende unit and roughly between the biotite-hornblende and hornblende-biotite units. A magmatic crenulation, defined by synmagmatic folding of the primary foliation, occurs within most of the pluton's margin and along several internal structural boundaries. Coherent zones of foliation are divided by structural discontinuities defined in part by abrupt changes in foliation dip and/or intensity/crenulation. The major boundaries are located within the biotite-hornblende unit, around the pyroxene-biotite-hornblende unit pods, close to the biotite-hornblende/hornblende-biotite contact, around the core facies, circling the aplitic unit pods, and along the Hauser Canyon lineament.

Foliation field data reported by Hoppler (1983) confirm the strike and dip measurements made by the author in the Long Potrero pluton.

Figure 39. Foliation map of the Long Potrero pluton. The six rainbow color contours represent foliation intensity and are coded red (strong), orange (weakly strong), yellow (moderate), green (weakly moderate), blue (weak), and violet (very weak).

Magnetic Susceptibility. Magnetic susceptibility is concentrically zoned in the pluton with highest values in the margin systematically decreasing to lowest values in the core and eastern side of the pluton (Figure 40). Volume percent magnetite distribution is deduced from the magnetic susceptibility. Rocks with relatively high susceptibility are located along the western, southwestern and southern margins of the pluton and also along an arcuate zone near the biotite-hornblende/hornblende-biotite contact on the north side of the Hauser Canyon lineament. Rocks with very low values occur along the eastern, northeastern and northern margins of the pluton and in the core. Concordant zones of either high, moderate, or low values are separated by abrupt changes in MS.

Figure 40. Contour map depicting the magnetic suscepti-bility of the Long Potrero pluton. The ten rainbow color contours represent both magnetic susceptibility and magnetite volume percent. The contour intervals reds and orange = high MS and magnetite content (up to 5000 x 10^{-5} SI/1.4 volume%); yellow and greens = moderate MS and magnetite (up to 1300 x 10^{-5} SI/0.46 volume%); blues and violet = low MS and magnetite (up to 300 x 10^{-5} SI/0.14 volume%).

Mafic Enclaves. Mafic enclaves are found throughout the Long Potrero pluton and vary systematically in abundance, morphology, size, and grouping characteristics, both on outcrop-scale and pluton-scale. Pluton-scale contouring of mafic enclave abundance and morphology displays patterns that are concentric to the margins and center of the pluton (Figures 41 and 42, respectively). Mafic enclave abundance is progressively zoned from greatest numbers along the pluton's margins to few in the core. Mafic enclave flattening is most pronounced near the pluton's margins and systematically decreases to zero in the core. However, an anomalous arcuate zone of strong flattening is located along the inner margin of the hornblende-biotite where the biotite unit is very narrow. Abrupt changes in mafic enclave abundance and elongation are noted along the same boundaries as described above for the other field data sets.

Figure 41. Contour map depicting the mafic enclave distribution of the Long Potrero pluton. The ten rainbow color contours represent zones of similar abundance (the average number of mafic enclaves per 9 m^2 of outcrop). The contour intervals reds and orange = high abundance (up to 400 per 9 m^2); yellow and greens = moderate abundance (up to 99 per 9 m^2); blues and violet = low abundance (up to 14 per 9 m^2).

Figure 42. Contour map depicting the mafic enclave morphology in the Long Potrero pluton. The seven rainbow color contours represent the average degree of flattening (expressed as a width to length ratio of the enclave). The color contour intervals reds and orange = high degree of flattening (up to 1:20); yellow and green = moderate degrees of flattening (up to 1:5½); blue and violet = low degrees of flattening (up to 2½).

Structural Discontinuities. Several structural discontinuities are present between, and within the lithologic units, and are defined in part by abrupt changes in lithologic gradation. Major breaks normal to the gradual lithologic zoning are located within the biotite-hornblende unit, around the pyroxene-biotite-hornblende unit pods, along the inner margin of the biotite-hornblende unit, between the hornblende-biotite and biotite units, between the biotite and muscovite-biotite units, and along the Hauser Canyon lineament. These boundaries suggest both magmatic and tectonic dislocations.

Geochemistry. Major, trace, and rare earth element geochemical analyses of the Long Potrero pluton (Appendix B, Table 5) indicate a comagmatic suite of rocks that originated from several geochemically distinct pulses; each pulse underwent limited crystallization differentiation at the present erosional level. Major element Harker diagrams have smooth curvilinear to linear trends except for K_2O (Figures 43 and 44). This diverse suite of rocks has major element spectras that overlap the entire respective spectras of the combined La Posta-type plutons, e.g., a SiO_2 range from 57 wt% to 77 wt% (Walawender and others, 1990).

Geochemical discrimination diagrams show the Long Potrero suite plotting tightly with the Morena Reservoir suite with the exception of the Long Potrero pluton's leucotrondhjemite core rocks (Figure 45).

Trace element Harker diagrams of the Long Potrero suite have trends that range from smooth and tight to irregular and dispersed, with distinctive groupings according to rock type (Figures 46 and 47).

Tonalite REE patterns are generally similar to those of the Morena Reservoir pluton, and can be organized into three distinctive groups that appear to be semi-independent of rock-type (Figure 48).

Two mafic enclave samples have irregular major and trace element patterns with that of the host tonalites (Figures 43-47). However, the mafic enclave REE signatures are virtually identical with that of several host rock samples; this is good supporting evidence for a cognate origin (Figure 48).

An examination of the combined geochemical data from the Long Potrero and Morena Reservoir plutons show tight correspondence in most of the major elements and several trace elements, but significantly divergence in other trace elements (Figures 43-44 and 46-47).

Summary and Discussion. In summary, the lithology, texture, foliation, magnetic susceptibility, and mafic enclave abundance and morphology variations all form patterns concordant to the margins and concentric about the center of the pluton. Several correlative abrupt pattern changes displayed by the contoured data define several concentric magmatic/structural boundaries that are inferred to be intra-plutonic pulse septa. These septa divide the pluton into four major pulse units and two minor ones (Figure 49). The pulses appear to have intruded in a sequential nested pattern, most likely through a central conduit; the oldest formed the pluton's margins and the youngest formed the core.

Hoppler (1983) draws a similar conclusion on the pluton's emplacement by suggesting that the diapir's differentiated felsic tail intruded its mafic head, i.e., a silica-rich pulse intruded an older, comagmatic mafic
rich pulse.

The similarities in geochemistry, particularly in the major- and rare-earth-elements, between the hornblende-bearing tonalite rocks of the Long Potrero and Morena Reservoir plutons indicate close petrogenetic ties in respect to source regions and parental melts. However, the differences in some major element and several trace element trends indicates independent magmatic systems and crystallization histories.

Figure 49. Structure map depicting the inferred sequence of pulses comprising the Long Potrero pluton. The six color zones represent the sequential pulses. Red = first pulse; orange = second pulse; magenta = third pulse; green = fourth pulse; yellow = fifth pulse; and violet = last pulse.

The La Posta Pluton

The 94 Ma La Posta pluton is an irregular elliptical-shaped, 1400 sq km, nonmagnetic, lithologically and concentrically zoned tonalite/granodiorite pluton to the east of the Morena Reservoir and Long Potrero plutons (Figures 4 and 5; Plate II). This body is the type-pluton for the large, eastern-zone, La Posta-type plutons of the Peninsular Ranges batholith (Walawender and others, 1990).

The pluton's lithology grades inward from a sphene-bearing, hornblende-biotite tonalite in the margins into a hornblende-biotite granodiorite, into biotite granodiorite into a two-mica granodiorite core. A north-trending, prebatholithic, amphibolite-grade, metasedimentary screen divides the pluton in half. The western half of the pluton appears to truncate the southern end of the regional Cuyamaca-Laguna Mountain Shear Zone (CLMSZ).

The western, marginal facies of the La Posta pluton in contact with the eastern edge of the Morena Reservoir and Long Potrero plutons consists of either, a homogeneous, weakly foliated, medium-grain, hypidiomorphic granular, nonmagnetic, sphene-bearing, hornblende-biotite tonalite, or a "border zone" of gradationally interleaved, alternating medium-grained, melanocratic and leucocratic tonalite.

For detailed descriptions and interpretations of its petrology, structure, emplacement history and petrogenesis, see Clinkenbeard (1987), and Walawender and others (1990).

Geology of the Granodiorites

The Granodiorite of Barrett

The granodiorite of Barrett (K_b) is typically coarse-grained, hypidiomorphic granular to moderately foliated (magmatic foliation), moderately magnetic, and schlieren-rich (Figures 4 and 5; Plate II). It contains 0.5 cm to 2.0 cm potassium feldspar phenocrysts, up to 10 percent biotite and 4 percent hornblende as coarse-grained phenocrysts, and variable abundances of mafic enclaves. This unit is equivalent to the monzogranite of Mother Grundy Peak of V.R.Todd (oral communication, 1994). The fresh rock is salmon pink to gray in color and weathers to a tan-orange color. It is uncertain whether the granodiorite of Barrett is a pluton or a composite body.

In many places between the International border and the Long Potrero pluton, the granodiorite of Barrett contains abundant massive xenoliths of fine- to medium-grained gabbro (1x1 m to 40x10 m); the host granodiorite is either coarse-grained up to the sharp contact with the xenoliths, or the granodiorite has a finer-grained texture within a mingled/transitional contact with the mafic xenoliths. The southern part of the granodiorite due south of the Long Potrero pluton contains more mafic enclaves and synplutonic dikes than does the granodiorite due west of the Long Potrero pluton.

Both southern and western parts of the granodiorite of Barrett are notable for their interior facies which feature 1- to 2-cm potassium feldspar phenocrysts, very coarse-grained, schlieren-rich, moderate foliation and magnetic susceptibility. This interior facies changes drastically to a contact facies that consists of a virtually non-mafic, well foliated, nonmagnetic, fine-grained, recrystallized? aplitic rock.

The interior facies of the 1 km-long, northeast-trending granodiorite body near Canyon City (K_b) is a moderately to well foliated, nonmagnetic, fine- to medium-grained leucogranodiorite (Figures 4 and 5; Plate II). Rocks in the margins of the body have a fine-grained aplitic texture with a partly to wholly recrystallized microscopic "triple-point" texture. In places, the rocks within 10 m of the contact contain a large number of 5- to 20 cm, highly attenuated, biotite-rich enclaves.

The Granodiorite of Corte Madera

The granodiorite of Corte Madera (K_{cm}) (V.R. Todd, oral communications, 1993) adjoins the northeastern margin of the Morena Reservoir pluton and underlie Morena Butte (Figures 4 and 5; Plate II). These rocks are the equivalent of the granodiorite of Corte Madera of V.R. Todd (oral communication, 1993). Gina Corolla (1993) studied the geochemistry of several samples of this rock type including samples from Corte Madera, Morena Butte, and from the granodiorite adjacent to the Morena Reservoir pluton's northern margin. The granodiorite of Corte Madera was not examined in the field during this study and the contact relation between it and the far northwestern corner of the Morena Reservoir pluton is presently uncertain. Descriptions of this unit are from V.R. Todd (1983).

The Granodiorite of Morena Butte

The granodiorite underlying Morena Butte (K_{cm}) is a 3 km-long, elongated, lens-shaped, northwest-trending pluton, between the Morena Reservoir and Long Potrero plutons (Figures 4 and 5; Plate II). The Morena Butte pluton appears to bend eastward at its southern end and pinches out on the north side of Hauser Canyon; no granodiorite is found on the south side of the canyon. Several small gabbro bodies are found along its southern margin. Also noted along the contact between the Morena Butte body and the Long Potrero pluton, are 10- to 40 m-wide vertical sheets of strongly foliated, partially melted(?), fine- to medium-grained, garnet-bearing, biotite-rich quartzofeldspathic gneiss. These metamorphic rocks, which may represent highly metamorphosed roof pendants, are similar in appearance and composition to a much larger body located west of Lawson Peak and north of Lyons Peak, 15 km west northwest of Morena Butte in the Barrett Lake 7.5 quadrangle. Interestingly, both bodies are found between granodiorite on their northern side and the Hauser Canyon lineament to the south.

The granodiorite of Morena Butte consists of a intensely foliated, partly recrystallized, nonmagnetic to moderately magnetic, fine-grained, biotite leucogranodiorite marginal facies (5 m to 20 m wide). The interior facies is a poorly foliated to moderately foliated, moderately magnetic, coarse-grained, K-feldspar megacryst-rich, pyroxene-bearing biotite granodiorite.

The granodiorite of Morena Butte appears to be an isolated southeastern extension of the much more extensive Corte Madera pluton.

The granodiorite pluton that include the granodiorites of Corte Madera and Morena Butte form a discontinuous northwest-trending mantle around the northeastern and southwestern flanks of the Morena Reservoir pluton, including the Los Pinos pluton. Interestingly, the granodiorite of The Narrows, which forms a structural septa between the eastern margin of the Morena Reservoir pluton and the western margin of the La Posta pluton, is north-trending.

The Granodiorite North of Morena Valley

The granodiorite underlying a part of the highlands north of Morena Valley is a west-northwest-trending, irregular-shaped body in contact with the Morena Reservoir pluton and the Morena metasedimentary screen (Figures 4 and 5; Plate II). This body consists of a weakly foliated, coarse-grained, biotite granodiorite that contains magnetite and very coarse-grained potassium feldspar megacrysts. The granodiorite is intensely foliated along its margins, particularly adjacent the Morena Reservoir pluton. Foliation appears locally concordant with foliation of the Morena Reservoir pluton. In general, this body has many similarities with the rocks underlying Morena Butte.

The Granodiorite of The Narrows

The granodiorite underlying The Narrows, and to the north and south, is a north-elongated body located between the Morena Reservoir and La Posta plutons (Figures 4 and 5; Plate II). It is a fine- to coarse-grained, hypidiomorphic granular, biotite granodiorite that has a lithology very similar to the granodiorite underlying the highlands north of Morena Valley, and the granodiorites underlying Corte Madera and Morena Butte. It is rich in magnetite and appears to be an eastern extension of the granodiorite body that underlies the highlands north of Morena Valley.

Interplutonic Contact Relations

The Contact Between the Los Pinos Gabbro and the Morena Reservoir and Long Potrero Plutons

The southeastern margin of the Los Pinos gabbro pluton is in sharp contact with a small part of the northern margin of the Long Potrero pluton and the western margin of the Morena Reservoir pluton (Figures 4 and 5; Plate II). In several places the contact between the Los Pinos gabbro pluton and the Morena Reservoir tonalite pluton can be seen, and in one place, the three bodies form a "triple junction" at the southeastern end of the Los Pinos pluton. Outcrops of the Los Pinos pluton/Long Potrero pluton contact are scarce, outcrops exhibiting the Los Pinos pluton/Morena Reservoir pluton contact are more numerous.

Within the contact zone, the Los Pinos pluton has an intensely crenulated magmatic fabric and contains anastomosing, extremely magnetite-rich, sinuous, ropey to lumpy appearing, light-colored, fine-grained hornblende diorite(?)/gabbro(?) dikes; these dikes form sharp, but irregular, contacts within the gabbro. Rock similar in composition, both as dikes and as synplutonic massive bodies are found in the adjoining Morena Reservoir tonalite.

The Morena Reservoir tonalite near the contact consists of intensely foliated, medium-grained, hypidiomorphic equigranular, very highly magnetic ($8,000 \times 10^{-5}$ SI), mafic-rich biotite-hornblende tonalite and a highly magnetic ($4,000 \times 10^{-5}$ SI), light-colored, medium-grained synplutonic biotite-hornblende tonalite that form an intermingled synplutonic fabric. Both rock types are crosscut by numerous slightly magnetite-bearing, dark- and highly magnetic, light-colored synplutonic dikes. Spatially associated mafic enclave sheets in the tonalite rocks appear to be fragmented dark dike material.

The contact zone between the Morena Reservoir tonalite and Los Pinos gabbro is dominated by massive highly magnetic, fine-grained, relatively light-colored, hornblende diorite/gabbro rocks that include sheet-like, dense clusters of relatively dark-colored enclaves. The light-colored diorite(?)/gabbro(?) dikes in the gabbro and the local Morena Reservoir tonalite host rock are extremely rich in magnetite (1% by volume), whereas the local host gabbro rock including its hornblende pegmatite, and the dark synplutonic dikes within the Morena Reservoir tonalite host rock contain small amounts of magnetite (less than 0.1% by volume).

Cross-cutting relations suggest that the very highly magnetic, light-colored, fine-grained ropey dikes and masses within the gabbro body are the medium-grained equivalent of the magnetite-rich, medium-grained Morena Reservoir tonalite; this would indicate that the tonalite intruded the gabbro. This contact displays fabric and structure that suggest near equilibrium thermal/rheologic conditions between the tonalite and gabbroic rocks.

The Morena Reservoir Gabbroic Core Contact

The Morena Reservoir tonalite nearest the contact with the gabbro core is notably mafic, intensely foliated, and contains magnetite-rich synplutonic dikes; these field relations are similar to that observed along the Los Pinos pluton/Morena Reservoir pluton contact.

Lack of crosscutting evidence between the tonalite and gabbro bodies and similarities in rocks of the Los Pinos pluton and Morena Reservoir gabbro core suggest that both gabbro bodies where a single body and were subsequently isolated by the emplacement of the Morena Reservoir pluton.

The Contact between the Long Potreroand Morena Reservoir Plutons

Exposures of contacts between these tonalite plutons are numerous, but many of the interplutonic tonalite contacts are difficult to interpret. The contact between the Long Potrero and Morena Reservoir plutons is fairly sharp and regular, and has a convex-to-the-northeast curvalinear shape. Two significant irregularities along this contact are the Morena Butte granodiorite and the offset along the Hauser Canyon lineament. The contact zone between the tonalite plutons from the southern end of Morena Butte southward includes 1) 5- to 20 m-wide, inter-leaved sheets of fine- and medium-grained, dark tonalite gneiss (mylonite?); 2) very intensely foliated and crenulated, medium-grained tonalite; and 3) contact-parallel, 1- to 2 m-wide, leucocratic tonalite/granodiorite dikes. The magnetic susceptibility variation in the contact zone ranges from magnetite-free (0-30 x 10^{-5} SI) to magnetite-rich (8,000 x 10^{-5} SI) and the tonalite and leucocratic dikes are vary moderately in magnetite content.

The tonalite on opposite sides of the contact are mineralogically very similar, but have significant local textural and structural differences. The Long Potrero tonalite has a much more deformed character than the Morena Reservoir tonalite. Typically, the rocks of the Long Potrero pluton have a coarse medium-grained, hypidiomorphic granular texture with very coarse-grained clots of biotite. The Long Potrero rocks have a pervasively intense magmatic foliation with intense to extreme, cm-scale, conjugate-set magmatic(?) intense folding of the foliation. Meter-scale, cross-cutting, shear planes in the rock indicate sinistral oblique east-over-west displacement. The character of the foliation suggests an overlapping of solid-state flow on a magmatic flow. Mafic enclave abundance and flattening is highly variable. Within 50-m of the Long Potrero pluton adjacent the contact zone, are dikes with 1) fine-grained, magnetite-free, leucocratic late-plutonic rock; 2) magnetite-bearing, mesocratic, fine-grained synplutonic rock; and 3) highly magnetite, melenocratic, medium- to fine-grained, late-plutonic tonalite rock.

The Morena Reservoir tonalite adjacent to the contact has a variably developed magmatic foliation with magmatic crenulation and a medium- to coarse medium-grained, hypidiomorphic granular texture. These rocks contain abundant hornblende, idiomorphic sphene and little to no magnetite. Isolated, 1- to 10 m-long, magmatically inter- min-gled bodies of poorly foliated, medium-grained, hypidiomorphic granular, cumulate-textured(?), magnetite-rich mafic tonalite, with very coarse-grained, idiomorphic hornblende phenocrysts, grade into highly elongated, m-scale mafic inclusions and schlieren. Numerous granite pegmatite dikes intrude the Morena Reservoir tonalite near the contact.

North of Morena Butte, and roughly following a concave-to-the-east, north-south trend, the two tonalite plutons share a contact that is missing the gneissic (mylonite?) sheet structures. The differences in texture and structure across the contact are significantly less than that seen along the southern contact noted above. The tonalite of the Long Potrero pluton adjacent to the contact is moderately to intensely foliated and crenulated with medium-grained, hypidiomorphic granular texture. The rock is magnetite-bearing and contains strongly flattened mafic enclaves. The Morena Reservoir tonalite has poorly to moderately developed foliation with no crenulation and medium-grained, hypidiomorphic granular texture; it contains low abundances of slightly flattened mafic enclaves and is magnetite-free. The contact is marked by 5 to 10 m-wide, dense clusters of magnetite-rich mafic enclaves on the Long Potrero side. Well exposed outcrops of the tonalite-tonalite contact are located between the northwestern end of Morena Butte and the southeastern end of the Los Pinos gabbro.

The inferred contact along the Long Potrero pluton's southeastern margin is drawn through an area (hundreds of meters wide) that consists of numerous tonalite facies and synplutonic contacts. Here a curvilinear boundary is mapped that is defined by the roughly overlapping traces of abrupt changes in lithology, texture, foliation, mafic enclave abundance and morphology, and magnetic susceptibility.

The Morena Reservoir Pluton's Southernmost Contact

Tonalite near the Morena Reservoir pluton's inferred southernmost limit might be either a narrow southern extension of the Morena Reservoir pluton, or most likely, these rocks represent an older, outermost pulse of the Long Potrero pluton. Or, this might be another tonalite pluton.

The Contact Between the Morena Reservoir and La Posta Plutons

The La Posta pluton appears to have intruded both the Morena Reservoir tonalite and a small granodiorite body found between the La Posta pluton and the Morena Reservoir pluton. The contact appears to be covered by alluvial-filled valleys. There is however, an excellent exposure of the contact between the La Posta pluton and the inferred northern appendage of the Morena Reservoir pluton in a large roadcut along Interstate 8 at Cameron Valley. Here, the La Posta pluton intrudes the magnetic western tonalite with a dark-/light-colored, meter-scale layered rock.

The Contact Surrounding the Granodiorite of Morena Butte

The contact between the granodiorite of Morena Butte and the Morena Reservoir and Long Potrero plutons and adjoining gabbros are presently uncertain. Inclusions of gabbro are found in the Morena Butte marginal facies near the contact. Dark dikes (tonalite?) are also noted in several localities in the Morena Butte body. Granodiorite xenoliths are not observed in the tonalites or gabbros, but, biotite quartzofeldspathic gneiss/schist (metasedimentary(?) or meta-igneous(?)) inclusions are found in parts of the Morena Reservoir tonalite near the Morena Butte body. Enclaves in the gabbros and Long Potrero tonalite along the southern margin of the Morena Butte body are mafic consisting of plagioclase and hornblende, with some pyroxene.

The best exposed contacts are located at the southeast end of the Morena Butte pluton, and along the base of the near vertical southwestern edge of Morena Butte, where it is adjoined by the Long Potrero pluton. Here the granodiorite is intensely foliated, fine- to coarse-grained with recrystallized fabrics, nonmagnetic to weakly magnetic, biotite-bearing and leucocratic. At the contact, the granodiorite has localized zones of densely-spaced, 10- to 20 cm-long, biotite-rich enclaves.

CHAPTER VI

PETROGENETIC AND EMPLACEMENT MODELING OF THE MORENA RESERVOIR PLUTON

Introduction

The generation and emplacement of the Morena Reservoir pluton coincides with the transition between older western- zone and younger eastern-zone magmatism. The subduction-related, tectonic setting of the Morena Reservoir pluton was likely due to initiation of a major plate reorganization between the Farallon and North American plate. This tectonic transition is characterized by high-rate, strong dextral-sense convergence, during the height of the proposed mid-proto-Pacific superplume around 105 Ma (Scotese and others, 1988; Larson, 1991). The tonalitic magmas generated at this time have mineralogic and major-, trace-, and rare-earth-element signatures that are similar to the older western-type tonalites. However, compared to the older western-type tonalites, the younger tonalites have more euhedral hornblende and megacrystic quartz in an overall coarser-grained texture within the more differentiated parts of the pluton.

The local transitional-type tonalites, e.g., the Morena Reservoir and Long Potrero plutons, have a very distinctive submagmatic, emplacement-related, microfabric deformation which is attributed to their diapiric style of emplacement. The transitional-type tonalite plutons have euhedral sphene and pluton-scale zoning of magnetite. They also have slightly higher Sr contents, and REE patterns with smaller or absent negative Eu anomalies; features that partly approach that of the younger, eastern-type tonalites. Closely following the transitional tonalite magmatism was the eastern-type magmatism of the 100 to 90 Ma La Posta-type plutons, magmatism that is very distinct from the older western zone magmatism. Emplacement of the 101 Ma Morena Reservoir pluton coincides with the location and timing of regional extensional shearing along the axial zone of the batholith in the Cuyamaca-Laguna Mountain Shear Zone, a deformational event pre-dating 105 Ma and post-dating 94 Ma (Girty and others, 1994).

A petrogenetic model of the Morena Reservoir pluton must take into account the field relationships, petrographic nature, mineralogic and geochemical variations, and isotopic age relations of the rocks, along with the above regional tectonic and structural setting. However, the limiting of data from a roughly horizontal cross section of a moving magma body that may be 5 to 10 km in length makes petrogenetic and emplacement modelling of an abstract exercise. A general plutonic model will be presented below for the Morena Reservoir pluton from the interpretations of the limited data.

Petrogenesis

A relatively simple crystallization fractionation model for a single parent magma, such as that proposed by Bateman and Chappel (1979) for the Tuolumne Intrusive Suite, predicts singular, continuous, curvilinear trends for major- and trace-element relationships. Rare-earth-element patterns should typically be subparallel and systematically vary in absolute abundance with degree of differentiation in the magma body, depending on the fractionating phase(s).

The geochemical data support a petrogenetic model dominated by limited crystallization differentiation of a single parental magma for the Morena Reservoir pluton. However, crystallization differentiation alone cannot account for all the observed geochemical variations of the Morena Reservoir pluton, particularly for pluton-scale Na_2O and Zr variations, and the geochemical anomalies in the marginal samples. Additional petrogenetic mechanisms such as partial melting of a heterogeneous source material, magma/host equilibration, assimilation, volatile diffusion differentiation, magma recharge and mixing, or interstitial fluid migration, are needed in the model to explain the anomalous patterns in the data.

The composition, depth, and degree of partial melting of the Morena Reservoir pluton's source region can be crudely modeled by looking at the pluton's bulk rock composition and its REE patterns. The quartz diorite to tonalite compositions of the rocks, reasonable degrees of partial melting of somewhere between 10% to 60%, and constraining physical and phase equilibrium parameters, together limits source rock compositions to that of basalts. The gradual negatively-sloped, slightly concave-up rare-earth-element patterns, with 6X (heavy REE's) to 50X (light REE's) chondrite abundances and little to no negative Eu anomalies suggest a partial melting scheme very similar to that proposed for the western tonalites: melt derivation from the 20% to 40% partial melting of a plagioclase-bearing, gabbroic or amphibolitic source rock (Gromet and Silver, 1987). The transitional nature of these tonalites is partly defined by the very small to absent europium anomaly (plagioclase-poor? source region). The REE patterns modestly contrasts with the western tonalites which typically have larger europium anomalies (plagioclase-richer? source region). This places a maximum depth constraint on partial melting where plagioclase is still in equilibrium with the residuum, and this would roughly correspond to a maximum depth of 30 to 45 km.

Early-forming plagioclase, hornblende and accessory minerals sphene, apatite, opaque minerals, and zircon are noted in all samples. Sphene, apatite, opaque minerals, and zircon occur in trace quantities that are not precisely known; a large fraction of the whole rock's REE's reside in these accessory phases. The contribution by high-level differentiation processes to the REE patterns, such as early crystallization differentiation and removal of sphene, apatite, zircon, opaque minerals, hornblende, and plagioclase, would wholly or partially modify the source region signature. This would include the addition or exaggeration of 1) an enriched, steepened, fractionation pattern of the light REE's and depleted pattern in middle to heavy REE's by sphene; 2) a fractionated middle to heavy REE pattern by zircon; 3) a depleted concave-up pattern in the middle REE's by apatite; 4) a concave-up, depleted fractionation pattern of the middle and heavy REE's by hornblende; and 5) a negative Eu anomaly by plagioclase. These high-level, early-stage fractional crystallization effects should be reflected in the REE patterns in terms of the differentiation sequence of the Morena Reservoir rocks. Examination of the REE patterns in relation to inferred differentiation reveals no correlation; the most mafic-rich, silica-poor quartz diorites have patterns virtually identical to the most mafic-poor, silica-rich tonalites. Furthermore, the REE patterns do not appear to systematically change in absolute abundance in respect to increasing differentiation (decreasing FeO+MgO, increasing SiO_2). The gabbroic core rocks have identical middle to heavy REE patterns with the quartz diorites and tonalites.

The above observations indicate that the REE element patterns are most likely a direct image of the source rocks, as relayed by the rocks' primary melt, and that high level crystallization differentiation by the removal of early-forming plagioclase, hornblende, and accessory minerals from the parental melt was probably limited. If this is true and other processes such as assimilation and interstitial melt migration are negligible, then the Morena Reservoir rocks may closely represent emplacement-level melt compositions. However, with the ever-growing realization that nearly all plutons are multi-scale open-systems, the above inference is probably untrue. Open-system dynamics are further supported by anomalous geochemical compositions in the marginal samples. The disequilibrium microfabric features in virtually all samples, including strong oscillatory zoned plagioclase and resorbed hornblende indicate crystal-melt disequilibrium, features most likely caused by either, response to changing pressure-temperature-H_2O fugacity conditions, or melt hybridization by systematic mixing of evolved melt with fresh parental magma. Without stable isotopic data, however, the influences made by open-system exchange dynamics on the Morena Reservoir pluton's crystallizing parental melt cannot intelligently be made.

The zoned transitional nature of the Morena Reservoir, and Long Potrero tonalites, i.e, the intraplutonic zoning between tonalites having western-zone affinities and tonalites having more eastern-like affinities, is an enigmatic petrogenetic problem. The difficulty lies in the fact that much if not most of the information reflecting the key contrasting mediums, processes, and mechanisms that created the distinctive differences appears to be deleted by latter magmatic events that dominated the final rock fabrics and compositions. Perhaps the key information is cryptically encoded in the rock but has not been recognized petrographically or geochemically. Several possible petrogenetic models for the pluton's transitional nature are crudely sketched below.

The basic differences in the contrasting tonalite fabrics within the Morena Reservoir pluton, western-type versus eastern-type, could be controlled by differences in the pressure-temperature-rheology character of the different pulses in terms of the dynamically changing volumetric and compositional proportions of the cooling magma's crystal-melt system as it ascends. Two parental melts of similar composition, (but not of necessarily similar crystal-melt proportions and temperatures), leave similar source regions; however they ascend, cool, crystallize, and emplace by different styles and rates that are controlled in part by significant differences in initial magma crystal content and temperature, wall rock rheology and temperatures, lithostatic pressure, and regional tectonic forces.

Alternately, temporal contrasts in magma storage duration, either in or near the source region(s), or at the site of deep-level ponding (base? of the crust), could produce small but significant differences in the once identical primary magmas, such that older magmas had shorter residence time and hence less time to either segregate from their restitic residuum, differentiate, or interact/equilibrate with the country rocks. Whereas, the younger pulses came from primary melts that had longer deep-level incubation periods, and greater potential to interact and equilibrate with the containment rocks, or differentiate, prior to its major ascent.

The transitional contrasts could also be accounted for by differences in temperature and mechanical and chemical insulation between the ascending and emplacing older and younger pulses. Older pulses may have been more greatly subjected to mechanical and chemical interaction with the cooler wall rocks, including its own carapace, which may have included assimilation of, and contamination by, the wall rocks. The above interactions might have had significant cooling effects on the older pulses which would have initiated crystallization early in the pulses' ascent. The later pulses might have been better insulated from the above effects by their older, marginal pulses, and ascend at higher temperatures with considerably less crystallization during ascent.

The influence of a melt recharge mechanism, probably driven by density-controlled convective overturn in the magma chamber, would effectively mix cooler, more-dense differentiated high-level residual melts with hotter, less-dense, fresh parental melt from depth; this could partly explain the rather homogenous REE patterns, the limited pluton-scale lithologic and geochemical zoning, and the strong plagioclase oscillatory zoning.

Emplacement

The major mechanical parameters controlling the emplacement of the Morena Reservoir pluton would include the depth of emplacement, magma and country rock temperatures, rheology of the magma and country rocks, density contrast between the magma and the country rocks, and finally the local tectonic and structural setting. Several studies indicate that crystal-poor, intermediate-composition (quartz diorite/tonalite) magmas have the greatest density contrasts with crustal-composition country rock than do either the more silicic- or mafic-composition melts (references). Therefore, tonalite magmas would have the greatest buoyancy of the typical batholithic magma types, and hence would probably have the highest rates of ascent through the crust. These magmas would be the best candidates for "forceful" diapiric-style ascent and emplacement, with evidence of emplacement-related high strain and concordant/concentric flattening within and around these bodies. This would especially be true if the country rock were either partially molten or near its solidus temperature, and mass transfer would be accommodated under conditions of high ductility.

A general emplacement model for this pluton will be based on the above controls, field structural relations, emplacement-related features, and the regional tectonic setting. Interestingly, the Morena Reservoir pluton and the Long Potrero pluton, have all the above qualities in terms of composition and fabric, strain features, cooling history, and style of emplacement that classify them as rapidly ascending, upper mid-crustal, diapiric emplacement-style, tonalite plutons. The general premise for the emplacement of the Morena Reservoir pluton is as follows: the nearly completely molten Morena Reservoir pluton ascended rapidly from its source region, or zone of ponding without very much apparent assimilation or contamination with the wall rocks. The pluton crystallized under limited differentiation, disequilibrium conditions while it ascended through a dynamically changing temperature-pressure-rheology continuum. The pluton emplaced diapir-style in one or more pulses into a mid- to lower upper-crust that consisted mostly of older, near-solidus, plutonic rocks and genetically associated, synplutonic magmas by a combination of passive stoping and forceful dilation, under compressive conditions while in a variably submagmatic state. Potentially migrating interstitial fluids solidified while in disequilibrium with the adjoining crystals. Intensive uplift and rapid exhumation of the roof occurred immediately after the pluton solidified. And finally, the solidified pluton was tectonically deformed by sinistral oblique? faulting along the Hauser Canyon lineament between 101 Ma and 94 Ma; this deformational event is probably coeval with the pluton's rapid uplift, adjacent emplacement of the La Posta pluton, and temporally associated extensional faulting in the Cuyamaca-Laguna Mountain Shear Zone.

The above emplacement model for the Morena Reservoir pluton presumes a nearly completely molten state at time of ascent from its source; initial magma temperature would be approximately 1000 C°. This would provide maximum density contrast and hence most rapid ascent. The evidence for near molten, high temperature, rapid ascension conditions with little wall rock interaction for the incipient pluton is basically missing. However, the apparent absence of restite, lack of assimilation or contamination, presence of fabrics that indicate disequilibrium conditions, favor the idea of the rapid ascent of a relatively hot, highly buoyant magma. Crystallization during emplacement is also indicated by the significant disruption of the likely pressure-sensitive crystal-melt equilibrium conditions existing between hornblende and the melt. Strong oscillatory zoning in plagioclase may also indicate crystallization during ascent. Quantitative constraints on pressure and temperature for the Morena Reservoir pluton are lacking.

Stoping of the roof to facilitate emplacement space is evident in the pluton's core. Here a cone-shaped roof pendent of older gabbro has apparently sharp contacts that truncate its uniform northwest trending lithologies and structures, whereas the surrounding tonalite has concentric lithologies and structure around the gabbro. A similar contact is seen between the Morena Reservoir and gabbroic Los Pinos pluton.

Evidence for diapiric-style emplacement of one to several pulses under compressive, submagmatic conditions at the present erosional level is found in the pluton's concordant structural aureole, its concentric magmatic foliation trajectories, its flattened mafic enclave radial-form fabrics, and its significant submagmatic microfabric deformation. In the wall rocks is an apparently narrow (10's to 100 m), variably concordant macroscopic structural aureole around much of the pluton's margin. This aureole is particularly narrow and discordant in the gabbros, and highly concordant and somewhat broader in the granodiorites (even more so in the adjacent tonalites). This aureole would indicate relatively small amounts of near-field accommodation in the wall rocks, and minimal apparent mass transfer of the adjacent wall rocks. However, to make a proper evaluation of a ballooning mechanism for accommodating pluton displacement would require detailed structural mapping of the pluton's wall rocks. Also, a highly quantitative strain analysis of the flattened enclaves would be needed to better define the pluton's internal mass transfer space problem by means of lateral shortening.

John and Blundy (1993) have studied a smaller-sized, stronger-zoned pluton in the Italian Alps that has many lithologic and structural similarities with the Morena Reservoir pluton, particularly the emplacement-related, mafic enclave flattening and submagmatic deformation. Their results point towards a combination of emplacement mechanisms, that includes "passive" stoping, ballooning, and magma contraction. Their idea of magma contraction, as a consequence of solidification and interstitial melt migration, provides up to a 20% volume contribution towards the "space" problem. Average flattening of mafic enclaves in the Morena Reservoir pluton give a vague estimate of intra-pluton shortening of 30% to 50%. The combining of the above volume percentages would provide for roughly 50% to 70% of the necessary space for pluton emplacement. A volume of 30% to 50% of the pluton displacement would then be unaccounted for in the "space" problem, however a large part of this could be taken up in the structural aureole in the wall rocks.

Conclusion

The generation of the magma to form the Morena Reservoir pluton occurred close to the peak of the proposed mid-Cretaceous, mid-Pacific superplume (Larson, 1991). At this time, the Farralon and North American plates were part of a major plate reorganization, and their shift in combined velocities produced rapid dextral-oblique interplate motion in the subduction zone around 105 Ma. Walawender and others (1990) argued that the generation of transitional magmatism, such as the El Topo pluton, resulted from an abrupt increase in convergence rate around 105 Ma which shallowed the angle of subduction causing buckling and detachment in the down-going slab. They also consider the transitional magmatism to be derived from two source regions. Gromet and Silver (1987) consider the transition from western- to eastern-zone magmatism to be abrupt and concluded that the western-zone rocks had a relatively shallow, plagioclase-rich basalt/amphibolite source region, and the eastern-zone magmatism had a relatively deeper, plagioclase-poor, garnet-rich eclogite source region.

The zoned Morena Reservoir pluton has parts of its margin and core that are characterized by magnetite-rich, well foliated, western-type mineralogies and fabrics, and much of its intermediate regions that feature magnetite-free, poorly foliated, more eastern-type mineralogies and fabrics. Relic pyroxenes, magnetite, well-developed foliation, and lack of megacrystic quartz and sphene characterize the western-like rocks. Abundance of sharp euhedral hornblende, sphene, and biotite, lack of magnetite, megacrystic quartz, and little to no magmatic foliation define the more eastern-like tonalites. The western-type rocks are considered to be the older pulses, and the more eastern-like rocks represent the younger pulses. The zoning of the Morena Reservoir pluton is variably gradational, suggesting that the younger pulses intruded into the older pulses while the older pulses were still partly molten.

The Morena Reservoir tonalite has very similar mineralogies and textures to that of the Long Potrero pluton's biotite-hornblende and hornblende-biotite facies. However, the Morena Reservoir tonalite is generally less magnetic, foliated, and strained, as compared to the hornblende-bearing facies of the Long Potrero pluton. The Morena Reservoir tonalite also contains prismatic primary sphene which is absent in the Long Potrero pluton.

The geochemistry of the Morena Reservoir pluton is more similar to western-zone tonalites than eastern-zone tonalites, particularly the SiO_2 range, the paucity of K_2O, the local presence of pyroxene, and the REE patterns. These rocks have REE patterns with little or no negative Eu anomalies, and with significantly higher Sr content than most of the older western-zone tonalites. The Morena Reservoir pluton's REE patterns are similar to those of the Alpine and Long Potrero tonalite suites (Lombardi, 1990 and Rector, 1993, respectively).

Most of the Morena Reservoir pluton's major- and rare-earth-element geochemistry resembles that of the neighboring hornblende-bearing tonalites of the Long Potrero pluton. However, several of the major elements and many of the trace elements plot as diverging trends in respect to the plutons. This suggests a common source that generated independent magma diapirs that had similar yet isolated crystallization histories. In conclusion, the striking geochemical, mineralogical, structural, spatial, and temporal similarities and differences between the Morena Reservoir and Long Potrero plutons suggests an intimate, yet unique, petrogenetic and emplacement history for these plutons.

The geochemical data of the Morena Reservoir pluton, as well as that of the Long Potrero pluton, indicate that these intermediate composition magmas originated from a basaltic/amphibolitic source very similar to that of the older western tonalites. And yet only the mineralogy and textures of the more marginal rocks (older pulses?) are similar to the older western-type tonalites such as the Japtul- and Alpine-type tonalites of Todd (1978). Contrarily, much of the intermediate and interior parts of the pluton have more La Posta-like textures, i.e., Granite Mountain of Todd (1978). It appears that the Morena Reservoir pluton is zoned in crystallization- and emplacement-related tectonic conditions, and this zoning is interpreted as mainly being a consequence of time, with the growth of the pluton by two or more sequential pulses through a dynamically changing set of emplacement conditions over a period of perhaps 5 to 10 Ma. The question as to the roles of source region(s) versus ascent and emplacement conditions in respect to controls on the transitional nature of the pluton's rock texture, mineralogy, and geochemistry is difficult to decipher. How by blending the mediums and conditions that generated the western-zone Alpine tonalite with the ones that generated the eastern-zone La Pluton tonalite, might a Morena Reservoir or Long Potrero tonalite be produced?

To answer to this question would require the utilization and integration of the known fundamentals of tonalite-composition magma generation, ascension, and crystallization, together with the field, petrologic, and geochemical information provided from the rocks. The transitional differences are most likely a function of 1) source region depth and phase composition, 2) degree of partial melting and equilibration with source rock residuum, 3) temperature, density, water fugacity, and rheology of the incipient magma and containing country rocks, 4) degree of cooling, crystallization, and wall rock interaction of magma during ascension, and 5) final emplacement conditions, which would make the greatest reflection upon the solidus phase compositions and fabrics; these would include, depth of emplacement, water and oxygen fugacity, cooling history, wall rock/magma interaction, and tectonic (external), and magmatic (internal) processes that include emplacement-related deformational forces. The marginally-situated western-type rocks crystallized from the early-stage pulse(s) of the pluton. These pulses probably originated from a relatively shallower source region, ascended as crystalline-rich magmas through relatively cooler wall rocks, and emplaced in a submagmatic state under compressive conditions than the eastern-"like" rocks of the central part of the pluton . The eastern-"like" rocks probably crystallized from the later-stage pulses which originated from a slightly deeper source region, and ascended more rapidly as more crystalline-poor magmas through hotter wall rocks, i.e., through the older pulses, and emplaced under self-induced ballooning conditions mostly within the older pulses, and against the Long Potrero pluton. This caused significant radial flattening deformation in the older, marginal parts of the plutons and in the adjoining near- or above-solidus wall rocks.

The 101 Ma Morena Reservoir pluton emplaced into a much larger structurally concordant multi-pluton complex informally termed the Hauser complex. This western-zone complex contains small peripheral gabbro plutons, core-centered, zoned, transitional-type tonalite plutons, and complex-rimmed, massive granodiorite plutons. The silicic two-mica core of the 103 Ma Long Potrero tonalite pluton represents both the spatial and structural core of the Hauser complex, and may have a crystallization age very close to, or identical with, the Morena Reservoir rocks. The Morena Reservoir pluton is the youngest documented tonalite unit in the complex, but appears to be slightly? older than some of the adjacent granodiorites. The Long Potrero and La Posta plutons and the granodiorite plutons are structurally concordant with the Morena Reservoir pluton, whereas the gabbro bodies have generally discordant lithologic and structural trends in relation to that of the local marginal tonalites of the Morena Reservoir pluton.

The Morena Reservoir pluton intruded the biotite-hornblende facies of the Long Potrero pluton. The Ar-Ar cooling ages for these plutons indicate a similar cooling history, in respect to the western side of the Morena Reservoir pluton (LP-R18 versus MR-R591), but a significantly younger cooling history in respect to the Morena Reservoir pluton's eastern side (LP-R18 versus MR-R756). Yet, the zircon U-Pb ages of these plutons differ by only 2 Ma (LP-R18 versus

MR-R756), and the analytical error overlaps these ages. Thus, it would appear that the eastern side of the Morena Reservoir pluton was either, more deeply buried for a longer period of time than the Long Potrero pluton, or it was thermally overprinted by the emplacement of the La Posta pluton, an overprint not recorded by the sampled Long Potrero rocks. In either case, the Ar-Ar cooling histories of the Morena Reservoir pluton indicate rapid cooling of the pluton after solidification. This may indicate rapid uplift and exhumation of the pluton's roof during and/or immediately after emplacement and crystallization of the pluton.

Sinistral faulting along the Hauser Canyon lineament offsets the Long Potrero and Morena Reservoir plutons, but does not appear to fault or deform the 94 Ma La Posta pluton. However, the La Posta pluton has a 1.5 km left-lateral, curving jog in its undeformed western boundary corresponding to the extrapolated eastern extension of the Hauser Canyon lineament. Motion along this lineament may be genetically related to the emplacement of the La Posta pluton, and possibly with the closely coeval extensional shearing in the Cuyamaca-Laguna Mountain Shear Zone.

The megascopic magmatic foliation, flattened mafic enclaves, microscopic submagmatic deformation, and potential interstitial fluid migration, indicate considerable anisotropic compressional stresses in the pluton during emplacement, and possibly during the pluton's ascent to its level of emplacement. The significant disequilibrium textures noted in these rocks, such as hornblende resorption and mermikite textures, may be in part a symptom of the same processes that generated the deformation. Defining the roles between magmatic versus tectonic processes in the pluton's emplacement history and emplacement-related deformation is difficult. However, the pluton's core-centered, concentric foliation and radial flattening strains, with least deformation in the core and greatest in the margins, indicates emplacement-related stresses dominated by internal magmatic forces, such as forceful magma injection of younger pulses into the center of the body.

Several regional lineaments run through or along side the Morena Reservoir pluton and are locally concordant with the structural grain of the pluton, except for the Hauser Canyon lineament. These regional lineaments bracket or geometrically divide the large, multi-pluton, western zone complexes, including the Hauser complex, which the Morena Reservoir pluton is a part of; the La Posta-type plutons truncate the lineaments. Even though these lineaments appear to have great regional structural significance, their role in the development of the batholith is unclear, and evidence for movement or pluton accommodation along them is presently unknown. Therefore, their role in the emplacement of the Morena Reservoir pluton is uncertain.

REFERENCES

Ague, J. J. and Brimhall, G. H., 1988, Regional variation in bulk chemistry, mineralogy, and compositions of mafic and accessory minerals in the batholith of California: Geological Society of America Bulletin 100, p. 891-911.

Batchelor, M.D. and Bowden, P., 1985, Petrogenetic interpretation of granitoid rock series using multicationic parameters: Chemical Geology, v. 48, p. 43-55.

Bateman, P.C. and Chappell, B.W., 1979, Crystallization, fractionation, and solidification of the Toulumne Intrusive Series, Yosemite National Park, California: Geological Society of America Bulletin, v. 90, p. 465-482.

Bergreen, R.G. and Walawender, M.J, 1977, Petrology and metamorphism of the Morena Reservoir roof pendant, southern California: Division of Mines and Geology Special Report 129, p. 61-65.

Clark, D.B., 1992, Granitoid Rocks; Topics in the earth sciences, v. 7: London, Chapman and Hall, p. 283.

Clinkenbeard, J.P., 1987, The mineralogy, geochemistry, and geochronology of the La Posta pluton, San Diego and Imperial counties, California [M.S. thesis]: San Diego, California, San Diego State University, p. 215.

Corolla, G., 1993, Petrography and geochemistry of the western zone granodiorites, San Diego California: [M.S. thesis] San Diego, California, San Diego State University, 124 p.

De Paulo, D.J. 1981, A neodymium and strontium isotopicstudy of the Mesozoic calc-alkaline granitic batholiths of the Sierra Nevada and Peninsular Ranges, California. Journal of Geophysical Res. 86, p. 10470-10488.

Erskine, B.G. 1986, A paleomagnetic, rock magnetic, and magnetic mineralogic investigation of the northern Peninsular Ranges batholith, southern California [Masters thesis], Department of Geological Sciences, San Diego State University, 146 p.

Faure, G. 1977, Principles of isotope geology: New York, John Wiley and Sons, 589 p.

Gastil, R.G. 1975, Plutonic zones in the Peninsular Ranges of southern California and northern Baja California: Geology, v. 3, p. 361-363.

Gastil, R.G. and Miller, R., 1984, Prebatholithic paleogeography of peninsular California and adjacent Mexico: in Frizzell, V.A., Jr., ed., Geology of Baja California Peninsula: Pacific Section, Society of Economic Paleotologists and Mineralogists, v. 39, p. 9-16.

Gastil, R.G., 1990, Zoned plutons of the Peninsular Ranges in southern and Baja California: University Museum, University of Tokyo, Nature and Culture n. 2, p. 77-90.

Gastil, R.G., Diamond, J.,Knaack, D., Walawender, M. Marshall, M., Boyles, C., Chadwick, B., Erskine, b., Davis, T. and Smith, B., 1989, The problem of the magnetite-ilmenite boundary in southern and BajaCalifornia: in Anderson, J.L., ed., The Nature and Origin of Cordilleran Magmatism, Geological Society of America Memoir 174, p. 19-32.

Gill, J., 1981, Orogenic andesites and plate tectonics: New York, Springer-Verlag, 390 p.

Girty, G.H., Thomson, C., Miller, J., Carmicheal, D.L., Netto, S.P., 1994, Cretaceous extension, Scove Canyon segment, Cuyamaca-Laguna Mountains shear zone (CLMSZ), Peninsular Ranges, southern California (abs.): Geological Society of America Abstracts with Programs, v. 26, no. 2, p. 54.

Green, T.H., 1980, Island arc and continental building magmatism: a review of petrogenetic models based on experimental petrology and geochemistry. Tectonophysics v. 63, p. 367-385.

Gromet, L.P., and Silver, J.T., 1987, REE variations across the Peninsular Ranges batholith: Implications for batholithic petrogenesis and crustal growth in magmatic arcs: Journal of Petrology, v. 28, Part 1, p. 75-125.

Grove, M., 1994, Contrasting denudation histories within the east-central Peninsular Ranges batholith (33° N): Geological Society of America Field Trip Guidebook, Cordilleran Section, San Bernardino, California, p. 235-240.

Hearn, T.M., and Clayton, R.W. 1986, Lateral velocity variations in southern California, I. Results for the upper crust from Pg waves, II. Results for the lower crust from pn waves: Bulletin of the Seismological Society of America, v. 76, n. 2, p. 495-520.

Herzig, C.T., submitted, Petrology and eruptive setting of the Santiago Peak Volcanics: a primitive island arc association erupted over transitional oceanic-continental basement: American Journal of Science.

Hoppler, H., 1983, Petrology and emplacement of the Long Potrero pluton - a tail of one tonalite: [B.S. thesis], San Diego, Califor nia, San Diego State University, 54 p.

Hudson, F.S., 1922, Geology of the Cuyamaca region of California, with special reference to the origin nickeliferous pyrrhotite: University of California Department of Geological Sciences Bulletin, v. 13, p.175-252.

Hurlbut, C.S. Jr., 1935, Dark inclusions in a tonalite of southern California: American Mineralogist, v. 20, p. 609-630.

Jaffey, A.H., Flynn, K.F., Glenenine, L.F., Bentley, W.C., and Essling, A.M., 1971, Precision measurement of half- lives and specific activities of 235U and 238U: Physics Review, C4, p. 1889-1906.

John, B.E. and Bundy, J.D., 1993, Emplacement-related deformation of granitoid magmas, southern Adamello Massif, Italy: Geological Society of America Bulletin, v. 105, p. 1517-1541.

Kadik, A.A., and Lukanin, O.A. 1985, paths for mantle out-gassing during melting: changes in fluid composition and conditions in basaltic magmas during migration to the surface, Internat. Geological Review, v. 27, p. 573-586.

Kay, S.M., Kay, R.W., Citron, G.P. and Perfit, M.R., 1990, Calc-alkaline plutonism in the intra-oceanic Aleutian arc, Alaska, in Kay, S.M. and Rapela, C.W., eds. Plutonism from Antarctica to Alaska: Boulder, Colorado, Geological Society of America Special Paper 241, 263 p.

Kimbrough, D.L., in preparation, The Santiago Peak Volcanics: Early Cretaceous volcanic cover of the Peninsular Ranges batholith, southern and Baja California.

Kimsey, J.A., 1982, Petrography and geochemistry of the La Posta granodiorite: [Masters thesis], Department of Geological Sciences, San Diego State University, 150 p.

Krogh, T.E., 1973, A low-contamination method for hydro-thermal decomposition of zircon and extraction U and Pb or isotopic age determinations: Geochim. Cosmochim. Acta, v. 37, p. 485-494.

Krogh, T.E., 1978, Vapor transfer for the dissolution of zircons in a multi-sample capsule at high pressure,Short paper of the 4th international conference on geochronology, cosmochronology, and isotope geology: United States Geological Survey. Open File Report 78-101.

Krogh, T.E, 1982, Improved accuracy of U-Pb zircon ages by creation of more concordant systems using an air abrasion technique: Geochim. Cosmochim. Acta., v. 46, p. 637-649.

Krummenacher, D.K., Gastil, R.G., Bushee, J., and Doupont, J. 1975, K-Ar apparent ages, Peninsular Ranges batholith, southern California and Baja California: Geological Society of

America Bulletin, v. 86, p. 760-768.

Larsen, E.S. 1948, Batholith and associated rocks of Corona, Elsinore, and San Luis Rey quadrangles, southern California: Geological Society of America Memoir 29, 182 p.

Larson, R.L., 1991, Latest pulse of Earth: Evidence for a mid-Cretaceous superplume: Geology, v. 19, p. 547-550.

Larson, R.L., 1991, Geological consequences of superplumes: Geology, v. 19, p. 963-966.

Le Marchand, F., Villemant, B., and Calas, G., 1987, Trace element distribution coefficients in alkaline series: Geochemica Cosmochimica Acta. v. 51, p. 1071-1081.

Lombardi, M., 1991, The Alpine tonalite: image of a gabbroic source, San Diego County, California: [M.S. thesis], San Diego California, San Diego State University, 153 p.

Ludwig, K.R. 1989, A computer program for IBM PC-compatible computer for processing raw Pb-U-Th isotope data, version 1.05: United States Geological Survey Open File Report 88-542.

Maniar, P.D. and Piccoli, P.M., 1989, Tectonic discrimination of granitoids: Geological Society of America Bulletin, v. 101, p. 635-643.

Means, W.D. and Park, Y., 1994, New experimental approach to understanding igneous texture: Geology, v. 22, p. 323-326.

Miller, Joan, Girty, G.H., Girty, M.S., 1993, Harper Creek and Cuyamaca Reservoir Gneisses, CLMSZ: Late Jurassic plutons of the Peninsular Ranges batholith: Geological Society of America Abstracts with Programs, v. 25, n. 5, p. 122.

Oliver, H.W., 1982, ed., Interpretation of the gravity map of California and its continental margin: California Division of Mines Geological Bulletin, n. 205.

Parrish, R.R., 1987, Improvements in micro-capsules for zircon dissolution in U and Pb geochronology, in New developments and applications in isotope geosciences: Chemical Geology, (Isotope Geoscience Section), v. 66, p. 99-102.

Paterson, S.R., Vernon, R.H., and Tobisch, O.T., 1989, A review of criteria for the identification of magmatic and tectonic foliations in granitoids: Journal of

Structural Geology, v. 11, p. 349-363.

Peacock, M.A. 1931, Classification of igneous rock series, Journal of Petrology, v. 39, p. 54-67.

Pearce, J.A. 1982, Trace element characteristics of lavas from destructive plate boundaries, in Andesites: orogenic andesites and related rocks, R. S. Thorpe (ed) p. 525-548. Chichester: Wiley.

Pearce, J.A., Harris, N.B.W. and Tindle, A.G., 1984, Trace element discrimination diagrams for the tectonic interpretation of granitic rocks: Journal of Petrology, v. 25, p. 956-983.

Pitcher, W.S., 1978, The anatomy of a batholith: Journal of the Geological Society of London, v. 135, p. 157-182.

Pitcher, W.S., 1982, Granite type and tectonic environment, in Hsu, K.J., Mountain building processes: London, Academic Press, p. 19-40.

Pitcher, W.S., Atherton, M.P., Cobbing, E.J., and Beckinsale, R.D., 1985, Magmatism at a plate edge: New York, John Wiley and Sons, 328 p.

Pupin, J.P., 1980, Zircon and granite petrology: Contribu- tions to Mineralogy and Petrology, v. 73, p. 207-220.

Rector, R.R., 1989, The magnetic susceptibility of the Long Potrero pluton: San Diego County, California: [B.S. thesis], San Diego, California, San Diego State University, 54 p.

Rector, R.R., 1989, The magnetite distribution of the Long Potrero pluton, San Diego County (abs.): Geological Society of America Abstracts with Programs, v. 21, n. 6, p. 67.

Rector, R.R., 1993, The internal nature and dynamics of the Long Potrero pluton, San Diego County (abs.): Geological Society of America Abstracts with Programs, v. 25, n. 6, p. 303.

Sack, R.O., Carmichael, I.S.E., Rivers, M., and Ghiorso, M.S., 1980, Ferric-ferrous equilibria in natural silicate liquids at 1 bar, C.M.P. v. 75, p. 369-376.

Shimazu, Y., 1961, Physical theory of generation, upward transfer, differentiation, solidification and explosion of magmas: Journal of Earth Science, Nagoya University, v. 9, p 185-223.

Silver, L.T., Early, T.O. and Anderson, T.H., 1975, Petrological, geochemical and geochronological observations of the Peninsular Ranges batholith. Geological Society of America Abstracts with Programs, v. 7, p. 375-376.

Silver, L.T., Taylor, H.P. Jr, Chappell, B.W., 1979, Some petrological, geochemical and geochronological observations of the Peninsular Ranges batholith near the International border of U.S.A and Mexico, in Abbott, P.L., Todd, V.R., eds., Mesozoic crystalline rocks: Geological Society America Annual Meeting Guidebook, p.83-110.

Stacy J.S., and Kramer, J.D., 1975, Approximation of terrestrial Pb isotope evolution by a two stage model: Earth and Planetary Science Letters, v. 26, p. 207- 221.

Streckeisen, A.L., 1975, To each plutonic rock its proper name: Earth Science Reviews, v. 12, p. 1-33.

Taylor, H.P., Jr. and Silver, J.T., 1978, Oxygen isotope relationships in plutonic igneous rocks of the Penin- sular Ranges batholith, southern and Baja California, in Zartman, R.E., ed., Short papers of the Fourth International Conference on Geochronology, Cosmochemistry and Isotope Geology: United States Geological Survey Open File Report 78-701, p. 423-426.

Thompson, C., Girty, G.H., and Bracchi, K.A., 1994, Jurassic continental arc magmatism and the record of Early Cretaceous normal convergence between the North American and Farallon plates: Cuyamaca-Laguna Mountain shear zone (CLMSZ), Peninsular Ranges batholith (PRB), southern California (abs.): Geological Society of America Abstracts with Programs, v. 26, n. 2, p. 98.

Tobisch, O. T., Cruden, A.R., and O'Keefe, M.D., 1993, Constraints on the use of mafic enclaves as magmatic strain markers (abs.): Geological Society of America Abstracts with Programs, v. 25, n. 6, p. 305.

Todd, V.R., 1979, United States Geological Survey Open File 1:24,000 geologic mapping survey of the northern Peninsular Ranges batholith.

Todd, V.R. and Shaw, S.E., 1979, Structural, metamorphic and intrusive framework of the Peninsular Ranges batholith in southern San Diego County, California, in Abbott, P.L., and Todd, V.R., eds., Mesozoic crystal line rocks: San Diego, California, San Diego State University, Department of Geological Sciences, p. 177- 231.

Todd, V.R., 1985, S-type granitoids and an I-S line in the Peninsular Ranges batholith, southern California: Geology, v. 13, p. 231-233.

Todd, V.R., Erskine, B.G., and Morton, D.M., 1988, Metamorphic and tectonic evolution of the northern Peninsular Ranges batholith, southern California, in Ernst, W.G., ed., Metamorphic and crustal evolution of the western United States (Rubey Volume VII): Englewood Cliffs, New Jersey, Prentice-Hall, p. 894-937.

Todd, V.R., Kimbrough, D.L., and Herzig, C.T., 1994, The Peninsular Ranges batholith in San Diego County, California from volcanic arc to mid-crustal intrusive and metamorphic rocks: Geological Society of America Field Trip Guidebook, Cordilleran Section, San Bernardino, California.

Todd, V.R., in review, Geological map of the El Cajon 30'x 60' quadrangle, San Diego, California: US Geological Survey Open-File Report, scale 1:100,000.

Turekian, K.K. and Wedepohl, K.H. 1961, Distribution of the elements in some major units of the Earth's crust, Bulletin, Geological Society of America v. 72, p. 175-191.

Walawender, M.J., Gastil, R.G., Clinkenbeard, J.P., McCormick, W.V., Eastman, B.G., Wernicke, R.S., Wardlaw, M.S., Gunn, S.H., and Smith, B.M., 1990, Origin and evolution of the zoned La Posta-type plutons, eastern Peninsular Ranges batholith, southern and Baja California, in Anderson, J.L., ed., The nature and origin of Cordilleran magmatism: Boulder, Colorado, Geological Society of America Memoir 174, p. 1-18.

Walawender, M.J., 1976, Petrology and emplacement of the Los Pinos pluton, southern California: Canadian Journal of Science, v. 13, p. 1288-1300.

Wooden, J. L. and Stacy, J. S., 1987, Lead isotopic constraints on the origin of Cordilleren granitic magmatism in the western US (abs.): Geological Society of America Abstracts with Programs, v. 19, p. 465.

APPENDIX A

MODES FROM POINT COUNTING AND PETROGRAPHIC ANALYSES

Table 3.

Modal Analyses of the Morena Reservoir Suite

Sample #	Plag	Qrtz	Kfs	Hnbd	Biot	Op	Ap	Sp	Zr	Mz
MR-R581	56.9	22.6	3.0	7.1	10.4	Tr	NA	Tr	NA	NA
MR-R635	47.1	28.9	8.3	7.4	8.3	Tr	NA	Tr	NA	NA
MR-R650	54.1	22.3	1.0	13.3	8.3	Tr	Tr	Tr	Tr	NO
MR-R689	63.0	17.2	3.3	10.9	5.6	Tr	Tr	Tr	Tr	Tr
MR-R749	61.4	21.9	0.7	7.4	8.6	Tr	Tr	Tr	Tr	Tr
MR-R750	57.7	18.5	0.6	13.6	9.3	Tr	Tr	Tr	Tr	Tr
MR-R751	54.7	21.0	0.9	10.4	12.9	Tr	Tr	Tr	Tr	Tr
MR-R755	63.6	18.1	3.3	7.9	7.1	Tr	NA	Tr	NA	NA
MR-R756	50.3	21.9	2.4	8.4	17.0	Tr	NA	Tr	NA	NA
MR-R757	64.3	10.6	0.1	15.3	9.7	Tr	NA	NO	NA	NA
MR-R758	59.4	13.6	0.3	16.4	10.3	Tr	Tr	Tr	Tr	Tr
MR-R759	56.7	18.7	2.5	12.7	9.4	Tr	NA	NO	NA	NA
MR-R760	58.4	17.7	0.4	13.6	9.9	Tr	NA	NO	NA	NA
MR-R761	58.1	19.6	2.4	8.1	11.7	Tr	Tr	Tr	Tr	Tr
MR-R767	60.6	12.4	Tr	14.3	12.7	Tr	NA	Tr	NA	NA
MR-R772	55.2	18.9	1.7	10.7	13.6	Tr	NA	Tr	NA	NA
MR-R773	52.2	22.3	1.9	10.9	12.7	Tr	Tr	Tr	Tr	Tr

Symbol Key to Table 3. Plag = plagioclase; Qrtz = quartz; Kfs = potassium feldspar; Hnbd = hornblende; Biot = biotite; Op = opaques; Ap = apatite; Sp = sphene; Zr = zircon; Mz = monozite; Tr = trace; NO = not observed; NA = not available/no slide. Note: Major mineral modes from slab analyses; trace mineral identification from petrographic and field observations.

APPENDIX B

MAJOR-, TRACE-, AND RARE-EARTH-ELEMENT GEOCHEMICAL DATA

Geochemistry Sample Localities

The locations for the geochemistry samples found in Tables 4 and 5 below are shown in Plate I (in back pocket). Note however, that several of the samples for the Long Potrero pluton are located off the left edge of the map. The locations for these samples are alternately defined by longitude and latitude as follows:

LP-R18 = 116° 38' 2'' W longitude and 32° 38' 32'' N latitude; LP-R23 = 116° 37' 12'' W longitude and 32° 38' 17'' N latitude; LP-R26 and LP-R26I = 116° 36' 23'' W longitude and 32° 38' 21'' N latitude; LP-R27A = 116° 36' 12'' W longitude and 32° 38' 21'' N latitude; LP-R28 = 116° 38' 48'' W longitude and 32° 37' 48'' N latitude; LP-R56 = 116° 40' 3' W longitude and 32° 38' 13'' N latitude; LP-R57 = 116° 39' 57'' W longitude and 32° 38' 3'' N latitude; LP-R60B = 116° 39' 30'' W longitude and 32° 38' 16'' N latitude; PP-R16 = 116° 36' 21'' W longitude and 32° 36' 49'' N latitude; and PP-R754 = 116° 37' 48'' W longitude and 32° 37' 32'' N latitude.

Table 4. Geochemical Data on the Morena Reservoir Rocks

Unnormalized Results (Weight %)

Sample	MR-R652	MR-R689	MR-R749	MR-R750	MR-R751	MR-R758
SiO_2	61.99	63.76	63.19	59.07	63.07	56.81
Al_2O_3	16.98	16.52	17.02	18.00	16.84	17.83
TiO_2	0.81	0.76	0.84	0.96	0.82	1.05
FeO*	5.53	4.62	4.86	6.34	4.85	6.59
MgO	2.73	2.42	2.48	3.07	2.51	3.74
MnO	0.10	0.08	0.08	0.11	0.08	0.12
CaO	5.98	5.38	5.54	6.53	5.72	7.09
Na_2O	3.50	3.76	3.74	3.55	3.77	3.61
K_2O	1.78	1.98	1.96	1.63	1.76	1.36
P_2O_5	0.16	0.15	0.16	0.17	0.16	0.21
Total	99.56	99.42	99.83	99.43	99.59	98.42

Normalized Results (Weight %)

Sample	MR-R652	MR-R689	MR-R749	MR-R750	MR-R751	MR-R758
SiO_2	62.82	64.62	64.01	59.86	63.92	57.57
Al_2O_3	17.21	16.74	17.25	18.24	17.07	18.07
TiO_2	0.81	0.77	0.85	0.97	0.83	1.07
FeO*	5.60	4.68	4.93	6.43	4.92	6.68
MgO	2.77	2.45	2.51	3.11	2.54	3.79
MnO	0.10	0.08	0.08	0.11	0.08	0.12
CaO	6.06	5.45	5.61	6.62	5.80	7.19
Na_2O	3.55	3.81	3.79	3.60	3.82	3.66
K_2O	1.80	2.01	1.99	1.65	1.78	1.38
P_2O_5	0.16	0.15	0.16	0.17	0.16	0.21
Total	100.00	100.00	100.00	100.00	100.00	100.00

Trace and Rare Earth Elements (ppm)

Sample	MR-R652	MR-R689	MR-R749	MR-R750	MR-R751	MR-R758
Ni	6	10	7	5	8	6
Cr	20	21	22	24	23	41
Sc	17	11	11	11	10	18
V	106	99	111	129	98	136
Ba	458	629	480	540	531	372
Rb	69	82.12	96.91	57.49	80.92	41.14
Sr	380	384	421	400	406	447
Zr	142	127	130	142	150	10
Y	19.65	19.80	15.35	22.96	23.82	18.56
Nb	4.55	4.72	4.38	5.27	5.99	4.47
Ga	18	18	14	17	16	19
Cu	21	11	1	10	7	6
Zn	82	73	75	95	74	98
Pb	6.35	7.75	6.96	5.49	6.06	7.01
La	7.21	10.07	10.82	12.55	11.86	11.11
Ce	17.75	19.60	22.25	27.59	29.37	23.80
Pr	2.62	2.50	2.79	3.61	4.10	3.22
Nd	12.38	10.65	11.99	16.10	18.11	14.14
Sm	3.51	2.81	2.95	4.28	4.73	3.73
Eu	1.01	0.88	0.92	1.17	1.23	1.20
Gd	3.31	2.93	2.75	4.15	4.23	3.37
Tb	0.55	0.51	0.42	0.70	0.72	0.61
Dy	3.38	3.32	2.60	4.25	4.19	3.28
Ho	0.67	0.71	0.50	0.85	0.79	0.64
Er	1.84	2.05	1.39	2.29	2.24	1.75

Table 4. Trace and Rare Earth Elements (ppm)(continued)

Sample	MR-R652	MR-R689	MR-R749	MR-R750	MR-R751	MR-R758
Tm	0.26	0.30	0.19	0.31	0.32	0.25
Yb	1.59	1.91	1.20	1.94	1.90	1.48
Lu	0.24	0.31	0.19	0.29	0.29	0.23
Th	0.97	3.08	4.86	1.34	3.81	3.53
U	0.51	0.53	1.63	0.47	1.39	0.86
Hf	3.13	5.42	3.25	3.20	4.45	2.42
Ta	0.29	0.20	0.30	0.33	0.41	0.34
Cs	2.91	2.46	4.22	2.34	3.40	2.43
Sn	2.2	2.2	2.5	2.5	2.8	3.5

Unnormalized Results (Weight %)

Sample	MR-R761	MR-R768	MR-R773	MR-R579	MR-R790	MR-R794
SiO_2	67.40	54.41	61.68	65.55	45.33	49.37
Al_2O_3	15.54	17.57	17.36	16.41	22.08	19.36
TiO_2	0.55	1.29	0.73	0.64	1.73	1.54
FeO*	4.18	7.77	4.81	4.57	9.98	9.02
MgO	1.09	4.51	2.51	2.18	4.84	6.08
MnO	0.10	0.14	0.08	0.08	0.13	0.15
CaO	4.08	7.41	5.83	4.75	12.28	10.29
Na_2O	4.74	3.57	3.67	3.59	2.44	2.90
K_2O	1.26	0.87	1.93	2.49	0.12	0.28
P_2O_5	0.15	0.23	0.15	0.12	0.08	0.27
Total	99.10	97.77	98.74	100.38	99.02	99.27

Normalized Results (Weight %)

Sample	MR-R761	MR-R768	MR-R773	MR-R579	MR-R790	MR-R794
SiO_2	68.31	55.14	62.51	65.30	45.78	50.03
Al_2O_3	15.75	17.81	17.59	16.35	!22.30	19.62
TiO_2	0.56	1.31	0.74	0.64	1.75	1.56
FeO*	4.24	7.87	4.87	4.55	10.08	9.14
MgO	1.10	4.57	2.54	2.17	4.89	6.16
MnO	0.10	0.14	0.08	0.08	0.13	0.15
CaO	4.13	7.51	5.91	4.73	12.40	10.43
Na_2O	4.80	3.62	3.72	3.58	2.46	2.94
K_2O	1.28	0.88	1.96	2.46	0.12	0.28
P_2O_5	0.15	0.23	0.15	0.12	0.08	0.27
Total	100.00	100.00	100.00	100.00	100.00	100.00

Trace and Rare Earth Elements (ppm)

Sample	MR-R761	MR-R768	MR-R773	MR-R579	MR-R790	MRR794
Ni	7	8	7	4	0	9
Cr	4	37	24	18	24	92
Sc	19	28	16	20	27	26
V	49	180	101	93	362	236
Ba	459	438	451	562	59	100
Rb	30.55	21.74	76.73	92.20	2.19	4
Sr	268	493	427	330	522	545
Zr	195	163	119	106	64	81
Y	33.59	27.12	14.11	14.65	20.38	20
Nb	5.71	5.39	3.83	4.31	3.17	6.6
Ga	18	20	21	18	23	19

Table 4. (continued) Trace and Rare Earth Elements (ppm)

Sample	MR-R761	MR-R768	MR-R773	MR-R579	MR-R790	MRR794
Cu	3	18	3	7	29	14
Zn	71	!123	74	69	100	116
Pb	7.26	4.55	8.43	9.79	1.22	0
La	13.67	13.57	40.24	13.01	4.13	2
Ce	28.61	31.67	64.60	27.55	10.42	40
Pr	3.76	4.41	6.21	2.98	1.74	NA
Nd	17.24	20.12	20.84	11.91	9.36	NA
Sm	5.08	5.56	3.66	2.87	3.40	NA
Eu	1.38	1.29	0.96	0.79	1.42	NA
Gd	5.15	5.01	2.91	2.85	3.68	NA
Tb	0.96	0.84	0.44	0.44	0.69	NA
Dy	5.77	4.87	2.57	2.63	4.01	NA
Ho	1.18	0.96	0.48	0.51	0.77	NA
Er	3.37	2.62	1.30	1.49	2.06	NA
Tm	0.47	0.35	0.19	0.21	0.26	NA
Yb	2.94	2.19	1.17	1.29	1.54	NA
Lu	0.44	0.33	0.19	0.21	0.23	NA
Th	2.89	2.33	28.31	5.95	0.14	NA
U	0.78	0.55	3.58	1.20	0.04	NA
Hf	4.65	4.24	2.86	2.83	1.20	NA
Ta	0.34	0.34	0.24	0.59	0.18	NA
Cs	1.21	1.40	4.96	4.89	0.19	NA
Sn	3.0	2.8	2.5	NA	1.5	NA

Major elements are normalized on a volatile-free basis.
*Total Fe is expressed as FeO.
"!" denotes values >120% of lab's highest standard.
"NA" denotes no available data.\

Table 5. Geochemical Data on the Long Potrero Rocks

Unnormalized Results (Weight %)

Sample	LP-R18	LP-R23	LP-R26	LP-R27A	LP-R28	LP-R57
SiO_2	65.42	68.25	68.54	71.42	61.90	66.65
Al_2O_3	15.74	15.59	15.47	15.13	15.65	15.57
TiO_2	0.527	0.451	0.416	0.288	0.672	0.763
FeO*	4.19	3.47	3.40	2.31	5.57	4.89
MgO	2.11	1.53	1.44	0.79	3.12	1.71
MnO	0.082	0.074	0.067	0.058	0.110	0.079
CaO	5.36	4.07	3.97	3.49	6.15	4.53
Na_2O	3.66	3.87	4.13	4.30	3.15	3.95
K_2O	1.47	2.10	1.93	1.56	1.67	1.89
P_2O_5	0.116	0.101	0.104	0.084	0.095	0.199
Total	98.68	99.51	99.47	99.42	99.08	100.23

Normalized Results (Weight %)

Sample	LP-R18	LP-R23	LP-R26	LP-R27A	LP-R28	LP-R57
SiO_2	66.30	69.17	69.46	72.38	62.73	57.73
Al_2O_3	15.95	15.80	15.68	15.33	15.86	15.53
TiO_2	0.53	0.46	0.42	0.29	0.68	0.76
FeO*	4.25	3.52	3.45	2.34	5.64	4.88
MgO	2.14	1.55	1.46	0.80	3.16	1.71
MnO	0.08	0.07	0.07	0.06	0.11	0.08
CaO	5.43	4.12	4.02	3.54	6.23	4.52
Na_2O	3.71	3.92	4.19	4.36	3.19	3.94
K_2O	1.49	2.13	1.96	1.58	1.69	1.89
P_2O_5	0.12	0.10	0.11	0.09	0.10	0.20
Total	100.00	100.00	100.00	100.00	100.00	100.00

Trace and Rare Earth Elements (ppm)

Sample	LP-R18	LP-R23	LP-R26	LP-R27A	LP-R28	LP-R57
Ni	15	11	12	7	10	8
Cr	23	16	12	8	38	16
Sc	14	8	8	5	22	14
V	80	49	42	11	152	121
Ba	374	549	486	480	364	819
Rb	55.19	72.80	62.96	58.95	62.13	36.84
Sr	303	295	297	325	233	254
Zr	133	128	136	132	132	251
Y	20.61	14.38	13.58	6.07	24.87	27.10
Nb	3.37	3.34	3.24	3.28	3.36	3.78
Ga	15	14	16	15	13	17
Cu	11	4	4	6	7	10
Zn	61	58	57	49	66	67
Pb	7.66	9.14	8.83	8.75	7.24	8.27
La	19.99	5.01	16.42	4.60	15.32	17.37
Ce	38.58	9.96	33.67	7.15	29.33	37.27
Pr	4.30	1.43	3.94	0.88	3.46	4.71
Nd	16.80	7.07	15.31	3.73	14.19	20.80
Sm	3.81	2.08	3.11	0.84	3.70	5.56
Eu	0.97	0.72	0.77	0.77	0.96	1.39
Gd	3.58	2.48	2.63	1.34	3.55	6.06
Tb	0.62	0.40	0.40	0.14	0.67	1.02
Dy	3.67	2.39	2.34	0.90	4.13	6.21
Ho	0.76	0.49	0.48	0.19	0.90	1.28
Er	2.15	1.42	1.40	0.60	2.68	3.68

Table 5. (continued)

Trace and Rare Earth Elements (ppm)						
Sample	LP-R18	LP-R23	LP-R26	LP-R27A	LP-R28	LP-R57
Tm	0.31	0.20	0.20	0.10	0.38	0.51
Yb	2.02	1.35	1.38	0.75	2.45	3.21
Lu	0.33	0.23	0.22	0.14	0.40	0.50
Th	8.43	4.16	9.46	2.77	22.83	1.05
U	1.24	0.91	2.26	1.12	2.19	0.40
Hf	3.81	3.24	3.47	3.45	3.12	6.08
Ta	0.29	0.30	0.31	0.28	0.34	0.30
Cs	3.96	4.95	6.05	5.01	3.98	1.02
Sn	NA	NA	NA	NA	NA	2.0

Unnormalized Results (Weight %)						
Sample	LP-R60B	LP-R26I	LP-R199B	LP-R199BI	LP-R212	LP-R236
SiO_2	61.57	56.97	75.39	60.04	73.79	74.99
Al_2O_3	16.49	18.06	13.74	16.61	14.20	14.45
TiO_2	0.707	0.805	0.122	0.741	0.188	0.099
FeO*	6.02	6.43	1.33	7.26	1.88	1.14
MgO	2.84	3.59	0.17	3.29	0.32	0.13
MnO	0.113	0.182	0.044	0.454	0.079	0.050
CaO	6.14	6.49	2.50	4.56	2.19	2.42
Na_2O	3.59	4.67	4.64	2.25	4.55	5.02
K_2O	1.37	1.65	0.96	3.05	1.72	1.19
P_2O_5	0.125	0.158	0.075	0.367	0.067	0.081
Total	98.97	99.00	98.95	98.63	98.98	99.57

Normalized Results (Weight %)

Sample	LP-R60B	LP-R26I	LP-R199B	LP-R199BI	LP-R212	LP-R236
SiO_2	62.40	57.73	76.40	60.85	74.78	76.00
Al_2O_3	16.71	18.30	13.92	16.83	14.39	14.64
TiO_2	0.72	0.82	0..12	0.75	0.19	0.10
FeO*	6.10	6.52	1.35	7.36	1.91	1.16
MgO	2.88	3.64	0.17	3.33	0.32	0.13
MnO	0.11	0.18	0.04	!0.46	0.08	0.05
CaO	6.22	6.58	2.53	4.62	2.22	2.45
Na_2O	3.64	4.73	4.70	2.28	4.61	5.09
K_2O	1.39	1.67	0.97	3.09	1.72	1.19
P_2O_5	0.13	0.16	0.08	0.37	0.07	0.08
Total	100.00	100.00	100.00	100.00	100.00	100.00

Trace and Rare Earth Elements (ppm)

Sample	LP-R60B	LP-R26I	LP-R199B	LP-R199BI	LP-R212	LP-R236
Ni	9	13	10	5	10	9
Cr	29	28	5	32	2	0
Sc	22	19	0	11	1	0
V	123	140	0	71	0	0
Ba	436	505	441	268	709	510
Rb	45.90	78.63	15.61	76.03	29.00	15.30
Sr	278	351	326	324	294	310
Zr	139	97	134	102	137	104
Y	30.97	30.99	10.13	18.22	15.57	13.21
Nb	4.07	5.48	2.50	4.14	4.32	2.35
Ga	15	20	10	15	15	12
Cu	8	!459	7	86	6	7

Table 5. (continued)

Trace and Rare Earth Elements (ppm)

Sample	LP-R60B	LP-R26I	LP-R199B	LP-R199BI	LP-R212	LP-R236
Zn	75	107	38	!212	63	38
Pb	6.29	6.90	6.29	2.04	6.47	6.89
La	9.21	10.80	8.87	7.36	11.98	7.70
Ce	21.87	28.59	16.79	16.18	22.69	14.83
Pr	3.24	4.44	1.97	2.26	2.69	1.75
Nd	14.94	21.17	8.02	11.18	10.86	7.21
Sm	4.35	5.40	1.82	3.21	2.46	1.80
Eu	1.03	1.38	0.68	0.86	0.67	0.60
Gd	4.43	4.64	1.62	3.00	2.33	1.78
Tb	0.81	0.81	0.24	0.51	0.37	0.32
Dy	5.07	4.73	1.61	3.06	2.31	2.06
Ho	1.09	0.98	0.33	0.60	0.48	0.44
Er	3.17	2.89	0.94	1.61	1.37	1.23
Tm	0.45	0.41	0.14	0.22	0.20	0.18
Yb	2.90	2.86	0.93	1.37	1.42	1.18

Table 5. (continued)

Trace and Rare Earth Elements (ppm)

Sample	LP-R60B	LP-R26I	LP-R199B	LP-R199BI	LP-R212	LP-R236
Lu	0.46	0.49	0.16	0.21	0.23	0.19
Th	1.99	0.46	1.46	0.62	1.62	1.09
U	0.71	0.42	0.29	0.29	0.29	0.21
Hf	3.42	2.61	3.09	2.91	3.34	2.61
Ta	0.29	0.35	0.19	0.24	0.28	0.17
Cs	1.61	3.00	1.00	6.07	0.79	0.48
Sn	NA	NA	3.6	1.6	2.1	2.0

Unnormalized Results (Weight %)

Sample	LP-R338B	LP-R338C	LP-R346	LP-R402B	LP-R431B	LP-R554B
SiO_2	64.03	62.94	59.62	67.36	63.23	61.14
Al_2O_3	16.22	16.61	16.69	15.66	16.31	15.89
TiO_2	0.612	0.612	0.823	0.498	0.697	0.641
FeO*	4.95	5.28	6.78	3.74	5.40	5.03
MgO	2.41	2.53	3.94	1.67	2.86	2.63
MnO	0.092	0.100	0.126	0.068	0.098	0.087
CaO	5.50	6.29	7.19	4.32	6.14	5.89
Na_2O	3.67	3.66	3.23	3.80	3.23	3.32
K_2O	1.59	1.20	1.24	1.82	1.65	1.76
P_2O_5	0.125	0.117	0.126	0.104	0.112	0.109
Total	99.19	99.34	99.40	99.04	99.72	96.50

Table 5. (continued)

Normalized Results (Weight %)

Sample	LP-R338B	LP-R338C	LP-R346	LP-R402B	LP-R431B	LP-R554B
SiO_2	64.89	63.79	60.06	68.26	64.08	61.96
Al_2O_3	16.44	16.83	16.91	15.87	16.53	16.10
TiO_2	0.62	0.62	0.83	0.50	0.71	0.65
FeO*	5.02	5.35	6.87	3.79	5.47	5.10
MgO	2.44	2.56	3.99	1.69	2.90	2.67
MnO	0.09	0.10	0.13	0.07	0.10	0.09
CaO	5.57	6.37	7.29	4.38	6.22	5.97
Na_2O	3.72	3.71	3.27	3.85	3.27	3.36
K_2O	1.61	1.22	1.26	1.84	1.67	1.78
P_2O_5	0.13	0.12	0.13	0.11	0.11	0.11
Total	100.00	100.00	100.00	100.00	100.00	100.00

Trace and Rare Earth Elements (ppm)

Sample	LP-R338B	LP-R338C	LP-R346	LP-R402B	LP-R431B	LP-R554B
Ni	11	7	14	10	10	18
Cr	23	21	44	16	31	43
Sc	14	19	21	8	22	21
V	103	118	167	54	126	111
Ba	447	324	341	442	314	392
Rb	48.97	46.99	33.59	53.27	79.27	73.44
Sr	307	304	291	306	275	276
Zr	132	150	109	128	117	156
Y	21.84	29.70	25.06	13.88	24.02	29.33
Nb	3.46	3.26	3.26	2.74	3.57	3.88

Table 5. (continued)Trace and Rare Earth Elements (ppm)

Sample	LP-R338B	LP-R338C	LP-R346	LP-R402B	LP-R431B	LP-R554B
Ga	19	16	17	14	18	17
Cu	10	13	26	5	10	19
Zn	68	69	76	61	67	61
Pb	7.04	6.09	5.81	8.07	6.04	6.64
La	7.06	6.37	10.73	9.20	8.83	8.20
Ce	17.44	16.08	23.16	17.82	19.73	20.27
Pr	2.54	2.60	3.01	2.14	2.65	2.94
Nd	12.24	13.29	13.51	9.03	12.00	13.71
Sm	3.64	4.22	3.78	2.27	3.38	4.05
Eu	0.99	1.04	1.05	0.76	0.90	0.92
Gd	3.55	4.32	3.86	2.27	3.30	3.98
Tb	0.62	0.79	0.68	0.37	0.60	0.73
Dy	3.90	5.00	4.36	2.27	3.76	4.61
Ho	0.81	1.05	0.92	0.47	0.80	0.97
Er	2.28	2.97	2.69	1.33	2.27	2.74
Tm	0.33	0.42	0.37	0.20	0.32	0.38
Yb	2.09	2.71	2.45	1.29	2.09	2.50
Lu	0.33	0.42	0.38	0.21	0.33	0.38
Th	2.19	1.48	3.71	3.76	2.65	6.31
U	0.95	0.85	0.93	0.59	1.91	1.00
Hf	3.52	3.55	3.09	3.21	4.04	4.32
Ta	0.31	0.26	0.28	0.25	0.33	0.30
Cs	3.73	2.83	2.21	3.52	4.44	3.11
Sn	2.9	2.3	2.5	2.7	2.1	2.9

Table 5. (continued) Unnormalized Results (Weight %)

Sample	LP-691	PP-R16	PP-R754	LP-R56
SiO_2	61.84	51.42	49.76	50.86
Al_2O_3	16.46	19.01	20.43	20.18
TiO_2	0.671	1.102	1.018	0.886
FeO*	5.45	9.31	8.80	9.07
MgO	3.17	5.37	4.93	4.45
MnO	0.101	0.171	0.159	0.159
CaO	5.89	9.68	9.99	10.47
Na_2O	3.34	3.01	3.23	3.05
K_2O	1.57	0.30	0.09	0.23
P_2O_5	0.094	0.175	0.179	0.110
Total	99.24	97.55	98.59	99.45

Normalized Results (Weight %)

Sample	LP-R691	PP-R16	PP-R574	LP-R56
SiO_2	62.67	52.11	50.43	51.54
Al_2O_3	16.68	19.27	20.70	20.45
TiO_2	0.68	1.12	1.03	0.90
FeO*	5.52	9.44	8.92	9.19
MgO	3.21	5.44	5.00	4.51
MnO	0.10	0.17	0.16	0.16
CaO	6.06	9.81	10.12	10.61
Na_2O	3.38	3.05	3.27	3.09
K_2O	1.59	0.30	0.09	0.23
P_2O_5	0.10	0.18	0.18	0.11
Total	100.00	100.00	100.00	100.00

Table 5. (continued)
Trace and Rare Earth Elements (ppm)

Sample	LP-691	PP-R16	PP-R574	LP-R56
Ni	11	9	7	5
Cr	45	50	42	31
Sc	23	34	30	33
V	137	244	201	298
Ba	429	108	31	95
Rb	30.55	6.57	2.06	2.79
Sr	269	360	481	359
Zr	156	53	36	39
Y	19.76	16.04	9.35	10.96
Nb	2.59	2.45	1.38	0.66
Ga	15	15	22	19
Cu	16	16	1	13
Zn	61	104	100	90
Pb	5.98	2.58	0.94	1.23
La	10.07	6.42	4.00	3.97
Ce	19.60	12.77	8.13	8.10
Pr	2.50	1.71	1.18	1.15
Nd	10.65	8.02	5.71	5.48
Sm	2.81	2.47	1.70	1.87
Eu	0.88	1.14	0.93	1.07
Gd	2.93	2.18	2.62	1.75
Tb	0.51	0.49	0.31	0.45
Dy	3.32	2.99	1.81	2.62
Ho	0.71	0.63	0.36	0.56
Er	2.05	1.78	0.97	1.67

Table 5. (continued)

Trace and Rare Earth Elements (ppm) Sample	LP-R691	PP-R16	PP-R574	LP-R56
Tm	0.30	0.24	0.13	0.23
Yb	1.91	1.57	0.82	1.49
Lu	0.31	0.25	0.13	0.24
Th	3.08	0.74	0.17	0.35
U	0.53	0.21	0.05	0.08
Hf	5.42	0.84	0.30	0.60
Ta	0.20	0.18	0.09	0.07
Cs	2.46	0.35	0.08	0.25
Sn	2.2	2.2	2.0	4.5

Major elements are normalized on a volatile-free basis.
*Total Fe is expressed as FeO.
"!" denotes values >120% of lab's highest standard.
"NA" denotes no available data.

APPENDIX C

MAGNETIC SUSCEPTIBILITY INSTRUMENTATION, FIELD METHODS AND DATA

Data Collection Instrumentation

The susceptibility data was collected using a hand-held Geoinstruments JH-8 low-field intensity meter. The magnetic susceptibility meter works in the following way: two electro-magnetic coils are orthogonally positioned inside the contact-end of the meter, one transceiving, the other receiving; the transceiving coil is connected to a 9V battery, and generates a weak magnetic field; the applied, weak magnetic field permeates the sample to be measured, without saturating it, and induces magnetization of the sample; in turn, the temporarily induced magnetic field of the sample generates an induced voltage in the receiving coil of the meter, which is proportional to the magnetic susceptibility of the sample; this signal is amplified, rectified, and then used to drive a thermally compensated, directly calibrated, analog panel meter on the head-end of the magnetic susceptibility meter. The JH-8 meter is sensitive to 5×10^{-5} SI, and has a measuring range of 0 to 100000×10^{-5} SI. A four-setting knob separates the range into four ten-fold subranges.

Field Data Collection Procedures

The field measuring procedure is now described. Turn meter on, from the "0" setting to the "B" setting on the right hand side of meter. The battery is good if the needle is above the red mark on the scale. Now turn knob to the "1" setting. Zero the instrument by turning the left hand knob back and forth. Make sure to keep the meter at least 30 cm away from metal objects when zeroing or measuring. Find the flattest, least weathered outcrop surfaces for measuring. Press the sensitive end of the meter against the rock snugly. Turn the magnitude knob, if necessary, to the magnitude level which allows the meter needle to rest between 10 and 100. Record the reading. The most simple and efficient way to record the magnetic susceptibility value is to write down the value as two numbers separated by a hyphen. The first number represents the value read off the scale of the panel meter, and the second number represents the magnitude knob setting number. For example, a scale reading of 45 taken on a knob setting of "2" would be written as 45-2. This would translate to $45x10x10^{-5}$ SI, or $450x10^{-5}$ SI. This method is used to avoid making math conversions in the field.

Take as many measurements as needed, over the entire exposure, to get a characteristic range of values for the particular outcrop. A minimum of ten to forty readings, depending on observed variation, should be made for each outcrop. The characteristic value range can then be averaged for a susceptibility value that best represents that particular outcrop. However, the averaging process can be difficult for a small number of outcrops, e.g., in some instances, two values might be assigned to a single outcrop if two distinct ranges are noted after taking many measurements.

Magnetic susceptibility data that include ranges and averages for located outcrops in the Morena Reservoir pluton (see Plate I for locations) are found below in Table 3. Averaged magnetic susceptibility values for stationed outcrops were contoured over geology and topography (Plate IV)

Table 6.

Magnetic Susceptibility Data of the Morena Reservoir Pluton

Locality Code	M.S. (SI units) (values x10^{-5}) Range	Average	Locality Code	M.S. (SI units) (values x10^{-5}) Range	Average
MR-R321	35-40	38	MR-R569	15-25	20
MR-R322	38-50	45	MR-R570	40-90	65
MR-R323	600-780	650	MR-R571	80-240	160
MR-R324	820-2200	1800	MR-R572	80-250	180
MR-R325	200-200	200	MR-R573	300-2000	800
MR-R437	400-900	500	MR-R574	2000-7000	3500
MR-R535	200-780	530	MR-R575	80-400	210
MR-R535	800-2200	1800	MR-R576	40-2000	900
MR-R536	200-1600	1600	MR-R577	20-2000	800
MR-R537	35-75	60	MR-R578	20-1300	300
MR-R538	300-2000	1500	MR-R579A	400-1300	500
MR-R557	200-300	250	MR-R579B	N.R.	700
MR-R558	250-800	600	MR-R580	N.R.	700
MR-R559	30-40	35	MR-R581	40-120	75
MR-R560	25-33	29	MR-R582	160-480	300
MR-R561	28-35	32	MR-R583	180-380	280
MR-R562	45-150	75	MR-R584	200-300	250
MR-R563	80-400	250	MR-R585	40-180	110
MR-R566	20-40	30	MR-R586	65-180	130
MR-R567	28-32	30	MR-R587	80-210	170
MR-R568	40-90	60	MR-R588	35-120	80

Table 6. (continued)

Locality Code	M.S. (SI units) (values x10^{-5})		Locality Code	M.S. (SI units) (values x10^{-5})	
	Range	Average		Range	Average
MR-R589	65-240	150	MR-R617	80-200	150
MR-R590	80-220	160	MR-R619	300-1300	550
MR-R591	25-150	55	MR-R620	15-400	50
MR-R592	60-200	130	MR-R621	100-550	300
MR-R596	95-160	13	MR-R622	200-400	300
MR-R597	20-30	25	MR-R623	200-400	300
MR-R598	N.R.	N.R.	MR-R624	200-600	350
MR-R599	100-200	150	MR-R625	200-400	300
MR-R600	100-200	140	MR-R626	250-950	450
MR-R604	15-80	20	MR-R627	300-950	700
MR-R605	200-600	400	MR-R628	20-2500	500
MR-R606	200-2000	550	MR-R629	400-1000	600
MR-R607	200-900	550	MR-R630	300-1500	700
MR-R608	40-500	200	MR-R631	300-1500	70
MR-R609	10-450	165	MR-R632	250-1500	900
MR-R610	10-30	165	MR-R633	200-600	300
MR-R611	250-800	500	MR-R634	40-130	80
MR-R612	20-80	40	MR-R635	200-800	300
MR-R613	10-200	40	MR-R636	200-1000	600
MR-R614	20-200	60	MR-R637	600-1700	1200
MR-R615	100-500	300	MR-R638A	30-200	80
MR-R616	200-450	320	MR-R647	30-40	35

Table 6. (continued)

Locality Code	M.S. (SI units) (values x10^{-5})		Locality Code	M.S. (SI units) (values x10^{-5})	
	Range	Average		Range	Average
MR-R648	20-40	30	MR-R731	10-25	20
MR-R649	30-36	33	MR-R732	10-23	18
MR-R650	25-35	30	MR-R733	15-23	20
MR-R651	28-38	33	MR-R734	20-25	22
MR-R652	23-34	30	MR-R735	20-30	25
MR-R653	33-43	38	MR-R749	30-2200	1100
MR-R654	55-95	70	MR-R750	25-30	30
MR-R655	90-350	200	MR-R751A	60-250	100
MR-R656	60-80	70	MR-R755	600-2200	1400
MR-R657	45-150	75	MR-R756	20-27	25
MR-R658	25-95	70	MR-R757	60-260	130
MR-R659	25-250	200	MR-R758	60-200	130
MR-R660	50-150	75	MR-R759	All 20	20
MR-662	55-210	130	MR-R760	80-230	150
MR-R681	50-210	140	MR-R761	40-380	190
MR-R682	80-200	160	MR-R767	270-1300	670
MR-R687	22-35	30	MR-R768	2500-4500	3700
MR-R688	70-150	120	MR-R769	1600-2600	2200
MR-R689	60-130	90	MR-R770	380-1200	720
MR-R705	55-140	85	MR-R771	270-440	360
MR-R728	15-20	18	MR-R772	170-250	210
MR-R729	10-270	20	MR-R773	40-220	68

Table 6. (continued)

Locality Code	M.S. (SI units) (values x10^{-5})		Locality Code	M.S. (SI units) (values x10^{-5})	
	Range	Average		Range	Average
MR-R774	450-2100	1300	MR-R795	400-2000	800
MR-R775	750-2300	1400	MR-R796	500-1100	700
MR-R776	10-30	20	MR-R797	15-45	28
MR-R777	200-2000	800	MR-R798	20-40	30
MR-R778	1500-2500	2000	MR-R799	20-40	30
MR-R779	1000-2300	1600	MR-R800	35-55	45
MR-R780	600-1800	1100	MR-R801	35-55	45
MR-R781	150-8500	2800	MR-R802	20-30	25
MR-R782	600-1200	900	MR-R803	40-80	60
MR-R783A	600-4000	1800	MR-R804	80-240	160
MR-R783B	2000-4000	3000	MR-R805	280-810	560
MR-R784	600-4000	2800	MR-R806	350-1100	650
MR-R785	600-3000	2600	MR-R807	400-900	600
MR-R786	200-400	300	MR-R808	350-850	480
MR-R787	200-600	400	MR-R809	250-950	680
MR-R788	400-1000	700	MR-R810	250-900	700
MR-R789	200-2500	400	MR-R811	15-2200	850
MR-R790	3500-6500	4800	MR-R812	600-3000	1600
MR-R791	200-3000	2000	MR-R813	20-40	33
MR-R792	400-800	650	MR-R814	30-40	35
MR-R793	800-2200	1200	MR-R815	60-300	85
MR-R794	200-450	320	MR-R816	500-900	680

Table 6. (continued)

Locality Code	M.S. (SI units) (values x10^{-5})		Locality Code	M.S. (SI units) (values x10^{-5})	
	Range	Average		Range	Average
MR-R817	25-40	34			
MR-R818	200-800	320			
MR-R819	20-35	27			
MR-R820	130-240	180			
MR-R821	15-35	31			
MR-R822	250-400	320			
MR-R823	45-55	51			
MR-R824	15-25	20			
MR-R825	45-200	80			
MR-R826	35-45	39			
MR-R827	18-23	20			
MR-R828	15-22	18			
MR-R829	18-25	20			
MR-R830	45-600	230			
MR-R842	200-350	280			
MR-R843	100-200	150			
MR-R851	30-600	280			
MR-R852	100-420	280			
MR-R853	60-580	370			
MR-R854	200-600	430			

APPENDIX D

MAFIC ENCLAVE FIELD METHODS AND DATA

Measurement of Mafic Enclave Abundance

Abundance defined by the number of mafic enclaves in a given area is roughly proportional to the volume percent of mafic enclave material in the host-rock. Abundances were obtained by counting the number of mafic enclaves in several separate $9m^2$ areas within a single outcrop and taking the average count per $9m^2$. The abundance determined by the number of mafic enclaves in a $9m^2$ area can be less than or greater than the true abundance as defined by mafic enclave volume percent. This is due to the inhomogeneous variations in mafic enclave size within an outcrop and the limitations of the measuring system. The ranges in mafic enclave abundance for outcrops are typically +/- 15% of the recorded average. A small percentage of outcrops have abundance ranges that exceed +/- 100% of the recorded average. These outcrops typically have either very high abundances characterized by the presence of mafic enclave sheet-like clusters, or very low abundances.

The variation in abundance for $9m^2$ squares on gridded outcrops ranges from less than 1 in very low abundance outcrops to 200 in outcrops with sheet-like clusters of densely grouped mafic enclaves. Most outcrops have a relatively small size distribution for over 80% of their mafic enclaves. Averaged mafic enclave abundances for stationed outcrops in the Morena Reservoir pluton are found below in Table 4 (see Plate I for locations) . Mafic enclave abundances were contoured over topography and geology (Plate V).

Measurement of Mafic Enclave Morphology

Mafic enclaves were measured for size and shape in outcrop. Size ranges and averages were made for mafic enclave widths (short axis) and lengths (long axis), and where possible, heights (other long axis). The width-length-height nomenclature was used because virtually all of the mafic enclave flattened orientations measured in the pluton are subvertical. Length: width, and where possible, length:width:height are used to express the degree of flattening. Most outcrops lack well-exposed, vertical surfaces, and this greatly limits the observation of three dimensional variations in mafic enclave morphology. Thus, the analysis of enclave elongation is dominated by a sub-horizontal, two-dimensional data set represented by length and width. The variation in mafic enclave width:length in outcrops can be attributed to both real variation in short-axis:long-axis of the mafic enclaves, and in the apparent variation in cross-section shape of mafic enclaves (cross-sections that do not correspond to short-axis:long-axis plane of the mafic enclave. However, in a high percentage of the outcrops that afforded good three dimensional observation, it is noted that the height component of mafic enclaves typically matches the length component and that the averaged width:length of all mafic enclave cross-sections in outcrop corresponds to the average short-axis:long axis measured in three dimensions.

Mafic enclave width to length ratios for stationed outcrops in the Morena Reservoir pluton, both ranges and averages, are found below in Table 7 (see Plate I for locations). Mafic enclave width to length ratios were contoured over topography and geology (Plate VI).

Table 7.

Mafic Enclave Data of the Morena Reservoir Pluton

Locality	Abundance (count/9m^2)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range		Avg	Avg	Max
MR-R321	50	1:2 - 1:4	1:3	7	40	
MR-R322	30	1:2 - 1:3	1:2½	6	25	
MR-R323	60	1:1 - 1:4	1:2½	4	40	
MR-R324	40	1:1 - 1:2	1:1½	3	10	
MR-R325	100+	1:4 - 1:10	1:7	6	50	
MR-R534	50	N.R.	1:2½	20	N.R.	
MR-R535	70	N.R.	1:2	16	N.R.	
MR-R536	60	1:1 - 1:2	1:1½	N.R.	N.R.	
MR-R537	200+	N.R.	N.R.	N.R.	N.R.	
MR-R538	250	N.R.	N.R.	N.R.	N.R.	
MR-R558	N.O.	---	---	---	---	
MR-R559	30	1:2 - 1:3	1:2	4	10	
MR-R560	40	N.R.	N.R.	11	23	
MR-R561	45	1:1 - 1:4	1:3	N.R.	N.R.	
MR-R562	45	1:3 - 1:5	1:4	5	35	
MR-R563	20	N.R.	N.R.	N.R.	N.R.	
MR-R566	40	N.R.	N.R.	N.R.	N.R.	
MR-R567	40	N.R.	N.R.	N.R.	N.R.	
MR-R568	25	1:1 - 1:2	1:1½	N.R.	N.R.	
MR-R569	30	1:2	1:2	N.R.	N.R.	
MR-R570	40	1:3	1:3	N.R.	N.R.	

Table 7. (continued)

Locality	Abundance (count/9m^2)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R571	25	1:1 - 1:2	1:1½	N.R.	N.R.	
MR-R572	25	1:1 - 1:2	1:1½	N.R.	N.R.	
MR-R577	20	ALL 1:2	1:2	5	20	
MR-R578	20	N.R.	N.R.	N.R.	200	
MR-R581	120	1:2 - 1:2	1:4	4	100	
MR-R582	90	N.R.	1:4	4	25	
MR-R583	65	1:2 - 1:4	1:3	4	20	
MR-R584	35	1:1 - 112	1:1½	4	20	
MR-R585	45	1:1 - 1:3	1:2	4	20	
MR-R587	30	1:2 - 1:5	1:3	12	N.R.	
MR-R588	30	N.R.	N.R.	N.R.	N.R.	
MR-R591	20	1:2 - 1:5	1:3	N.R.	N.R.	
MR-R592	15	N.R.	1:3	15	N.R.	
MR-R597	30	N.R.	N.R.	N.R.	N.R.	
MR-R598	50	N.R.	N.R.	N.R.	N.R.	
MR-R599	8	N.R.	N.R.	N.R.	N.R.	
MR-R600	20	N.R.	N.R.	N.R.	N.R.	
MR-R606	2	N.R.	N.R.	N.R.	N.R.	
MR-R607	2	N.R.	N.R.	N.R.	N.R.	
MR-R608	2	N.R.	N.R.	N.R.	N.R.	
MR-R609	N.O.	---	---	---	---	
MR-R611	60	1:1 - 1:3	1:2	6	N.R.	

Table 7. (continued)

Locality	Abundance (count/9m²)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R612	40	N.R.	N.R.	6	N.R.	
MR-R613	20	N.R.	N.R.	N.R.	N.R.	
MR-R615	10	N.R.	N.R.	N.R.	N.R.	
MR-R616	10	N.R.	1:2	12	N.R.	
MR-R619	100	N.R.	1:3	12	N.R.	
MR-R620	30	N.R.	1:3	11	N.R.	
MR-R621	15	N.R.	N.R.	N.R.	N.R.	
MR-R624	15	N.R.	1:3	12	200	
MR-R625	<1	N.R.	N.R.	N.R.	N.R.	
MR-R626	20	N.R.	1:2½	20	N.R.	
MR-R627	150	N.R.	1:3	30	N.R.	
MR-R628	30	N.R.	1:4	16	N.R.	
MR-R629	30	N.R.	N.R.	N.R.	N.R.	
MR-R630	55	N.R.	1:4	16	N.R.	
MR-R631	30	N.R.	N.R.	N.R.	N.R.	
MR-R632	40	N.R.	1:3	55	1000	
MR-R633	80	N.R.	1:4	N.R.	16	
MR-R634	5	N.R.	N.R.	N.R.	N.R.	
MR-R635	10	1:1 - 1:3	1:2	N.R.	N.R.	
MR-R636	60	N.R.	N.R.	N.R.	N.R.	
MR-R637	40	N.R.	1:4	N.R.	16	
MR-R638A	15	N.R.	1:4	50	100	

Table 7. (continued)

Locality	Abundance (count/9m^2)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R647	60	N.R.	1:2	20	N.R.	
MR-R648	35	N.R.	N.R.	N.R.	N.R.	
MR-R649	40	N.R.	1:2	10	N.R.	
MR-R650	50	1:3 - 1:8	1:5	15	N.R.	
MR-R651	50	1:1 - 1:3	1:2	10	N.R.	
MR-R652	35	1:1 - 1:4	1:2½	12	N.R.	
MR-R654	60	N.R.	1:3	14	N.R.	
MR-R655	50	1:3 - 1:6	1:4	16	N.R.	
MR-R656	N.R.	N.R.	1:3	14	N.R.	
MR-R657	8	N.R.	N.R.	N.R.	N.R.	
MR-R660	30	N.R.	1:2	15	N.R.	
MR-R662	25	1:1 - 1:2	1:1½	10	N.R.	
MR-R666	30	N.R.	1:4	40	N.R.	
MR-R681	8	N.R.	1:1½	12	N.R.	
MR-R682	15	N.R.	N.R.	N.R.	N.R.	
MR-R687	25	1:2 - 1:5	1:3	12	N.R.	
MR-R688	10	N.R.	N.R.	N.R.	N.R.	
MR-R689	150	N.R.	N.R.	N.R.	1000	
MR-R705	8	1:1½ - 1:3	1:2	10	N.R.	
MR-R728	9	1:3 - 1:4	1:3½	N.R.	N.R.	
MR-R729	30	1:3 - 1:1	1:6	N.R.	N.R.	
MR-R730	12	1:2 - 1:4	1:3	N.R.	N.R.	

Table 7. (continued)

Locality	Abundance (count/9m^2)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R731	N.R.	1:3 - 1:6	1:4	N.R.	N.R.	
MR-R732	35	1:2 - 1:8	1:3½	N.R.	N.R.	
MR-R733	20	1:2 - 1:8	1:3½	N.R.	N.R.	
MR-R734	10	N.R.	1:2	N.R.	N.R.	
MR-R735	4	1:1 - 1:2	1:1½	15	N.R.	
MR-R749	100	1:3 - 1:10	1:6	12	N.R.	
MR-R750	50	1:1 - 1:6	1:2	N.R.	N.R.	
MR-R751A	15	1:1 - 1:3	1:2	12	N.R.	
MR-R751B	15	1:1 - 1:3	1:2	12	N.R.	
MR-R755	65	1:2 - 1:6	1:3	9	40	
MR-R756	45	1:1 - 1:3	1:2	12	200	
MR-R757	20	1:2 - 1:3	1:2½	12	N.R.	
MR-R758	50	1:1½ - 1:6	1:3	N.R.	N.R.	
MR-R759	40	1:2 - 1:2	1:8	15	50	
MR-R760	60	1:3 - 1:15	1:6	N.R.	N.R.	
MR-R761	70	1:3 - 1:3	1:9	20	200	
MR-R767	120	1:2 - 1:12	1:3	16	100	
MR-R768	50	1:2 - 1:4	1:3	15	80	
MR-R769	40	1:2 - 1:15	1:4	16	N.R.	
MR-R770	40	1:3 - 1:20	1:6	N.R.	N.R.	
MR-R771	20	1:2 - 1:8	1:4	N.R.	N.R.	
MR-R772	65	1:2 - 1:15	1:4	N.R.	N.R.	

Table 7. (continued)

Locality	Abundance (count/9m²)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R773	80	1:2 - 1:8	1:3	N.R.	100	
MR-R774	110	1:1 - 1:4	1:2	5	25	
MR-R775	150	1:1 - 1:5	1:3	5	N.R.	
MR-R776	230	1:4 - 1:13	1:7	N.R.	100	
MR-R777	200	1:8 - 1:15	1:10	200	N.R.	
MR-R778	110	1:3 - 1:8	1:5	11	N.R.	
MR-R779	70	1:2 - 1:8	1:4	50	N.R.	
MR-R780	90	1:1 - 1:5	1:3	N.R.	150	
MR-R781	130	1:2 - 1:12	1:4	6	30	
MR-R782	120	1:2 - 1:20	1:6	N.R.	100	
MR-R783A	60	1:4 - 1:30	1:10	N.R.	N.R.	
MR-R783B	350	N.R.	N.R.	N.R.	N.R.	
MR-R784	200	N.R.	1:20	N.R.	N.R.	
MR-R785	3	1:1 - 1:2	1:1½	N.R.	N.R.	
MR-R786	3	1:1 - 1:2	1:1½	N.R.	N.R.	
MR-R792	40	1:2 - 1:3	1:2½	N.R.	N.R.	
MR-R795	100	1:2 - 1:5	1:3	13	100	
MR-R796	80	1:2 - 1:5	1:3	13	100	
MR-R797	65	1:2 - 1:6	1:3	16	80	
MR-R798	45	1:1 - 1:3	1:2	9	30	
MR-R799	35	1:2 - 1:4	1:3	N.R.	N.R.	
MR-R800	N.R.	1:2 - 1:8	1:3	7	35	

Table 7. (continued)

Locality	Abundance (count/9m²)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R801	30	1:1 - 1:4	1:2	N.R.	N.R.	
MR-R802	35	1:2 - 1:30	1:5	17	1000	
MR-R803	23	1:2 - 1:4	1:2	11	25	
MR-R804	23	1:2 - 1:4	1:2	11	25	
MR-R805	50	1:1 - 1:5	1:3	10	40	
MR-R809	60	1:1 - 1:2	1:1½	10	30	
MR-R810	80	N.R.	1:4	N.R.	200	
MR-R806	55	N.R.	N.R.	N.R.	N.R.	
MR-R807	30	1:1 -1:2	1:1½	10	30	
MR-R808	50	N.R.	N.R.	N.R.	N.R.	
MR-R812	5	1:1 - 1:2	1:1½	17	N.R.	
MR-R813	80	1:2 - 1:4	1:3	12	150	
MR-R814	32	1:1 - 1:3	1:2	6	25	
MR-R815	25	N.R.	1:2	7	30	
MR-R816	40	1:1 - 1:2	1:2	15	150	
MR-R817	28	1:2 - 1:5	1:2½	8	25	
MR-R818	20	1:1 - 1:3	1:2	10	60	
MR-R819	17	N.R.	1:2	8	25	
MR-R820	20	1:1 - 1:2½	1:2	7	35	
MR-R821	16	1:2 - 1:5	1:2½	8	30	
MR-R822	26	1:2 - 1:4	1:3	10	40	
MR-R823	62	1:2 - 1:8	1:8	10	40	

Table 7. (continued)

Locality	Abundance (count/9m^2)	Shape (width:length)		Size (length(cm))		Code
	Avg	Range	Avg	Avg	Max	
MR-R824	25	1:3 - 1:8	1:4	N.R.	N.R.	
MR-R825	120	1:2 - 1:10	1:4	15	200	
MR-R826	33	1:2 - 1:7	1:4	7	100	
MR-R828	45	1:3 - 1:10	1:4	N.R.	N.R.	
MR-R829	28	1:2 - 1:5	1:3	10	100	
MR-R830	20	1:2 - 1:4	1:3	7	N.R.	
MR-R842	15	1:2 - 1:6	1:4	12	N.R.	
MR-R843	36	1:2 - 1:4	1:3	20	200	
MR-R851	25	1:3 - 1:6	1:4	15	700	
MR-R852	25	1:2 - 1:6	1:3	12	100	
MR-R853	30	1:3 - 1:20	1:10	N.R.	N.R.	
MR-R854	13	1:3 - 1:6	1:4	14	30	
MR-R855	32	1:2 - 1:5	1:3	12	40	

Abundance = an averaged count from three or more 9 m^2 grids of horizontal to subhorizontal outcrop surface; Shape = a dimensionless width (short axis) to length (long axis) ratio; Avg = Average, Max = maximum.

APPENDIX E

FOLIATION FIELD DATA

Table 8.

Foliation Data of the Morena Reservoir Pluton

Locality	Strike/Dip	Int.	Locality	Strike/Dip	Int.
MR-R321	N26W 75NE	2	MR-R570	N25W 75NE	4
MR-R322	N18W 79NE	3	MR-R571	N25E 70NE	2
MR-R323	N25W 87NE	4	MR-R572	N30W 70NE	2
MR-R324	N5E ---	2	MR-R573	N60W ---	1
MR-R325	N30E 70SE	4	MR-R574	N70W 70NE	4
MR-R437	N70E 70SE	C5	MR-R575	N35W 75NE	C3
MR-R534	N15E 80SE	C5	MR-R576	N30W 90	C6
MR-R536	N15E 80SE	C5	MR-R577	N25W 77NE	6
MR-R536	N.R.	4	MR-R578	N35W 80NE	5
MR-R537	N40E ---	5	MR-R579A	N.R.	C5
MR-R538	N80E 82NW	C6	MR-R579B	N.R.	C5
MR-R539	N.R.	6	MR-R580	N.R.	C6
MR-R558	N.R.	1	MR-R581	N10W 70NE	4
MR-R559	N12W 70NE	3	MR-R582	N8W 77NE	C4
MR-R560	N20W 68NE	3	MR-R583	N33W 72NE	C4
MR-R561	N17W 67NE	4	MR-R589	N48W 56NE	3
MR-R569	N50W 65NE	4	MR-R591	N6OW 65NE	4

Table 8. (continued)

Locality	Strike/Dip	Int.	Locality	Strike/Dip	Int.
MR-R592	N45W 75NE	4	MR-R622	N.R.	4
MR-R596	N.R.	4	MR-R623	N.R.	4
MR-R597	N8E 78SE	4	MR-R624	N66E 80SE	4
MR-R598	N40W 66NE	5	MR-R625	N70E 72SE	3
MR-R599	N80E 70SE	4	MR-R626	N5W 80NE	3
MR-R600	N50W 78NE	3	MR-R627	N15W 75NE	C6
MR-R603	N60E 84SE	C4	MR-R628	N20W 80NE	C4
MR-R604	N20E 80NW	4	MR-R629	N.R.	4
MR-R605	N.R.	1	MR-R630	N5W 70NE	4
MR-R606	N10E 76SE	4	MR-R631	N10W 70NE	3
MR-R607	N.R.	1	MR-R633	N20E 82NW	5
MR-R608	N.R.	3	MR-R634	N.R.	2
MR-R609	N.R.	1	MR-R635	N60W 70NE	3
MR-R610	N40W 80NE	C5	MR-R636	N45W ---	4
MR-R611	N48W 83NE	4	MR-R637	N.R.	5
MR-R612	N30W 72NE	C5	MR-R638A	N-S 80E	4
MR-R613	N.R.	C5	MR-R647	N15W 80NE	3
MR-R614	N.R.	4	MR-R648	N20W 75NE	4
MR-R615	N40W 67NE	3	MR-R649	N16W 68NE	4
MR-R616	N50W 80NE	3	MR-R650	N23W 70NE	5
MR-R617	N.R.	5	MR-R651	N24W 83NE	3
MR-R619	N40W 65NE	C5	MR-R652	N26W 69NE	4
MR-R620	N30W 75NE	C4	MR-R653	N15W 78NE	4
MR-R621	N50W 75NE	C4	MR-R654	N36W 72NE	4

Table 8. (continued)

Locality	Strike/Dip	Int.	Locality	Strike/Dip	Int.
MR-R655	N26W 78NE	4	MR-R751B	N15W 75NE	5
MR-R657	N30W 80NE	5	MR-R755	N25E 69NW	C5
MR-R658	N16W ---	4	MR-R756	N15E 70NW	4
MR-R659	N15W 80NE	4	MR-R757	N56W 81SW	3
MR-R660	N20W 78NE	4	MR-R758	N80W 78SW	4
MR-R662	N32W 78NE	3	MR-R759	N60E 83NW	3
MR-R666	N32 80SW	C6	MR-R760	N70W 75NE	C4
MR-R681	N29W 86NE	4	MR-R761	N50W 66NE	4
MR-R682	N55W 90	4	MR-R767	N27W 73NE	C4
MR-R687	N25W 70NE	4	MR-R768	N20E 80NW	C6
MR-R688	N30W 82NE	4	MR-R769	N26E 85NW	C5
MR-R689	N40W 88NE	4	MR-R770	N30E 78SE	5
MR-R705	N46W 90	3	MR-R771	N18E 73SE	5
MR-R728	N34W 79NE	4	MR-R772	N29E 71SE	4
MR-R729	N24W 86NE	4	MR-R773	N17E 73SE	4
MR-R730	N9W 86NE	4	MR-R774	N10E 82NW	C5
MR-R731	N-S 87E	C4	MR-R775	N-S 83NW	C5
MR-R732	N4W 72NE	C4	MR-R776	N23E 90	6
MR-R733	N10W 74NE	C4	MR-R777	N12E 86NW	C5
MR-R734	N15W 78NE	3	MR-R778	N20W 82SW	C4
MR-R735	N22W 90	2	MR-R779	N15W 80SW	C5
MR-R749	N40E 80NW	C4	MR-R780	N7W 79SW	C5
MR-R750	N10E 70SE	3	MR-R781	N15W 82SW	C4
MR-R751A	N15W 75NE	5	MR-R782	N4E 86NW	C5

Table 8. (continued)

Locality	Strike/Dip	Int.	Locality	Strike/Dip	Int.
MR-R783A	N-S 90	C5	MR-R815	N39W 78NE	3
MR-R783B	N10W 88SW	5	MR-R816	N63W 60SW	2
MR-R784	N10W ---	C6	MR-R817	N5E 58SE	3
MR-R794	N.R.	1	MR-R818	N39W 72NE	3
MR-R795	N30E 80NW	C4	MR-R819	N17E 75SE	3
MR-R796	N23E 78NW	C4	MR-R820	N14E 79SE	2
MR-R797	N22W 79SW	C4	MR-R821	N2W 81NE	2
MR-R798	N13W 81SW	C3	MR-R823	N71E 82SE	4
MR-R799	N15W 85SW	C5	MR-R824	N32E 80SE	4
MR-R800	N5W 80SW	4	MR-R825	N56E 79NW	4
MR-R801	N2W 85NE	3	MR-R826	N44E 82SE	3
MR-R802	N2W 77SW	C4	MR-R827	N38E 77NW	4
MR-R803	N4W 84SW	C4	MR-R828	N40E 90	4
MR-R804	N4W 84SW	C4	MR-R829	N32E 83SE	3
MR-R805	N12W 84NE	C4	MR-R830	N33E 815E	C5
MR-R806	N8W 79SW	C5	MR-R842	N66W 79NE	5
MR-R807	N15E 86NW	3	MR-R843	N52W 72NE	C6
MR-R808	N10E 80SE	C2	MR-R851	N17E 79SE	5
MR-R809	N5E 90	C3	MR-R852	N9E 72SE	4
MR-R810	N7W 80SW	C5	MR-R853	N13W 77NE	4
MR-R812	E-W 80S	2	MR-R854	N22W 75NE	4
MR-R813	N32E 73SE	4	MR-R855	N16W 74NE	4
MR-R814	N5E 82SE	3			

Foliation Table Key. Int. = foliation intensity; intensity number codes: 1 = weak; 2 = weakly moderate; 3 = moderate; 4 = moderately strong; 5 = strong; and 6 very strong. C = crenulated foliation (tightly folded foliation); N.R. and "---" = not recorded; and N.O. = no foliation observed.

ABSTRACT

The 101 Ma, 65 sq km Morena Reservoir pluton is a northwest-trending, vertically-tapered?, "Y"-shaped body. The pluton consists of a foliated and strained, medium-grained, compositionally zoned (color index range of 16 to 27), sphene-bearing biotite-hornblende tonalite and quartz diorite. The Morena Reservoir pluton represents PRB transitional-type tonalite magmatism that emplaced into the middle to upper crust under compressional conditions. Its complex crystallization history suggests long-term crystal-melt disequilibrium conditions. The Morena Reservoir pluton is part of a much larger, structurally concordant/concentric multi-plutonic complex informally named the Hauser complex that includes small peripheral gabbro plutons (oldest?); larger, concentrically zoned, complex-cored, tonalite plutons; and extensive, massive, complex-rimmed, granodiorite and leucomonzogranite plutons (youngest?). The gradual-sloped, slightly concave upwards REE patterns show 6X (heavy REE's) to 50X (light REE's) chondrite abundances and little to no negative Eu anomalies. The REE profiles suggest melt derivation from the partial melting of a plagioclase-bearing gabbroic or amphibolitic source rock. Limited fractional crystallization appears to be the dominant single control on compositional variation, as evidenced by the numerous tight, smooth, curvilinear Harker diagram trends. Sampled rock from the pluton's margin have anomalous geochemistry and small zircon inheritance?, and most likely signify modest assimilation or contamination from the wall rocks. The pluton's mafic enclaves are most likely either, entrained and disseminated fragments of the conduit walls and carapace that represent older cognate pulses, or

disseminated, synplutonic dikes. The concentric patterns of contoured field data suggest diapiric emplacement of one to several pulses via a central conduit, rather than emplacement by multiple injections of sheet-style dikes. Both the microfabric submagmatic deformation that includes fractured plagioclase and hornblende and strained quartz and biotite, and the likely synchronous migration of interstitial fluids are considered to be emplacement-related. The long axes of mafic enclaves are parallel to the magmatic foliation, and the degree of flattening is proportional to the foliation intensity. Mafic enclave flattening and foliation are believed to have resulted from the same emplacement-related compressional mechanism(s) such as ballooning and/or regional transpressional tectonics. Internal contraction estimated from mafic enclave flattening (20% to 40% volume reduction) plus potential magma solidification shrinkage (up to 20%) could account for roughly 30% to 60% of the space requirements for pluton emplacement. The general lack of a wide, strongly-deformed, structural aureole in the wall rocks suggests a low percentage of space accommodation by local wall rock contraction (up to 20%?). Roof stoping and far-field extensional tectonics are the most likely complimentary mass transfer mechanisms for the remaining 20% to 40% of the pluton's displacement needs. Sinistral oblique offsets of plutonic boundaries and contoured field data, and evidence of both left-lateral and dip-slip displacements in a west-striking/north-dipping shear zone, occur along the Hauser Canyon lineament. Field relations and isotopic dating constrain the timing of displacement between 101 and 94 Ma. Displacement along the Hauser Canyon lineament is closely synchronous with La Posta pluton emplacement and active extension in the Cuyamaca-Laguna Mountain Shear Zone.