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Geochemistry and origin of mafic schists from the Pelona, Orocopia, and Rand Schists; structure and metamorphism of the Orocopia Schist, southern California

Malcom Robert Dawson II

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Geochemistry and origin of mafic schists from the Pelona, Orocopia, and Rand Schists; structure and metamorphism of the Orocopia Schist, southern California

Dawson, Malcom Robert, II, Ph.D.
Iowa State University, 1987
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Geochemistry and origin of mafic schists from the Pelona, Orocopia, and Rand Schists; structure and metamorphism of the Orocopia Schist, southern California

by

Malcom Robert Dawson II

A Dissertation Submitted to the Graduate Faculty in Partial Fulfillment of the Requirements for the Degree of DOCTOR OF PHILOSOPHY

Department: Earth Sciences
Major: Geology

Approved:

Signature was redacted for privacy.

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Iowa State University
Ames, Iowa

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GENERAL INTRODUCTION

The Orocopia, Pelona, and Rand Schists of southern California are thought to have formed in an accretionary prism of Late Cretaceous age. These schists are oceanic in nature and consist of interlayered graphitic mica/quartz/feldspar schists, metabasites, metacherts, and marbles. Tectonically emplaced over these rocks are continental Precambrian, Paleozoic, and Mesozoic igneous and metamorphic rocks. Determination of the polarity of this subduction zone is important for reconstruction of the Mesozoic paleogeography of the southwestern United States.

This dissertation is divided into three "self-contained" chapters which describe three separate approaches to this problem. The first chapter describes a geochemical study (major, minor, and trace elements, including rare earth elements) of the metabasalts. The purpose of this project was to characterize the types of rocks which formed their protolith and to determine the similarities and differences between the Orocopia, Pelona, and Rand mafic schists.

The second chapter is a description and interpretation of the structures and microstructures of the Orocopia Schist of the Orocopia Mountains (OMS). Macroscopic structures of the OMS were studied to compare movement directions in the Orocopia and Chocolate Mountains. Microstructures within albite porphyroblasts were examined and interpreted to determine earlier stages of the deformational history and to constrain pressure-temperature histories of deformation and metamorphism. Folds and microstructures were studied to look for
evidence which might indicate how these folds formed, which could aid in the interpretation of stretching lineations, fold orientations, folding styles, and tectonic movement directions.

The third chapter describes the petrology and metamorphism of the Orocopia Schist of the Orocopia Mountains. It is important to understand the relationship of deformation to metamorphism because kinematic indicators which formed during uplift may have a shear sense opposite to those formed during subduction, and because late kinematic indicators have been used to interpret tectonic movement directions. The purpose of this study was two-fold: (1) to characterize the metamorphism which produced the matrix minerals of the OMS for comparison to correlative localities; and (2) to look for evidence of either prograde or retrograde P or T so tectonic models might be constrained. This was done through petrographic study and determination of mineral compositions by electron microprobe.

Each chapter contains a summary or conclusions for each project. To avoid repetition, figures, tables, and plates are consecutively numbered. Samples which are labeled with an "OR", "PR", "RA", "SG", or "SP" prefix were collected by Carl Jacobson. Samples with no prefix were collected by the author.
CHAPTER I.

GEOCHEMISTRY AND ORIGIN OF MAFIC ROCKS FROM THE PELOMA, OROCOPIA, AND RAND SCHISTS, SOUTHERN CALIFORNIA
ABSTRACT

Mafic schists from the Pelona-Orocopia-Rand (POR) subduction complex, southern California, have been analyzed for major, minor, and trace element concentrations to investigate the nature, origin, and tectonic implications of these rock bodies. Element abundances indicate these rocks are tholeiitic and alkalic basalts generated from several sources. Most samples (group 1a) resemble normal to enriched mid-ocean ridge basalts (N-MORB to E-MORB). Variation of MORB-like samples can be accounted for by low-pressure crystal fractionation. A second group of samples (group 2) resembles non-depleted basalts (P-type). Trace element and REE abundances indicate these basalts formed as an ocean island or seamount. A third group (group 1b, a subset of samples from the Rand Mountains) has elemental abundances similar to E-MORB, however, Y, Yb, Lu, and P are more depleted. Several of these samples (group 1b*) have high MgO, Ni, and Cr, characteristics which are commonly associated with crystal accumulation. The group 1b samples may be within-plate tholeiites.

The interlayering of these rocks with metagraywackes is possibly due to tectonic mixing within an accretionary prism. The data do not support tectonic models that call for protolith formation in either a rifted back-arc basin or an island arc. The data are consistent with subduction models in which oceanic crust originated in an open ocean or in a marginal basin formed either by entrapment of older oceanic crust or transform-style tectonics.
INTRODUCTION

The Pelona, Orocopia, and Rand (POR) Schists are thought to have formed in an accretionary prism of Late Cretaceous age now exposed in southern California and southwestern Arizona (Figure 1). The POR schists consist of approximately 90% quartzofeldspathic and 10% mafic schists which underlie Precambrian, Paleozoic, and Mesozoic continental igneous and metamorphic rocks along the Vincent-Chocolate Mountains thrust. Unlike the Franciscan Complex of northern California, the POR schists are located within the continent, and not at the continental margin. Models to explain the intra-continental location of the POR accretionary complex postulate that it either is a correlative of the Franciscan Complex emplaced during a period of low-angle easterly dipping subduction (Yeats, 1968; Crowell, 1968, 1981; Burchfiel and Davis, 1981; Dickinson, 1981) or that it formed in a westerly dipping zone associated with the closing of a marginal basin. This marginal basin could have formed by rifting and extension perpendicular to the continental margin (Japanese type: Haxel and Dillon, 1978; Haxel et al., 1985), closing of a larger ocean by migration of an exotic terrane toward North America (Ehlig, 1981; Vedder et al., 1983; Harding and Coney, 1985) or oblique extension (Gulf of California type: Tosdal et al., 1984).

The mafic rocks within the POR schists are generally thought to be oceanic in nature; however, conflicting opinions also exist regarding the specific magmatic origin of these rocks and the mode of emplacement which juxtaposed the protolith of the mafic and quartzofeldspathic
Figure 1. Map of southern California showing sample localities of the Pelona-Orocopia, and Rand Schist (after Jacobson et al., in press). BR, Blue Ridge area of San Gabriel Mountains; EF, East Fork area of San Gabriel Mountains; OM, Orocopia Mountains; PR, Portal and Ritter Ridges; SP, Sierra Pelona; RM, Rand Mountains. Mesozoic plutonic rocks are indicated by unpatterned outlined areas.
schists. Because of concordant contacts and a lack of observable feeder dikes, Ehlig (1958, 1968) interpreted the mafic schist to have been a tuff. Dillon (1976), Haxel (1977), and Haxel and Tosdal (1986) believed the mafic and felsic protoliths were not tectonically juxtaposed, but were deposited together and represent submarine eruptions onto a turbidite fan complex. Sharry (1981) believed the protolith was basaltic, but could not rule out calcareous sediments. He interpreted the mafic schist to represent tectonic blocks within a melange.

The Pelona, Orocopia, and Rand Schists are commonly correlated with each other on the basis of lithologic and metamorphic similarities and because all lie tectonically beneath rocks of continental origin. Unlike the Pelona and Orocopia Schists, which were adjacent to each other before 300 km of Late Cenozoic right-lateral displacement occurred along the San Andreas fault system, the Rand Schist lay distant from the Late Cretaceous Pelona and Orocopia locations, being 200 km to the northwest. Additionally, a 79 m.y. post-metamorphic intrusion of granodiorite into the Rand Schist (Silver and Nourse, 1986) indicates an apparent age difference between the Pelona (52-59 m.y.; Haxel and Dillon, 1978) and Rand metamorphism.

The purpose of this paper is to present a geochemical study (major, minor, and trace elements) of POR mafic schists to characterize the types of rocks which formed their protolith. If the mafic schists were derived from magmas, their compositions may constrain present tectonic models and help determine if the Rand mafic schist and the lithologically similar Pelona and Orocopia mafic schists are consanguineous.
DESCRIPTION OF THE MAFIC SCHIST

The Pelona and Orocopia Schists occur along and near the San Andreas fault in southern California and southwestern Arizona. The Rand Schist is exposed south of the Garlock fault in the Rand Mountains. The most abundant rock type (90%) in all areas is quartzofeldspathic schist, interpreted to be metagraywacke (Ehlig, 1958; Haxel et al., 1986). Mafic schists occur within the metagraywacke and comprise approximately 10% of the total formation. The mafic layers range in thickness from millimeters to several hundred meters, and can commonly be traced for several kilometers along strike. Ferromanganiferous quartzites, presumably derived from chert, are volumetrically minor but areally widespread. They are generally associated with the mafic schists. Marble and isolated pods of serpentine, actinolite plus talc, and pods of manganese oxides also are present.

The mafic schist is generally comprised of fine-grained amphibole and chlorite plus albite and epidote of variable grain size. The variations in mineral abundance and grain size define a small-scale compositional layering. With the exception of a few isolated blueschist localities (RA38, RA69A this study), the rocks are now predominantly greenschists to oligoclase amphibolites. At most locations, the schists are thoroughly recrystallized and foliated. Isoclinally refolded folds and as many as two additional sets of more open folds are present, as well as brittle deformation structures (Jacobson et al., in press). Few primary structures were seen in the mafic schist, although relict pillow
structures are preserved at one location in the Sierra Pelona (Ehlig, 1981; SP25 this study).
SAMPLING AND ANALYTICAL METHODS

Twenty-eight mafic samples were analyzed for major, minor, and trace elements by X-ray fluorescence (XRF). Of these samples, 19 were analyzed for trace elements by instrumental neutron activation analysis (INAA). The samples are Pelona Schist from the San Gabriel Mountains, Portal Ridge and the Sierra Pelona; Orocopia Schist from the Orocopia Mountains; and Rand Schist from the Rand Mountains (Figure 1). The analyzed samples were selected from a much larger sample suite collected from the five mountain ranges. Samples for analyses were chosen that represent: 1) varying distance to the thrust contact with the overlying continental rocks, which correlates to the range of metamorphic facies exhibited by the schist and 2) the range in bulk composition inferred from probe data and modal mineralogy.

Samples showing late-stage veining and alteration were generally avoided. It should be noted that, in some samples, minor effects of retrograde metamorphism can be seen in thin section. These features include alteration of garnet and biotite to chlorite.

Approximately 150 grams of each sample was ground in a corundum shatter mill to produce a single aliquot of powder. All analyses were performed on splits from these powders. Whole-rock concentrations of major and minor elements were determined by X-ray fluorescence (XRF) using a Siemens XRF spectrometer at Iowa State University. Analyses for Si, Al, Fe, Ca, Mg, K, and Ti were performed on glass discs prepared by fusing the sample with a lanthanum-bearing lithium borate mixture (Norrish and Hutton, 1969). Samples for fusion were first heated to
1000°C for four hours to drive off all volatiles and convert all iron to Fe₂O₃. The elements Na, Mn, and P were determined from pressed powder pellets. Accuracy is believed to be 3% relative for major elements and 5% relative for minor elements, based on standards run as unknown samples.

Some trace elements (Ni, Rb, Sr, Y, Zr, Nb) were determined from pressed powder pellets on a Siemens XRF at the Institute of Mining Research, University of Zimbabwe (UZ) and on a Kevex energy dispersive XRF at Iowa State University (ISU), using loose powders. Agreement of trace element values from the two laboratories was very good with the exception of Rb. Results from UZ are reported for Ni, Sr, Y, Zr, and Nb. Results from ISU are reported for Rb based on agreement of analyzed values with published values for standard BHVO when it was run as an unknown. The rare earth elements (La, Ce, Sm, Eu, Yb, Lu) and Cr, Sc, Hf, Ta, and Th were analyzed by instrumental neutron activation analysis (INAA) at the University of Missouri Research Reactor, Columbia, Missouri. INAA analytical techniques, sample preparation and accuracy are similar to that of Jacobs et al. (1977). Samples from this INAA run were also analyzed at Washington University, St. Louis, with very good agreement between the facilities.
RESULTS

Major Elements

To simplify the discussion of data, samples have been grouped based on similarity of elemental abundance and relative elemental proportions. These groups are referred to as "group 1a, group 1b, group 1b* and group 2". Group 1b* is a subgroup of group 1b.

Whole-rock compositions and normative analyses for all 28 samples are shown in Table 1. The major elements indicate the rocks are basaltic, with SiO$_2$ between 45-52 wt%. An AFM diagram (Figure 2; fields of Irvine and Baragar, 1971) indicates that most samples (group 1a and group 1b) are tholeiitic with an iron enrichment trend and minor alkali enrichment. One sample from the Orocopia Mountains (280B, group 2) lies away from the tholeiitic trend. Three samples (group 1b*) are enriched in MgO (RA66, RA73 and RA74; Table 1).

A tholeiitic nature for group 1a samples is also indicated by the relationship of alkalies (Na + K) to SiO$_2$ (not shown). Normative compositions confirm the tholeiitic character of group 1a samples (Table 2) and suggest that most rocks were olivine tholeiites. A few samples are quartz (Q) or nepheline (NE) normative. Of the samples that are nepheline normative, most are only slightly undersaturated.

The Pelona, Orocopia, and Rand (FOR) Schists have a wide range of Mg# (Mg-number) values, where Mg# = 100 Mg/(Mg+Fe$^{2+}$) in atomic proportions, and Fe$^{3+}$ is obtained by assuming Fe$^{3+}$/Fe$^{2+}$ = 0.15. Most of the group 1a basalts have Mg# < 63 which implies that the basalts have been modified by fractionation (cf. Green, 1971). Relatively
**TABLE 1. Major-element (weight percent) and trace-element (ppm) abundances of Pelona, Orocopia and Rand mafic schists**

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<th>Sample:</th>
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<th>OR180</th>
<th>OR200</th>
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Notes: XRF analyses on fusion discs and pressed powder pellets using Siemens XRF spectrometer at Iowa State University and University of Zimbabwe; REE,Hf,Ta,Th,Sc by INAA analysis at Columbia Research Reactor. Dashes indicate below limits of detection. ND = not determined. Loss on ignition (LOI) at 1000°C. Mg-number = 100 X Mg/Mg+Fe<sup>2+</sup>, FeO<sub tot</sub>/FeO = 0.15. <sup>a</sup> Group 2, alkalic within-plate basalts.
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Rb 20 34 18 30 19 0 ND 31
Sr 112 156 40 220 100 35 148 75
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| Ce | 3.75| 2.91| 3.38| 5.87| 2.03| 3.55| ND| 7.58|
| Sm | 1.44| 1.06| 1.30| 1.97| 0.83| 1.29| ND| 2.23|
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| Yb | 0.58| 0.21| 0.19| 0.32| 0.40| 0.24| ND| 1.10|
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| Th | 0.25| 0.64| 0.27| 0.84| - | 0.24| ND| 0.29|

b Group 1b, samples which may be within-plate tholeiites.
Compositions believed influenced by crystal accumulation.
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# TABLE 2. CIPW norms of Pelona, Orocopia and Rand mafic schists

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Notes: Norms calculated on Fe₂O₃/FeO = 0.15, deletion of LOI and normalizing.

Group 2, alkalic within plate basalts.
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*Group lb, samples which may be within-plate tholeiites.

*Group 1b compositions believed influenced by crystal accumulation.
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Figure 2. AFM (A=Na$_2$O+K$_2$O, F=FeO+Fe$_2$O$_3$, M=MgO) diagram with POR mafic schist data. Tholeiitic field is above the solid line (Irvine and Baragar, 1971). See Table 1 for sample grouping and text for discussion.
unfractionated samples are present, but uncommon (PR20A Mg#=68).

Variation of elements relatively immobile during metamorphism and seafloor alteration, such as iron and titanium, also indicates a tholeiitic character for the POR mafic schist. Samples of group 1a are nearly identical to average MORB in FeO and TiO₂ content (Figure 3, field from data of Nelson et al., 1976). Because Ti is generally an incompatible element, an increase of TiO₂ in the melt is expected as fractionation proceeds, while Mg/(Mg+Fe²⁺) decreases. Enrichment of TiO₂ correlating with a decrease of Mg-number is seen in Figure 3. The POR schists plot along the trend of open system fractionation of MORB (Shervais and Kimbrough, 1985). The group 1b* samples plot close to the main trend; however, they have relatively high Mg-numbers (74, 71, and 69) and high MgO values.

Within the Ca-Mg-Al-Si (CMAS) system, a modified basalt tetrahedron can be defined. The ol-cpx-plag-qtz subsystem contains the idealized non-oxide minerals present in oceanic basalts that crystallize at low pressure and can be used to evaluate the conditions of origin of oceanic magmas (Presnall et al., 1979; Walker et al., 1979; Elthon and Scarfe, 1984; and others). The POR schists are plotted on a CMAS pseudoternary diagram (Figure 4). This diagram is a projection from plagioclase onto the clinopyroxene-olivine-quartz plane (Elthon, 1983). The cotectic is defined by an experimental liquid-line-of-descent at 1 bar and confirmed by nearly 2000 MORB samples (Walker et al., 1979). POR basalts plot in the same region as mid-ocean ridge basalts in that olivine normative basalts plot in a trend to and approximately parallel
Figure 3. Plots of less mobile FeO, TiO₂, Al₂O₃, and more mobile Na₂O, K₂O, and CaO versus Mg# for POR mafic schist. Stippled area represents field for oceanic glasses (data from Nelson et al., 1976). For K₂O, the outlined area represents 90 percent of all points.
Figure 4. Data for POR metabasites plotted in Ol-Di-SiO$_2$ plane projected from plagioclase, a modified basalt tetrahedron (after Walker et al. 1979). See text for discussion. Symbols as in Figure 2. Sample OR00 plots off-diagram to the far left and is not included in this figure.
to the cpx-plag-liq pseudo-univariant curve. Parent magmas have not been calculated, however, actual magmas lie near the low-pressure (1 bar) cotectic for plagioclase saturated liquids, as do most MORBs. Data indicate crystallization, which resulted in fractionated FOR magmas, occurred either at comparatively low pressures or possibly that mineral phases reequilibrated to low pressures during magma ascent (Elthon and Scarfe, 1984; Siroky et al., 1985).

Alteration

If protolith compositions are to be inferred from metamorphic samples, it is important to evaluate whether compositional changes have occurred during seafloor alteration and subsequent metamorphism. Enrichment of Na$_2$O, K$_2$O, and Rb and depletion of CaO are characteristic of altered submarine basalts (Stillman and Williams, 1978; Alt and Honnorez, 1984). To determine which samples, if any, have been affected by seafloor alteration, oxide abundances as a function of differentiation for the FOR schist were compared to normal oceanic crust (cf. Siroky et al., 1985). The compositions of basaltic glasses from the Atlantic, Pacific, and Indian Oceans (Melson et al., 1976) were used to determine the field of oceanic crust (Figure 3). For these calculations it was assumed that Fe$^{3+}$/Fe$^{2+} = 0.15$. It should be noted that oceanic glasses may differ somewhat in composition from crystalline basalts. Glasses may have lower Ca, Al, and Mg, due to possible concentration of olivine and plagioclase phenocrysts in pillow interiors (Hawkins and Melchior, 1985); however, Al in crystalline MORB has a large range and may have either higher or lower abundances than MORB.
glasses (Perfit et al., 1980).

Samples of group 1a are nearly identical to average MORB in FeO and TiO$_2$ content. Al$_2$O$_3$ values plot lower than the oceanic basalt glass field. Most group 1a samples have Na values comparable to oceanic basalts. Samples OR00, SG258, RA72, and PR42 are likely to have had Na addition, probably the result of seafloor alteration. These points lie to the far left on the CMAS projection. Samples OR135, SP25, and RA38 plot only slightly above the ocean basalt field, and may have some Na enrichment. Samples RA76 and RA38 both plot slightly below the MORB field for CaO and above for K$_2$O. Samples OR00 and SG258 plot with very low CaO abundances, and appear to be the samples most affected by seafloor alteration. Although samples 280A and 280B are not MORB-like (below), low CaO abundances and some alteration of epidote in thin section indicate the more mobile major elements of these samples may have been affected by seafloor alteration or later metamorphic events.

The lack of Na enrichment and adherence to fractionation trends suggest that most samples are probably not greatly affected by seafloor alteration.

Trace Elements

Trace elements (e.g., Ta, Hf, Zr, Nb, Cr, Ni) are thought to be less mobile than many of the major elements and less likely to be affected by seafloor alteration and subsequent metamorphism; hence, they are more likely to represent original magmatic characteristics. On a
Ti-Zr diagram (Figure 5), both primitive and evolved basalts of group 1a and group 1b define a linear trend on or near the field of mid-ocean ridge basalts. Samples which are depleted in Ti and Zr are, in general, Mg-rich. Variation of Ti-Zr is parallel to vectors representing fractionation by crystallization of olivine, pyroxene, and plagioclase +/- ilmenite or magnetite (Pearce and Norry, 1979; Piboule and Briand, 1985), affirming the importance of low-pressure fractionation processes in the formation of these magmas. Samples of group 2 have much higher abundances of incompatible elements. These points plot in the field of within-plate basalts and clearly are not related to the fractionation trend of the other samples.

The proportions of Th-Ta-Hf (Figure 6, after Wood et al., 1979) for group 1a samples are similar to those of N-MORB and E-MORB. Group 1b samples have proportionally more Th and Ta compared to Hf, similar to E-MORB ratios or within-plate tholeiites. Group 2 samples have Th-Ta-Hf proportions which are identical to alkalic within-plate basalts.

REE distribution patterns normalized to chondritic abundances are shown in Figure 7. Most group 1a basalts are nearly flat, to somewhat LREE depleted, at about 4x to 30x chondrite abundances ((Ce/Sm)N=0.5 to 1.0). The more fractionated samples have a slight Eu anomaly. Samples from group 1b are enriched in LREE ((Ce/Sm)N=1.0), with chondritic abundances similar to most group 1a samples; however, the HREE are much more depleted than group 1a (5x to 10x). REE distribution patterns indicate that group 1a basalts are MORB to transitional MORB in character (Basaltic Volcanism Study Project, 1981). The nearly flat and
Figure 5. Ti versus Zr variation of POR mafic schists (after Pearce, 1982; Pearce and Norry, 1979). IAB=island arc basalt, WPB=within plate basalt.
Figure 6. Th-Hf-Ta variation of POR mafic schist (after Wood et al., 1979). Samples with values below detection limits are not plotted.
Figure 7. Rare earth element patterns normalized to chondritic abundances for POR mafic schists
parallel nature of these patterns and correlation of REE abundances to Mg# can be attributed to varying extents of crystal fractionation from a common parent magma or similar group of magmas. Group 2 REE abundances are quite unlike the nearly flat MORB-like patterns of group 1a and have higher overall REE concentrations (180x for LREE) and greater REE enrichment factors ((Ce/Sm)_N=2.0).

In Figure 8, various elements normalized to MORB values (Pearce et al., 1981) are plotted for the 19 samples analyzed by INAA. The elements are arranged on the basis of increasing ionic potential and increasing D-value to the right (D garnet lherzolite/melt; see Pearce et al., 1984; Saunders and Tarney, 1984). The large ion lithophile (LIL) elements (Sr–Rb), commonly mobile during metamorphism, are plotted on the left side of the diagram. The high field strength elements (HFS) are plotted to the right (Ta–Yb) and are increasingly more stable. The observed variation in abundance of incompatible elements (Ta, Nb, Zr, Hf, Ti, Y) between samples is expected if crystal fractionation was affecting the chemical variability of the magma. The samples of POR schist have variable enrichment and depletion of Sr and Rb compared to N-MORB. Enrichment of low ionic potential elements Sr, K, and, to a lesser extent Rb, may occur during seawater alteration (Hart, 1970) and spilitization (Vallance, 1974). Mobilization of these elements, along with the alkali and alkaline-earth elements, may again take place during subsequent metamorphism. In spite of the potential for mobilization, Sr, Rb, and K values are generally similar to MORB and E-MORB values for group 1a and group 1b samples.
Figure 8. Large ion lithophile (LIL) and high field strength (HFS) element patterns for selected POR metabasites. Data are normalized at N-MORB values of (Pearce, 1982). Bracketed points could not be separated from background, actual values are less than those plotted. A,B,C,F=group1a; D=group1b; E=group2
Figures 8A-C and F show group la elemental abundances. The patterns are similar to those of MORB or E-MORB, being generally flat to slightly enriched in Th-Ce. Uniform differences in elemental abundances between sample patterns can be correlated to Mg-number and are attributed to varying amounts of crystal fractionation. Unlike group la, in which the least incompatible elements Y and Yb are near or enriched relative to N-MORB values, group lb (Figure 8D) shows a depletion relative to MORB. Because of lower Y and Yb values, increasing enrichment with decreasing D-value is more pronounced than in group 1a samples which are generally flat (Figure 8B, C, and F) or show enrichment only from Zr through Th (Figure 8A). The exception to the group lb enrichment trend is the obvious negative phosphorus anomaly, which is seen only in samples from the Rand Schist. Enrichment of group 2 patterns (Figure 8E) is similar to group lb, however, abundances of group 2 samples are substantially greater and lack the phosphorus anomaly.
DISCUSSION

The POR mafic samples fall into distinct groups based on bulk-rock compositions. Group 1a metabasalts have normative compositions and sympathetic enrichment of iron and alkalis typical of tholeiitic basalts. A MORB-like character of the group 1a samples is shown by systematic variation of iron and magnesium with TiO₂, variation within the ol-cpx-plag-qtz modified basalt tetrahedron, and trace element abundances, confirming the preliminary reports of Bennett and DePaolo (1982) and Haxel et al. (1986). Enrichment and abundances of incompatible trace elements in some samples (e.g., Orocopia Schist, Figure 8A) are indicative of E-MORB.

Group 2 samples may be alkaline to subalkaline with Y/Nb < 0.6. These samples have low abundances of compatible elements and high abundances of Zr, Y, Hf, Ta, Th, and LREE. Incompatible element proportions indicate these basalts were erupted in a within-plate tectonic setting. Incompatible element abundances exceed P-MORB (Sun et al., 1979) and resemble those of off-axis (i.e., within-plate) plume volcanics (e.g., Bowie seamounts, west coast of British Columbia: Cousens et al., 1985; Loihi Seamount: Hawkins and Melchior, 1983; and other locations: Basaltic Volcanism Study Project, 1981). REE patterns similar to those of group 2 are observed in basalts from the Azores (Schilling et al., 1983), Walvis ridge DSDP sites, particularly site 528 (Humphris and Thompson, 1983), and dredged basalts from Iceland (Wood, 1976).

Group 1b samples are tholeiitic and similar to group 1a samples but
show characteristics commonly associated with tholeiitic "within-plate" or perhaps E-MORB mantle enrichments. Incompatible elements of group 1b generally show depletion of Y and Yb (the least incompatible elements) and enrichment with increasing elemental incompatibility (Figures 7 and 8D). Enrichment is also seen in higher Th and Ta values relative to Hf (Figure 6). The rare earth element trends of group 1b are LREE enriched and HREE depleted relative to N-MORB because of low abundances of Yb and Lu.

Interpretation of group 1b samples is not straightforward. All group 1b samples and RA72 are from a single extensive mafic body in the Rand Mountains. Three samples from group 1b (group 1b*) show characteristics commonly associated with cumulates. For example, they have high MgO values (Table 1), low FeO/MgO values and extremely high abundances of the compatible elements Ni and Cr. These samples also differ from other group 1b samples because of lower abundances of low ionic potential elements (Sr, K, and Th). These characteristics could be explained by the accumulation of olivine and clinopyroxene. However, if these samples were produced only by crystal accumulation, one would also expect TiO₂ and other relatively incompatible elements (LREE, Th, Ta, Nb, Hf, and Zr) to be depleted, which is not the case. Mixing of a partial melt or highly fractionated liquid with cumulate crystals may account for the observed anomalous enrichment of compatible and incompatible elements. A small amount of partial melting is also compatible with a within-plate tholeiite origin of these basalts, leading to enrichment of incompatible elements and depletion of the
least incompatible elements. An alternative explanation may be that garnet and apatite were stabilized during a hydrous melting event, leading to Y, Yb, and P anomalies in the Rand group 1b patterns. If these anomalies were absent, Rand group 1b patterns would be similar to the enriched Orocopia group 1a patterns (Figures 8A and D); however, the greater enrichments relative to Y and Yb and the P anomalies of group 1b imply a slightly different environment of origin. Additionally, only group 1b samples show high Cr, Ni, and MgO. If this variation is the result of interaction with cumulate crystals, their location in the topmost portion of the mafic body implies the body has been overturned during subduction.

None of the POR samples show island arc affinities. Island arc tholeiites are characterized by enrichment in low ionic potential elements (Sr-Th) and depletion in high ionic potential elements (Ta-Y: Perfit et al., 1980; Pearce, 1982) relative to MORB. Under conditions of hydrous melting or metasomatism, as found in orogenic systems, Ta is partitioned into stable Ti bearing phases such as sphene, rutile, or ilmenite (Weaver and Tarney, 1981). Ta and Nb depletion is not observed in the POR schist samples. Using these criteria, as well as Ta-Hf-Th, TiO₂/FeO*/MgO, and Ti-Zr ratios, the POR schist samples do not resemble island arc basalts. Rocks high in Mg, Cr, and Ni with low FeO/MgO, similar to group 1b samples, have been interpreted as primitive island arc tholeiites (Weaver et al., 1984), however, group 1b samples do not have the low trace-element abundances (Nb, Zr, Ti, REE) seen in noncumulate boninites or komatiites (Cameron et al., 1979). Group 2 and
group 1b samples are enriched in both the LIL and the high field strength elements.

The analyzed samples show no obvious evidence of eruption in a rifted back-arc basin (Mariana-type). The samples are not depleted in high ionic potential elements nor enriched in Th and other large-ion lithophile elements. Enrichment in Sr, K, Th, +/-La, +/-Ce is indicative of magmas erupted in successive backarcs (e.g., Mariana and Lau Basins, Hawkins and Melchior, 1985) and probably is the result of mantle source modification due to hydrous fluids or partial melts derived from a subducting slab (Hawkesworth et al., 1977; Saunders et al., 1980; Tarney et al., 1981). Group 1a lacks this "subduction component". The POR schist samples are not enriched in Th and other LIL elements. In addition, there is no indication of subduction of island-arc tholeiites which might be expected in the closing of a rifted basin and there is no indication of boninites or calc-alkaline volcanics, which are usually associated with initial stages of back arc extension and narrow ensialic basins (Pearce et al., 1984; Crawford et al., 1981).

Although there is a lack of support for the closing of a rifted back-arc basin, relatively low LILE abundance and lack of a "subduction component" are not unequivocal evidence against this origin of the protolith and a marginal basin origin of the protolith remains a possibility. Transitional to N-type MORBs, lacking LILE enrichment, are thought to occur in incipient ensialic basins, wide basins or those formed by entrapment of older lithosphere (Saunders and Tarney, 1984). The occurrence of MORB-like basalts and seamount-like basalts within the
POR subduction complex is compatible with several of the tectonic models which have been proposed for the schist. Oceanic crust could be incorporated into the complex if the subduction zone were east dipping off the continental margin (Franciscan-type). Marginal-basin models are consistent with these data if they formed either by capture of oceanic crust between the continent and an exotic block of continental crust or if a marginal basin formed by transform tectonics (Gulf of California-type). The occurrence of seamounts or ocean islands does not necessarily imply an open ocean origin. Seamount development is observed in both open-ocean and marginal-basin environments (e.g., Hawkins, 1976). However, if back-arc basin models are to be applied, the basin must have reached a high degree of maturity if seamounts were to develop.

The occurrence of high-Ti MORB-like basalt, transitional types and seamount-like basalts implies that the POR protolith was oceanic crust. The occurrence of seamount-like basalts, bounded both above and below by metagraywacke, also implies that at least some of the mafic bodies were emplaced tectonically and are not in situ flows or sills. If the majority of the mafic rocks are in situ, they must represent off-axis flows. In outcrop, POR mafic bodies can be separated by meters to hundreds of meters of quartzofeldspathic metasediments. In basins with high sedimentation rates, such as the Gulf of California, flows separated by only 15 meters of sediment are significantly younger than the basement basalts (1.8 my and 3.2 my respectively; hole 474, DSDP, p. 599). Based on 3 cm/yr half-spreading rates, these sills or flows may
have erupted up to 40 km from the spreading center (Saunders et al., 1983; Leg 64, DSDP). Off-axis flows may be distinguished from axis flows and layer 2 by Th/Hf abundances. When the POR schist samples are compared to axis and off-axis flows in the Gulf of California (hole 474, Leg 64) and off-axis flows in the Nauru Basin (hole 262, Leg 89; data of Floyd, 1984; Saunders, 1984), the POR schists more closely resemble oceanic layer 2 and axial flows from the Gulf of California (Figure 9).

MORB-like basalts and mafic rocks with geochemical characteristics similar to ocean islands or seamounts have been identified elsewhere in subduction melange (e.g., Wasowski and Jacobi, 1985) and modern trenches (Bloomer, 1983). Incorporation of slivers of subducting oceanic crust into accretionary wedges is common (Ernst, 1975; Karig, 1982) and topographic irregularities such as seamounts are more likely to be incorporated and deformed (Sills and Tarney, 1984; Storey and Meneilley, 1985).
Figure 9. Hf vs. Th variation for POR mafic schists, off-axis and axis basalts for the Gulf of California (Deep Sea Drilling Project (DSDP) leg 64, hole 474; Saunders et al., 1983) and off-axis basalts from the Nauru Basin (leg 89, data from Saunders, 1984; Floyd, 1984)
The POR mafic samples may be divided into several distinct chemical groups. Group 1a metabasalts are tholeiitic and have MORB-like compositions. Variation of major and trace elements can be attributed to extensive modification by crystal fractionation. Group 2 samples have high contents of Zr, P_2O_5, Y, and REE and show high LREE enrichment. Trace-element variations imply that these basalts were erupted within an ocean basin as the result of hot spot volcanism. High concentrations of incompatible elements suggest the source was not depleted in high ionic potential elements. A third group (group 1b) has elemental abundances similar to group 1a; however, Y, Yb, Lu, and P are more depleted. Several group 1b samples (group 1b*) have high MgO, Ni, and Cr, characteristics which are commonly associated with crystal accumulation. Group 1b samples may be within-plate tholeiites and derived from a source compositionally different from that of the other samples. Based on these data, the mafic schist of the POR probably represent remnants of oceanic crust and an oceanic island or seamount. Associated metamorphosed carbonates, ferromanganeseous cherts, relict pillow basalts, and Mn-nodules confirm this conclusion. It is more likely that the blocks are tectonically dismembered ocean floor or axis sills rather than in situ off-axis flows or tuff. Although the mafic schists may be derived from two or possibly three sources, the variety of compositions within and between the POR occurrences does not imply different tectonic environments for each locality, but may simply be a statement about the variety of compositions that can be incorporated.
into a subduction zone.

Low Ti, Ta, and Nb basalts characteristic of modern island arcs were not seen, nor do the samples show characteristics transitional to island arc magmas, as seen in mature back-arc basins, which are depleted in HFS elements from previous melting and enriched in LIL elements. FOR sample compositions are compatible with an open-ocean origin or a marginal basin formed by entrapment of older oceanic crust or transform-style tectonics (Gulf of California-type).
REFERENCES


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CHAPTER II.

STRUCTURAL GEOLOGY OF THE OROCOPIA SCHIST,
OROCOPIA MOUNTAINS, SOUTHERN CALIFORNIA
ABSTRACT

The Orocopia thrust of southern California is a segment of a regional system termed the Vincent-Chocolate Mountains thrust, which has been interpreted as a Late Cretaceous subduction zone. The Orocopia Schist of the lower plate is thought to represent an accretionary wedge.

The earliest deformational fabric of the Orocopia Schist of the Orocopia Mountains (OMS) is preserved as a foliation of graphite in pseudomorphed porphyroblasts. This foliation ($S_1$) has been transposed at least twice during deformation to produce the current matrix. The graphite layers are folded to produce either shear- or buckle-type of folds. The presence of relict crenulations in the matrix and graphite trails of pseudomorphed porphyroblasts oblique to and truncated by external inclusion trails within albite porphyroblasts indicate that a process of foliation regeneration produced the fabric in the Orocopia Schist.

The youngest structures are open to tight folds ("style 2 folds") which fold schistosity and compositional layering. Style 2 folds, in some cases, overprint isoclinal folds ("style 1 folds"). Both styles of folds, as well as stretching lineations, have variable orientations.

Evidence from the Orocopia Mountains indicates movement direction, in some areas, was oblique to an earlier shear strain. This is seen on an outcrop scale by quartzite layers with cylindrical asymmetric folds which have axes oriented in different directions and rootless folds of one axial orientation overlying folds with a different axial orientation.
Models which best explain these relationships and the parallelism of open folds and stretching lineations are most commonly applied to shallow thrust zones. These features may be the result of "packages" of material, on a kilometer or larger scale, moving at either different rates and/or different times. If package walls become pinned, strain can become locally reoriented, and deformation may be partitioned around preexisting structures.

This model implies a continuous process for the formation of folds, depending on the movement of material packages, rather than open folds forming during specific "generations".

A consistent fold asymmetry implies material movement, at the time of style 2 folding, was generally eastward. However, the lack of mylonite at the fault contact and extensive mineralization at the fault suggest reactivation of the Orocopia thrust.
INTRODUCTION

The Vincent-Chocolate Mountains thrust of southern California (Haxel and Dillon, 1978) separates continental upper-plate rocks from lower-plate rocks which are oceanic in origin. The upper plate consists of Precambrian, Paleozoic, and Mesozoic igneous and metamorphic rocks. Lower plate rocks consist of interlayered graphitic mica/quartz/feldspar schists, metabasites, quartzites, and marbles and are referred to as the Pelona, Orocopia, and Rand (POR) Schists. Deformation and metamorphism of the lower plate occurred during juxtaposition with the upper plate. This interpretation is based on (1) similarity in the orientation of foliation, folds, and lineations of the lower plate and mylonites, (2) similarity in metamorphic grade of both plates near the thrust and interposed mylonites, and (3) similarity of radiometric ages of lower-plate schists and mylonites (50 to 60 m.y. B.P., Haxel and Dillon, 1978; Ehlig, 1981). Metamorphic pressures indicate deformation in a deep crustal environment (20 to 30 km, Graham and England, 1976; Graham and Powell, 1984). Given the lithologies of both plates, lower-plate schists are commonly interpreted to represent a relict accretionary prism.

Several features of the POR schist cause interpretation to be nonplus. Because the schists and thrusts are located inboard of the continental margin, and because a northeast vergence has been inferred for the thrust in the Chocolate Mountains and vicinity (Haxel and Dillon, 1978), tectonic models have been proposed which suggest deformation took place in a west-dipping subduction zone between North
America and an outboard island arc or continental fragment. This subduction zone would be responsible for either the closing of a basin formed by rifting of the continent (Haxel and Dillon, 1978; Tosdal et al., 1984); collision of an exotic block into the continent (Ehlig, 1981; Vedder et al., 1983); or the closing of a basin which formed as the result of a mega-shear event (Haxel et al., in press). Because a suture zone has not been found inboard of the schist, and because the POR schists have rock types of oceanic character similar to the Franciscan Complex of northern California, other workers (Yeats, 1968; Burchfiel and Davis, 1981; Crowell, 1981) favor formation of the POR schists within an east-dipping subduction zone.

The most definitive evidence, but certainly not decisive, is the inferred movement direction of the thrust. Interpretation has been complicated by two factors: (1) the orientation of fold axes in the Pelona Schist of the San Gabriel Mountains is 90° to the orientation of fold axes in the Orocopia Schist of the Chocolate Mountains and (2) it is uncertain whether the folds and kinematic indicators used to determine movement direction formed during subduction or uplift.

A northeastern direction of overthrusting was inferred in the San Gabriel Mountains by Ehlig (1958, 1981) based on interpretation of a large macroscopic synform beneath the mylonites as a drag feature. Complicating this interpretation is the presence of a stretching lineation parallel to the axis of the synform (Jacobson, 1983a) and the possibility that the synform is the remains of a larger asymmetric structure which has been truncated by the thrust (Burchfiel and Davis,
The northeast direction of thrusting was later confirmed by Haxel and Dillon (1978) and Dillon et al. (in press) in the Chocolate Mountains by applying the methods of Hansen (1967) to minor folds and by the use of other kinematic indicators. The studies of Jacobson (1983a) in the San Gabriel Mountains, and Postlethwaite (1983) and Postlethwaite and Jacobson (1987) in the Rand Mountains yielded seemingly conflicting movement directions within those areas.

Because folds and additional kinematic indicators used to interpret movements postdate schistosity (Burchfiel and Davis, 1981; Crowell, 1981), and because of evidence of reactivation of the Rand thrust, Postlethwaite and Jacobson (1987), and Jacobson et al. (in press) argued that the minor folds could have formed after thrusting, presumably due to uplift. Jacobson (1983b, 1984) argued that open folds in the San Gabriel Mountains formed during thrusting, based on an interpretation of a prograde environment for the recrystallization of crenulated muscovites to more sodic and celadonitic compositions.

The ultimate goal of this study was to constrain tectonic models for the origin of the POR schists. More specifically the purpose of this study was three-fold: (1) The macroscopic structures of the Orocopia Schist in the Orocopia Mountains were studied to compare movement directions in the Orocopia and Chocolate Mountains. This represents an expansion of a previous study by Raleigh (1958), conducted in the extreme northwestern portion of the Orocopia Mountains. (2) Microstructures within albite porphyroblasts were examined and interpreted based on models of deformation partitioning (Bell, 1981;
1985) and progressive development of crenulation cleavage (e.g., Bell and Rubenach, 1983; Bell, Fleming and Rubenach, 1986; Jamieson and Vernon, 1987), to determine earlier stages of the deformational history; and to constrain pressure-temperature-time histories of deformation and metamorphism. (3) Folds and microstructures were studied to look for evidence which might indicate how folds in the FOR schist formed, which in turn could aid in interpretation of stretching lineations, fold orientations and folding styles.
DESCRIPTION OF LITHOLOGIES

The Orocopia Schist is probably correlative to the Pelona Schist and possibly correlative to the Rand Schist (Ehlig, 1958, 1981; Raleigh, 1958; Jacobson, 1980). Although mineral assemblages vary with the grade of each specific locality, the bulk chemistry and field appearance of lithologies are, in a general sense, very similar between localities (see Chapter 1 for bulk chemistry discussion). The principal variable between and within POR localities seems to be the relative abundance and positioning of lithologies.

In the Orocopia Mountains, and in all POR occurrences, the predominant lithology is a quartzofeldspathic schist interpreted to be a metagraywacke (Ehlig, 1958; Haxel et al., 1986). The schist typically contains albite porphyroblasts within a matrix of quartz and muscovite. Inclusions of graphite are common in the albite, even to the extent of giving the albite a gray to black color in hand sample. In this paper, quartzofeldspathic schists will be referred to as "grayschists". Epidote, chlorite, sphene, and calcite are common accessory minerals; actinolite or hornblende, and garnet are occasionally observed.

Mafic schists comprise about 10% of the total formation. They are basaltic in composition and contain porphyroblastic albite with epidote, calcic amphibole, and chlorite as the common mineral assemblage. At some localities in the Orocopia Mountains, variations in the size and abundance of albite and epidote define a compositional layering. The mafic schists can occur as laterally continuous bands which vary in thickness from less than a meter to thicknesses of 50 to 100 meters.
Raleigh (1958) noted that mafic schist which is interlaminated with quartzofeldspathic schists tends to be more micaceous than the massive mafic units. The mafic schists commonly display a schistosity parallel to the compositional layering, except in the hinge area of folds.

Quartzites and ferromanganiferous metacherts are commonly associated with mafic schists in the POR schist, however, quartzite layers also occur bounded by quartzofeldspathic schists. The metacherts are between 50% to 90% quartz; common accessories are calcite, biotite, spessartine garnet, muscovite, chlorite, hornblende or actinolite, apatite, and opaque minerals. In this paper the term "quartzite" will include quartz-rich (>50%) quartzofeldspathic rocks. The quartzites are commonly associated with marble.

Also present in the schist are isolated pods of serpentine and actinolite, and small pods (<0.5 m) of manganese oxides.

Mineral assemblages indicate albite-epidote amphibolite facies metamorphism. There is no apparent inverted metamorphism as is seen in other POR localities (Ehlig 1958, 1981; Graham and England, 1976; Jacobson, 1983a; Postlethwaite and Jacobson, 1987).
STRUCTURE

The Orocopia Schist, like all other POR occurrences, is exposed in a post-metamorphic arch which in this area trends NNW-SSE. The anticline is well defined, except in the western portion of the range (Plate 1). Minor faulting of the anticline is observed at some localities in the range. The contact between the schists and upper-plate rocks is exposed on the north limb of the anticline. Mylonitic rocks, developed primarily in the lower plate, are locally present and range up to several meters in thickness. Where mylonites are absent, there is a sharp contact between the upper and lower plates and, where present, the degree of mylonitization is variable. There is extensive oxidation and mineralization along the fault, and folds are not readily observable in either the mylonites or the upper plate. Lineations in the mylonite are variably oriented and consist of fine ribbing on foliation surfaces, elongated clusters of minerals, and pressure shadows.

Although very few folds were found in the upper plate, the lower plate is pervasively folded. Previous workers in the Orocopia, Pelona, and Rand Schists have all recognized that deformation was the result of superposed strains (Raleigh, 1958; Harvill, 1969; Vargo, 1972; Haxel and Dillon, 1978; Jacobson, 1983a, 1983b). The method of fold analysis commonly applied to the POR schists is based on that of Weiss and McIntyre (1957) in which rocks of the Loch Leven area in western Scotland were classified by similarities of fold style and orientation. Although the use of style has been criticized by some authors (e.g.,
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Plate 1. Orientations of style 1 and style 2 folds in the Orocopia Mountains, southern California
EXPLANATION

Axis of anticline

Orocopia thrust

Teeth on upper plate,
dotted where concealed
EXPLANATION

Axis of anticline

Orocopia thrust
teeth on upper plate,
dotted where concealed

trend and plunge of style 1 fold axis

trend and plunge of style 2 fold axis

Thrust contact and anticline mapped by
Jonathan Matti, Bob Powell, and Ed
Rodriguez (unpublished)
Williams, 1970, 1985), this method appears to work well in some locations (e.g., Pelona Schist, San Gabriel Mountains: Jacobson, 1983a) where folds could be classified based on the tightness of fold hinges. Isoclinal folds are referred to as "style 1 folds", open and tight folds are referred to as "style 2 folds". It is generally assumed that the isoclinal folds formed earlier than the more open folds, although it has been argued that both styles formed by a continuous process of folding during thrusting (Jacobson, 1983a). It will be argued later in this paper that in the Orocopia Mountains it not possible to assume the more open folds are earlier than the isoclinal folds. I have grouped the folds in the Orocopia Schist based on this criteria of style, although it is not believed that these groups represent folds that can be temporally separated.

There are two problems in applying a style classification to folds in the Orocopia Schist. First, the occurrence of a particular style of fold is obviously affected by the type of lithology present and, where the rocks are inhomogeneous, by the relative thicknesses and competencies of the different layers. This was also noted by Raleigh (1958) in the Orocopias, in the Rands by Vargo (1972), and in the San Gabriel Mountains by Jacobson (1983a). Isoclinal folds are most commonly seen in quartzites for two reasons: (1) As a fold hinge tightens, a strong axial planar schistosity will develop and form a weathering surface in the grayschists. (2) Quartzites are less likely to be overprinted by crenulation cleavage which tend to obscure isoclinal folds in the grayschist. The second problem with using style
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Plate 2. Orientations of schistosity in the Orocopia Mountains, southern California
EXPLANATION

Axis of anticline

Orocoaia thrust
toth on upper plate,
dotted where concealed
EXPLANATION

Axis of anticline

Oroopia thrust
teeth on upper plate,
dotted where concealed

Strike and dip

Thrust contact and anticline mapped by
Jonathan Matti, Bob Powell, and Ed
Rodriguez (unpublished)

ROAD CLASSIFICATION

Medium-duty
Light-duty
Unimproved dirt
State Route

MORTMAR, CALIF.
SW/4 COTTONWOOD SPRING IS QUADRANGLE,
N3330—W11552.5/7.5

1958
PHOTOREVISED 1974
AMS 2851 III SW—SERIES V895

Mapped, edited, and published by the Geological Survey
Control by USGS, USC&GS, USBR, and USCE
Topography from aerial photographs by ER-55 plotter
Aerial photographs taken 1956. Field check 1958
Polyconic projection. 1927 North American datum
10,000-foot grid based on California coordinate system, zone 6
1000-metre Universal Transverse Mercator grid ticks,
zone 11, shown in blue
Dashed land lines indicate approximate locations
Unchecked elevations are shown in brown
ROAD CLASSIFICATION

- - - - - - -
Light-duty

- - - - - - -
Unimproved dirt

OROCOPIA CANYON, CALIF.
SE/4 COTTONWOOD SPRING 15 QUADRANGLE
N3330—W11545/7.5

1958

WITH NATIONAL MAP ACCURACY STANDARDS

EVE, DENVER, COLORADO 80225, OR RESTON, VIRGINIA 22092

RAPID MAPS AND SYMBOLS IS AVAILABLE ON REQUEST
to classify the folds is that, unlike the Pelona Schist, where folds fall into distinct groups, in the Orocopia Schist there is a continuous range in tightness from isoclinal to open. In most areas, there is a continuous range of fold axes orientation, although there is little or no correlation between fold tightness and orientations.

**Style 1 Folds**

Isoclinal folds occur throughout the Orocopia Schist (Plate 2, back map pocket), and, in several locations, isoclinally refolded folds were observed (present study and Jacobson, 1983b). Isoclinal folds, as at other POR localities, are most evident in quartzites. Very few are preserved, or observable, in the grayschist, particularly where relatively open folds are present. Isoclinal folds have an accompanying axial planar schistosity defined by layering of muscovite. Examples are shown in Plate 3A, D, and E.

Isoclinal folds in the quartzites were observed with amplitudes up to several meters. Isoclinal folds with amplitudes of 50 m have been reported in the San Gabriel Mountains; consequently, larger isoclinal folds may be present but not visible because of the weathered nature of the outcrops, or they may have been present but are now overprinted by large-scale open folds.

**Style 2 Folds**

Most folds in the Orocopia Mountains are open folds with little or no development of an axial planar schistosity (Plate 3B, C). Open to tight
Plate 3. Variations in fold style in the Orocopia Mountains. (A) isoclinal "style 1" folds (B) open "style 2" fold (C) large style 2 fold (D) isoclinal fold which was refolded by style 2 folding and sheared
Plate 3. (continued)  (E) open and isoclinal folds of similar orientations, note sheared limb of lower isoclinal fold  (F) outcrop of rootless folds, fold in upper center has an axes into the outcrop, fold in lower photo has axes parallel to the outcrop (see text for discussion)  (G) Sheared layers and floating hinges below rootless folds  (H) asymmetric fold with two axial directions, upper hinge is parallel to outcrop and the lower hinge is at 90°
style 2 folds are not restricted to the vicinity of the contact with the upper plate, as seen in the San Gabriel and Chocolate Mountains, but are abundant throughout the greenschist (plate 2, back map pocket). These folds are generally asymmetrical and fold both compositional layering and schistosity. The axial surfaces of style 2 folds are commonly oblique to compositional layering, and, in some folds, a new axial planar schistosity is present. Spatially associated with most style 2 folds is a strong crenulation of an earlier foliation.

Style 2 folds in quartzites are generally cylindrical, with hinges which can be traced for up to ten meters. Folds in quartzite commonly have thickened hinges and tapered limbs. Parasitic folds are common.

Interlayered lithologies may have different fold styles than adjacent layers, presumably controlled by the different competencies of the layers. In layers of a single composition, such as quartzite, thinner layers tend to have smaller apical angles. It is also not uncommon for a fold hinge to range from gentle to tight through successive layers. Where the axial surface is curved, the tighter folds are more recumbent.

Some folds have one or both limbs truncated along a surface subparallel to the axial plane (Plate 3E). The truncated limbs in some cases appear fractured and, in others, pinched out. In some cases, folds with different orientations are separated by small shear zones. Some folds abut the attenuated limbs of other folds, or folded layers juxtapose non-folded layers which parallel schistosity. Shear surfaces were also seen in marble and quartzite layers in which one surface
truncates another.

Plate 3E shows two apparently isoclinal folds in which the upper limbs are sheared, however, a portion of the upper isoclinal fold can be traced into an overlying layer which contains several asymmetric folds.

At one outcrop (in subarea 5, Figure 10) rootless fold hinges appear within a zone 10 to 15 m thick (Plate 3F). Hinges are bounded by zones of relatively high shear strain, which juxtapose schistosity or other folds. The hinges of several larger rootless folds contain mafic schist, which is not seen elsewhere in the outcrop. Many of the larger rootless folds have quartz veins at their base. About 10 rootless folds were observed at this outcrop. The asymmetric ones all verge to the southeast. The fold axes are subparallel to a stretching lineation direction that tends northeast. Other folds in this zone, where limbs can be traced to some lateral extent, have a greater variability of orientations (Figure 11).

The folds below this zone have a fairly uniform orientation approximately perpendicular to that of the rootless folds (S10E, Figure 11). These folds are defined by quartz-rich grayschist layers (up to 0.5 m in thickness) interlayered with more micaceous layers.

Refolded Folds

In the outcrop described above, refolded folds and sheath folds occur near most of the larger rootless fold hinges. Examples of these folds are shown in Figure 12A-C. Very good exposures, in many cases, allowed determination of the shape and orientation of the axes of both early and late folds. Figure 12A is a series of folds in which the
Figure 10. Location of subareas within the western portion of the Oroopia Mountains. Reference Plate 1 (back map pocket) for location
Figure 11. Stereonet showing the orientation of rootless folds (axes oriented to the northeast) and the orientation of fold axes from folds immediately below the rootless folds. CCW=counterclockwise sense of shear, CW=clockwise sense of shear as seen when viewed down plunge.
Figure 12. Examples of refolded folds from the Orocopia Mountains. (A–C) Cartoons of refolded folds. (D) Cartoon depicting model for deformation partitioning with strain reorientation (see text for discussion).
upper-most folds are cylindrical and the lower ones are non-cylindrical, sheath folds. Figure 12B shows two asymmetric folds which are vertically juxtaposed and refolded about later axes of different orientations. Figure 12C is an isoclinal fold which has been asymmetrically refolded. The upper hinge of this fold was the first to form, is closest to a zone of shear, and has the same orientation as the rootless hinges.

Orientations

Figure 13 shows the orientations of style 1 and style 2 folds and lineations for the western Orocopias and for the north and south limbs of the anticline defined by schistosity. Style 1 lineations include ribbings on quartzite surfaces, muscovite streakings, intersection lineations, and mullion and rod structures. Style 2 lineations are primarily crenulation and intersection lineations. As noted earlier, folds were classified as style 1 or style 2 based primarily on their tightness and on the presence or absence of an axial planar schistosity. As can be seen from Figure 13, there is a large variability of orientations throughout the range. This is most noticeable in the orientations of style 2 folds in the western area. Figure 14 shows poles to schistosity for both limbs of the anticline and the western area. Figure 15 shows fold and lineation orientations of the western area divided into subareas (subarea locations are shown in Figure 10). As seen in this figure, there is a variability in orientation within subareas. This is particularly striking in subarea 4, where style 1 and style 2 folds appear to cluster but at different orientations.
Figure 13. Orientations of style 1 and style 2 fold axes and lineations for north and south limbs of the anticline and the western area
Figure 14. Orientations of poles to schistosity for north and south limbs of the anticline and the western area.
Figure 15. Orientations of style 1 and style 2 fold axes and lineations for subareas of the western area. Subareas are defined in Figure 10.
Abundant albite porphyroblasts, commonly with inclusions of quartz and graphite, are a characteristic of the grayschist of the Pelona and Orocopia Schists (e.g., Haxel and Dillon, 1978). Jacobson (1983a) noted that the graphite inclusions define complex patterns of internal schistosity and that albite porphyroblasts grew at different stages of fold and crenulation cleavage development.

Albite porphyroblasts in both the mafic schists and grayschists are subidioblastic and variably poikiloblastic. Albite porphyroblasts in the mafic schist are typically 1-5 mm but range up to several centimeters in samples with high sodium contents. Preservation of previous deformational features is not evident because of the coarse-grained nature of included quartz, epidote, sphene, actinolite, muscovite, and calcite in the mafic schist albites.

In the grayschists, a strong matrix foliation is defined mainly by muscovite, biotite, chlorite, and quartz bands. Commonly, the foliation is axial planar to matrix crenulations. Porphyroblasts vary from highly poikiloblastic to relatively inclusion free. The matrix foliation commonly wraps around porphyroblasts, truncating inclusion trails, but in some albite grains inclusion trails can be traced into the matrix. Quartz- and chlorite-rich strain shadows are common in the margins of porphyroblasts. Also included in some albite porphyroblasts are idioblastic to sub-idioblastic garnet and sphene (also with included graphite), muscovite, epidote, and actinolite. Garnet inclusions are usually unaltered and idioblastic, but, in some cases, garnet is
partially retrograded to chlorite, which retains the idioblastic shape of the original garnet. Generally, sphene and garnet are euhedral only where protected by albite. Matrix garnet is commonly partially altered to chlorite.

In some albite porphyroblasts, which contain an internal graphite foliation, the graphite is very finely disseminated and defines small equidimensional pseudomorph porphyroblasts with well-developed pressure shadows and asymmetric pressure trails. These can be seen wrapped by the main internal foliation within the albite (Plate 4). The pseudomorphed porphyroblasts have either opposing relict pressure shadows with a micaceous form or elongated, sometimes asymmetric trails resembling quartz layers. In the larger pseudomorphs, an internal foliation of graphite is oblique to and truncated by the graphite foliation in the main portion of the albite porphyroblast. Other pseudomorphs have an elongated form and also contain inclusion trails oblique to and truncated by the main albite foliation. In albite grains where pseudomorphed porphyroblasts are found, the graphite trails are generally straight and at a late stage of crenulation development (stage 5 or 6 of Bell and Rubenach, 1983). These albite grains have fine layers of concentrated graphite alternating with layers having less graphite. Garnet, sphene, and albite all overgrew the same graphite schistosity.

In several samples, graphite trails are "broken" by sub-circular zones where no graphite is present (Plate 4A, B, D, F, G) or are absent from the porphyroblast edges, generally on sides adjacent to pressure
Plate 4. Microstructures within albite porphyroblasts in grayschist of the Orocopia Schist. Small pseudomorphed porphyroblasts and pressure shadows can be observed in all photos (see text for discussion; width of the field of view is 696 μm for all photographs). (A) circular and concentric clear-zones in albite graphite trails (B) clear zones in graphite inclusion trails within a single albite porphyroblast (center of photo) (D) same as B but crossed polars (C) several albite grains with pseudomorphs in graphite inclusion trails, note solution seams and concentrations of graphite adjacent to albite porphyroblasts.
Plate 4. (continued) (E) relict crenulation in albite (F) euhedral garnet protected by albite porphyroblast (G) clear-zone in albite outside of which the graphite inclusion trails change orientation with the beginning of crenulation (H) pseudomorphed porphyroblast within albite showing S1 foliation and relict pressure shadows
shadows (Plate 4A-H). In these porphyroblasts, spacing between the
graphite trails is smaller than in the matrix or than in albites which
have overgrown local crenulations.

Both style 1 and style 2 folds commonly fold these previous
graphite layers, as well as a compositional layering defined by albite
porphyroblasts and quartz-rich layers. Albite grains kinked and
overprinted by later folding and albite grains with large crenulations
of inclusion trails which can be traced into the matrix imply albite
grew both before and after crenulation of the matrix. When 3" X 4" thin
sections of varying styles of fold hinges were observed, two types of
folding processes were seen. Macroscopic folding disrupted the graphite
foliation by either buckle-type folds, in which the graphite layers are
parallel to the folded compositional layering, or by shear-type folds,
in which the graphite layers are oriented perpendicular to the axial
surface rather than parallel to folded compositional layers.

In some folds, albite porphyroblasts contain relict asymmetric
pressure shadows which show the opposite sense of asymmetry as pressure
shadows within the matrix.
DISCUSSION

Interpretation of Microstructures

The earliest deformational event to affect the Orocopia Schist is recorded as a foliation of graphite in pseudomorphs which have been included by some albite porphyroblasts. Because this foliation is oblique to and truncated by the foliation of the albite, the pseudomorphed porphyroblasts are interpreted to have overgrown either a previous foliation (S₁) or an early stage of the albite foliation. These observations and the relict pressure shadows and tails, some of which are asymmetric, imply that the foliation of graphite within the albite is S₂ and developed by crenulation of an earlier foliation, S₁.

The foliation of the Orocopia grayschist has been transposed at least twice during deformation. The earliest deformation produced quartz bands with finely foliated graphite and mica layers. Small porphyroblasts which had overgrown the early foliation developed asymmetric quartz and mica pressure shadows. Because these porphyroblasts exist only as pseudomorphs within albite porphyroblasts, their composition is not known.

It is not clear what inclusion-free zones within and at the margins of albite porphyroblasts represent. The lack of graphite at the margins of albite porphyroblasts, which otherwise have graphite inclusions, has been attributed in other areas to a decarbonization event (Bell and Brothers, 1985) or extension (Meneilly and Storey, 1986). Several observations suggest a different process is affecting the Orocopia porphyroblasts. In most porphyroblasts the inclusion-free area occurs
at the margin of the albite, however, in some the inclusion-free zones are continuous circular zones, enclosing areas of $S_2$ graphite foliation (Plate 4A, B, D, F, G). In these cases, the $S_2$ graphite trails and relict pressure tails commonly can be traced across the zone without a change in orientation. In some cases the $S_2$ does change orientation outside the clear zone and appears to be a crenulation of $S_2$. Where $S_2$ can be traced continuously through separate areas enclosed by inclusion-free zones, the albite within one inclusion-free zone can have a different extinction angle than the rest of the albite porphyroblast (Plate 3G). In several samples, the clear zones are along grain boundaries in a triple-junction orientation.

These observations imply the areas enclosed by the inclusion-free zones were separate albite porphyroblasts at the time of $S_2$ foliation, and the inclusion free areas represent periods of very slow albite growth which precluded the incorporation of graphite. If this hypothesis is accepted, several stages of albite growth are implied by porphyroblasts with inclusion-free rings within the porphyroblast margins. This hypothesis is supported by concentric concentrations of inclusions within and parallel to the walls of some clear zones. A mechanism which could account for differential albite growth will be discussed along with the features of macroscopic folds.

Deformation Partitioning and Fold Formation

Ghosh and Ramberg (1968), Skjernaa (1975), Watkinson (1981), Cobbold and Watkinson (1981), and Watkinson and Cobbold (1981) have
demonstrated that the first fold hinges to form may add an anisotropy to a rock, strengthening it parallel to the hinges and affecting the orientation of later folds. This appears to be the case in areas of the western Orocopia Mountains. Based on the overprinting relationships seen in refolded folds in subarea 4, folds with axes that trend northeast developed before the southeast folds. Because lower quartzite layers do not show an overprinting, it is believed these quartzite layers were more resistant to folding than the micaceous ones and were unaffected by this early folding. When strain became reoriented, nearly perpendicular to the first axis, the strength of layers containing the developed fold hinges was greater than layers in which folds had not developed. Strain was then partitioned around the first fold hinges within the newly formed deformation field. Figure 1OD is a cartoon representation of this reorientation. The arrows represent deformation partitioning and show variation of the x-direction assuming a progressive simple shear. Bell and Hammond (1984) describe isotropic "pods" of rock in mylonites that posed abrupt strain transitions due to ductility differences between the block and the matrix. Models of deformation partitioning have also been applied to porphyroblast development within micaceous matrices (Bell and Rubenach, 1983; Bell, Fleming and Rubenach, 1986). As strain anastomoses around areas of shortening, folds of different orientations will develop as movement paths intersect layering. This may be the cause of sheath folds which are laterally associated with rootless folds. This plane of high shear strain anastomosing around the first fold hinges could also be
responsible for the decoupling of the large folds and the refolding of nearby folds.

The large, rootless folds within the zone of high shear strain described above are asymmetric folds with intact fold hinges. As explained above, shearing of layers oblique to the particle motion path and partitioning around existing structures is to be expected if the x-direction of shear strain were to become reoriented parallel to the fold axes. Directly below this zone, late folds have axes that trend southeast and it appears a high strain has also been partitioned around these fold hinges (Plate 4B, G); however, in this location the shear strain is oblique or perpendicular to the late fold axes. The short limbs of these folds was at a high angle to the shear surface which appears to have led to the fracture of the short limb of what were originally asymmetric folds.

Quartz veins at the base of rootless folds and the occurrence of "megaporphyroblasts" formed from groups of albite porphyroblasts in the shadow of fold hinges also support strain partitioning (Plate 4B, D). Bell and Brothers (1985), in a study of porphyroblast nucleation and growth, found that dissolution and solution transfer of material occurs in zones affected by a large component of shear strain. The highly strained areas associated with crenulation cleavage development are common locations of material dissolution. Possible sites for the precipitation of this material are low-strain areas such as pressure shadows and syntectonic veins. This is consistent with the observation of quartz veins at the base of rootless folds and in planes connecting
the limbs of sheared folds, as well as late stage growth of porphyroblasts in the "shadow" area of fold hinges.

Solution of matrix minerals also appears to be a mechanism for concentrating graphite into thinner, dense bands with progressive deformation. Hammond (1987) also noted this process in a study of dissolution and solution transfer in low-grade tectonic melange. He noted that the dissolution and recrystallization of quartz and re-equilibration of phyllosilicates produced dark anastomosing seams of extremely fine-grained phyllosilicates. He also noted that the seams tend to incorporate graphite inclusions with successive dissolution and enlarging grain size with recrystallization of muscovite.

In the Orocopia Schist, it is suggested that the finely spaced foliation of graphite in albite porphyroblasts has been segregated into the darker bands of albite and quartz, now seen as a compositional layering in the grayschist on a thin section scale.

Variability of Orientations and Models for Fold Formation

Several generations of foliations, which preceded the current matrix, are preserved within albite porphyroblasts and indicate that a process of foliation regeneration produced the fabric in the Orocopia Schist. This confirms earlier observations by Jacobson (1983a) of relict crenulations in isoclinal fold hinges and albite porphyroblasts in the Pelona Schist of the San Gabriel Mountains and correlative schists. These observations lead to speculation that the Pelona Schist underwent a process of repeated folding and development of new foliation surfaces. Refolded folds and crenulation lineations folded and oblique
to the fold axes in the Orocopia Mountains (Jacobson, 1983b and present study) also imply this process. However, several features, including open folds aligned with a stretching lineation and the variability of fold axes and lineations, indicate processes in addition to the refolding of foliation and previous folds, were important in producing the structural geometries now seen in the Orocopia Schist.

As seen in the Pelona Schist (Jacobson, 1980), the axes of both open and isoclinal folds in the Orocopia Mountains are parallel to a stretching lineation. However, in the Orocopia Mountains, the orientation of fold axes, lineations, and to a certain extent, the dominant schistosity, are variable. Variability in orientation has been observed in other localities interpreted to be subduction zones, such as the Otago Schist of New Zealand (Wood, 1978) and the rocks of Signy Island in the South Orkney Islands (Meneilly and Storey, 1986). These relationships are also observed in several shallow thrust zones and have been the source of much discussion in recent literature in an attempt to model the generation of open and polyclinal folds parallel to a stretching lineation (e.g., Coward and Kim, 1981; Sanderson, 1982).

Several features of the Orocopia Schist must be considered when trying to apply models of fold formation. First, both open and tight folds have variable orientations. This is exhibited on an outcrop scale by quartzite layers with cylindrical asymmetric folds which have axes oriented in two directions (Plate 3H) and folds of one axial orientation overlying folds with different axial orientations, some of which are rootless and not continuous into the layering (Plate 3F). Secondly,
fractured albite porphyroblasts filled with calcite and pressure shadows around porphyroblasts indicate the axis of more open folds are also parallel to the stretching direction.

A quantitative analysis of strain and subsequent interpretation of fold relationships is difficult in the Orocopia Schist due to a lack of strain indicators, a lack of overprinting of fold orientations, and the difficulty in establishing the relative timing of porphyroblast and matrix formation. Almost certainly the present foliation has developed from refolding of previous folds and foliations. Models of this process have been developed by Talbot (1979) while observing salt glaciers and Hudleston (1976) while studying ice glaciers. They determined that after nucleation of the axial surface at a high angle to layering and thrusting direction, a fold will flatten and rotate into the plane of shear, where it could again become refolded. These models, coupled with those of Bryant and Reed (1969), Esher and Watterson (1974), Quinquis and others (1978), have also been used to explain the variability of fold axis orientation through progressive rotation into the stretching direction. Although folds have probably formed in a manner similar to that described by Hudleston (1976), there is very little evidence for rotation of style 2 fold axis in the OMS (below).

In the Orocopia Mountains, most areas show little correlation between the range of fold-hinge appression and deviation of fold axes from the local stretching lineation. Most folds do show a progressive flattening of the axial surface with progressive strain, and, in some areas, it is possible to correlate the hinge angle with the orientation
of the fold axis relative to the local stretching lineation. Sheath folds seen in the schist do suggest rotation of some fold axes, and shear-type folds may have experienced some passive rotation, however, rotation models do not explain outcrops in which open folds have two orientations. Additionally, many folds are open asymmetric and chevron folds which have not developed an axial planar schistosity. These folds are commonly buckle folds which developed in nearly horizontal layers. Even asymmetric folds in which the short limb is at a high angle to layering have axes subparallel to stretching lineations.

Coward and Kim (1981) concluded that kink and chevron folds subparallel to stretching lineations on the Moine thrust could not have been rotated, but formed in this orientation when a component of shear normal to layering was present. Sanderson (1982) developed mathematical models to explain variation of strain with the differential transport of material. He showed that as boundary conditions changed by lateral or inclined ramping, the strain orientation will change. In side-wall ramps, cleavage will steepen and strike will rotate towards the ramp wall when slip is in the shear direction and normal to the shear plane. Watkinson (1975) has shown experimentally that buckle folds formed by vertical compression and lateral extension of vertically oriented planes have axes parallel to extension. Differential transport, or the movement of material at different rates or times, could rotate schistosity to vertical and, with continued deformation, produce folds with axes oriented in the direction of tectonic transport.
Coward (1984) explained a change in orientation of fold axes and stretching lineations as the result of a local rotation of thrusting direction due to pinning at the lateral tip of a thrust lobe. Differential transport of material commonly involves lobes advancing at different rates or times, and has been recognized in many shallow thrust zones but less commonly in deep thrust zones (e.g., Meneilley and Storey, 1986). A model of this type could explain many of the structural relationships seen in the Orocopia Schist.

Recent studies by Bell (e.g., 1985) have shown that albite porphyroblasts do not rotate during deformation. The orientation of graphite inclusion trails indicates that portions of a previous schistosity were steeply inclined. Additionally, on an outcrop scale, schistosity can be seen with a very steep orientation. Large-scale folds can produce a steeply dipping foliation, although the orientation will be normal to movement direction unless sheath folds are formed. The structural complexity of the Orocopia Schist makes it impossible at this time to determine which of these processes were active.

Differential transport of material is envisioned as en masse movement of "packages" of deformable material. These may move as lobes, at different rates or times, with lateral walls possibly "pinned" by slower moving adjacent material. The term "lobe" is applied only because of its application to shallow thrust zones. The term "large sheath fold" could also be applied.

This model implies a continuous process, as was proposed by Jacobson (1983b) for the formation of folds in the Pelona Schist. An
alternative suggestion is that folds formed as "generations" at specific times, (e.g., the NE oriented folds are one generation and the SE oriented folds are a later generation; or isoclinal folds are one generation and open folds are another). Several features argue against specific generations of fold formation in the OMS. (1) Specific generations of folds could not be correlated from one outcrop to another based on fold style and orientation. (2) Deformation partitioning, by its nature, implies changing strain fields and hence, the areas of deformation change with time. Changing strain with time is evident from the variable timing of porphyroblast growth. (3) Jacobson and Sorensen (1986) argue that because foliation is associated with isoclinal folds, deformation could not be due to static uplift. If deformation occurred in a subduction zone, then both subduction and non-isostatic uplift would result in changing boundary conditions in the accretionary prism with time. If folds formed under such conditions, it may be possible to determine the latest fold in any outcrop, but there is no reason to assume the latest folds from different outcrops formed at the same time.

Differential transport of materials could explain (1) the subparallel orientation of open and isoclinal folds to a variably oriented stretching lineation and (2) the consistent sense of shear seen in groups of asymmetric folds. I suggest that the western area in which the trace of the anticline axis in schistosity is lost, may represent a zone of overlapping thrust lobes.

Several important differences exist between the Orocopia Schist of
the Orocopia Mountains (OMS), the Pelona Schist of the San Gabriel Mountains, and the Orocopia Schist of the Chocolate Mountains. In the San Gabriel Mountains (Jacobson, 1980), fold orientation is less variable than in the OMS and deformation partitioning and late-stage shearing is less evident. One explanation may be that late deformation in the San Gabriel Mountains occurred at greater depths where strain tends to be more homogeneous. A difference between the OMS and the Orocopia Schist of the Chocolate Mountains is that the Hansen analysis was not effective in most areas of the OMS, although the overall asymmetry of the late folds is consistent with those of theChocolate Mountains (Haxel and Dillon, 1978).
The foliation of the Orocopia grayschist has been transposed at least twice during deformation. The earliest deformation produced quartz bands with finely foliated graphite and mica layers. Small porphyroblasts which had overgrown the early foliation developed asymmetric quartz and mica pressure shadows. Because these porphyroblasts exist only as pseudomorphs within albite porphyroblasts, their composition is not known.

Albite, idiomorphic garnet, and idiomorphic sphene porphyroblasts overgrew the S2 foliation. Epidote porphyroblasts may have been present in some locations. Macroscopic folds, of both shear and buckle type, have transposed this foliation at least once more, producing the present coarse-grained matrix and bands of graphite. The type of structures which developed depended on the orientation of the S2 surface and the competency of layers. Development of the present foliation may have been the result of a change in the direction of tectonic transport, based on the change in asymmetry of pressure shadows between S1 and S3. Tectonic transport, which was now generally in an easterly direction, produced open to isoclinal folds which are variable in orientation. Rapid albite growth may have accompanied lobe movement. Where lateral lobe surfaces overlapped, shear strain was partitioned around earlier folds, producing folds parallel to movement direction, layers with folds of different orientation, asymmetrical folds with axes in different directions, and sheath and refolded folds. During this phase of
deformation, garnet altered to chlorite in micaceous layers, was embayed in quartz-rich layers, and epidote altered to muscovite.
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CHAPTER III.
PETROLOGY AND METAMORPHISM OF THE OROCOPIA SCHIST,
OROCOPIA MOUNTAINS, SOUTHERN CALIFORNIA
The Orocopia Schist of the Orocopia Mountains (OMS) in southern California is believed to have been metamorphosed and deformed in a Late Cretaceous subduction zone. Because several tectonic models for the Orocopia and correlative schists are based on the interpretation of fold orientation, fold vergence, and additional kinematic indicators, it is important to understand the relationship of this deformation to metamorphic events. Kinematic indicators which formed during uplift may have a shear sense opposite to those formed during subduction.

Muscovite and amphibole compositions in the OMS metabasalts are similar to calcic series amphiboles found in high-pressure terranes. The OMS is metamorphosed to epidote amphibolite facies at temperatures of approximately 500°C. In the more than 2 km of estimated exposed structural thickness of schist, garnet and rutile are found at all levels and the occurrence of either hornblende or actinolite does not seem to be related to distance from the thrust, hence an inverted metamorphic zonation was not evident.

Some samples show evidence of limited recrystallization and retrograde metamorphism during the formation of late folds. Examples include garnet retrograded to chlorite in fractures oriented perpendicular to schistosity and epidote altered to muscovite which is crenulated.
INTRODUCTION

The Pelona, Orocopia, and Rand Schists (POR) of southern California and western Arizona have been tectonically placed beneath continental igneous and metamorphic rocks along a regional thrust system known as the Vincent-Orocopia-Rand-Chocolate Mountain thrust (Figure 1). The schist is believed to be a Late Cretaceous subduction complex.

Two basic types of tectonic models have been proposed to explain the location of these schists inboard of the continental margin. According to the first, the schists were deformed and metamorphosed in a low-angle easterly dipping subduction zone, now exhumed inboard of the continental margin (Yeats, 1968; Crowell, 1968, 1981; Burchfiel and Davis, 1981; Dickinson, 1981). The second type involves a westerly dipping subduction zone, responsible for the closing of a marginal basin and collision of a western block against an older continental margin (Haxel and Dillon, 1978; Ehlig, 1981; Vedder et al., 1983; Tosdal et al., 1984; Haxel et al., in press).

Structural data from the Orocopia Schist of the Chocolate Mountains and Picacho-Peter Kane Mountain area (Haxel and Dillon, 1978; Tosdal et al., in press) and the San Gabriel Mountains (Ehlig, 1981) indicate northeastward movement of the upper plate. Based on the assumption that late folds and additional kinematic indicators formed during subduction, this transport direction implies a westerly-dipping subduction zone. Structural evidence from the Orocopia Schist of the Orocopia Mountains (OMS) also indicates eastward movement of upper plate rocks (Chapter 2); however, there is some question as to whether late deformation should be
attributed to P-prograde or "down-going" subduction (Jacobson et al., in press). In the OMS, late deformation appears to be less homogeneous and less ductile than earlier deformation at any given locality (compare the isoclinal folds in Figure 11A, Chapter 2, to the asymmetrically refolded isoclinal fold with late shearing in Figure 11D, Chapter 2).

Displacement along sheared asymmetric folds is also eastward, indicating shearing was a late stage continuation of asymmetric folding. A reduction of strain homogeneity and less ductile deformation are features commonly associated with both P- and T-retrograde conditions. Additionally, mylonitic rock is absent or present only to a limited extent in the Orocopia Mountains as opposed to 1 km of upper-plate mylonite present in the San Gabriel Mountains (Jacobson, 1983a). In the Orocopia Mountains, a sharp contact between the upper and lower plates, with extensive mineralization, suggests the fault surface has been reactivated.

Recent modeling of subduction zone dynamics (e.g., Platt, 1986; Shreve and Cloos, 1986) indicates that return flow of material to higher levels within an accretionary prism is part of the subduction process. Because kinematic indicators which formed during uplift may have a shear sense opposite to those formed during subduction, it is important to understand the relationship of deformation to metamorphism. Determining whether late deformation is P-prograde or P-retrograde however is problematic in two respects. (1) The mineral assemblages present lack quantifiable pressure indicators. (2) Only limited recrystallization accompanied late folding.
The purpose of this study was two-fold: (1) to characterize the metamorphism which produced the matrix of the OMS, for comparison to other POR localities; and (2) to look for evidence which might indicate either prograde or retrograde P or T so tectonic models might be constrained. This was done through petrographic study and determination of mineral compositions by electron microprobe.
Grayschist

As seen at other POR localities (Figure 1), the most abundant lithology in the Orocopia Mountains is a graphitic albite/quartz/muscovite schist (referred to as "grayschist"), which comprises approximately 90% of the total formation and was presumably derived from a graywacke (Ehlig, 1958; Haxel et al., 1986). This lithology is characterized by numerous albite porphyroblasts within a matrix of muscovite and quartz layers. Graphite, interlayered with muscovite and included in albite, occurs in bands which range from 1 cm to many meters in thickness. These compositional layers are parallel to schistosity.

Albite porphyroblasts, 0.2-2 mm in size, are variably poikiloblastic. The most abundant inclusions are quartz or graphite, depending on whether the albite overgrew a quartz or muscovite layer. Garnet, sphene, epidote, and clinozoisite can overgrow graphite and be included by albite. Muscovite and actinolite are also found as inclusions in albite. Graphitic albite porphyroblasts, and included pseudomorph porphyroblasts defined by graphite, preserve evidence of at least two previous foliation surfaces and differential albite growth (see Chapter 2 for a complete discussion of microstructures). Mineral assemblages present in the grayschist are shown in Figure 16.

The most common mineral assemblages found in the grayschist are:

- alb + qtz + musc + epi + chl + bio + cal ± gar
- alb + qtz + musc + epi + chl + cal
- alb + qtz + musc + epi + cal + bio + gar
Figure 16. Schematic AFM projections for quartzofeldspathic OMS, projected from quartz, muscovite, albite, and epidote. AF203=Al2O3 + Fe2O3; abbreviations listed in Table 3
The frequency of mineral occurrence and definition of abbreviations are shown in Table 3.

Most minerals are oriented with their long dimensions parallel to schistosity, however, at some locations, a new axial planar schistosity is associated with tight folds and is superposed upon the earlier schistosity.

Garnet and sphene within albite porphyroblasts are idioblastic. All three minerals overgrew an S₂ foliation surface. The foliation defined by the coarser matrix minerals is interpreted to be S₃. Late folding of this schistosity caused kinking of some albite grains and crenulation of the matrix. Associated with late folds are epidote partly altered to muscovite; epidote with rims of higher birefringence than grain cores; and garnet which is embayed or rimmed and veined by chlorite, where not protected by albite.

Mafic Schist

Mafic schists comprise about 10% of the Orocopia Schist. These rocks occur as layers which range in thickness from less than a meter to tens of meters. Some layers are laterally continuous. Most samples are basaltic and similar in composition to transitional and enriched mid-ocean ridge basalts. Two samples (see Chapter 1 for discussion) have compositions similar to ocean island or seamount basalts.

In the mafic rocks, albite and epidote are porphyroblastic, with calcic amphibole, chlorite, epidote, and quartz as the most common matrix minerals. Variation in the size and abundance of the albite and epidote porphyroblasts, fine quartz lenses, and trails of sphene define
Table 3. Frequency of occurrence of minerals in different rock types of the Orocopia Schist of the Orocopia Mountains. The number of samples studied is indicated in parentheses beneath each rock-type title.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Gray Schist (78)</th>
<th>Mafic Schist (33)</th>
<th>Quartz-rich Schist (29)</th>
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<td>albite (alb)</td>
<td>94%</td>
<td>100%</td>
<td>45%</td>
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<td>quartz (qtz)</td>
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<td>100</td>
<td>100</td>
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<td>0</td>
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</table>
a compositional layering in the schist. Muscovite, biotite, sphene, calcite, magnetite, and garnet are accessory minerals and variably present in samples (Table 4). Mineral assemblages present in the grayschist are shown in Figure 14. The most commonly observed assemblages are:

\[
\begin{align*}
\text{alb + Ca-amp + chl + epi + qtz} \\
\text{alb + Ca-amp + chl + epi + qtz + musc} \\
\text{alb + Ca-amp + chl + epi + qtz + cal + gar}
\end{align*}
\]

Albite porphyroblasts are poikiloblastic and contain inclusions of quartz, garnet, epidote, Ca-amphibole, muscovite, and calcite. As seen in the grayschists, albite in the mafic schist commonly contain rims or faces adjacent to pressure shadows which are free of inclusions.

Amphiboles and micas are generally oriented with their long dimensions parallel to schistosity, although cross-cutting chlorite and muscovite are not uncommon. Schistosity is parallel to the compositional layering, except in the hinge area of folds. Some mafic bodies lack a well-developed foliation. These rocks generally also lack muscovite.

Quartzites

Ferromanganiferous metacherts are associated with the mafic schists, but are also found bounded by quartzofeldspathic schists. Quartz abundance varies between 60 and 90%. Calcite, biotite, spessartine garnet, muscovite, chlorite, hornblende or actinolite, apatite, and opaque minerals are common. Schistosity is defined by
Table 4. Mineral assemblages of analyzed metabasites from the Orocopia Schist

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<th>alb</th>
<th>qtz</th>
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<th>Ca-amp</th>
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<tr>
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<td>OR43</td>
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Notes: X=present in amounts >5%; M=present in minor amounts (<3%); tr=present in trace amounts (<1%); -=not present.
muscovite, biotite, and chlorite, and is parallel to a compositional layering defined by variation in abundance of garnet. Blue amphibole, probably riebeckite, and aegerine were found in two samples. Piemontite is present in some quartzites.

Minor Lithologies

Marble is spatially associated with quartzite. Replicate layers of marble and quartzite have thicknesses up to several meters. Isolated pods and layers of serpentine and actinolite and pods of manganese oxides are present but rare.
ANALYTICAL PROCEDURE

Amphibole, garnet, chlorite, and muscovite were analyzed at Iowa State University on an automated ARL-EMX microprobe using wavelength dispersive spectrometers. Accelerating voltage was 15 kV. For garnet, a sample current of 18 nA and beam diameter of approximately 2 microns were used. For other minerals, a sample current of 15 nA and beam diameter of about 5 to 10 microns were used in order to avoid loss of H2O, K, and Na. Inter-element and matrix effects were corrected using the Bence-Albee (1968) and Albee-Ray (1970) procedures. Standards were reanalyzed every two to four hours, depending on beam stability. Major elements were reproducible within 5 percent relative.

Standards used for amphibole were albite for Na; augite for Mg, Al, Si, and Ca; hypersthene for Fe; hornblende (Kaersutite) for Ti and K; and Mn-Ni-diopside for Mn. The standards used for chlorite were spessartine-almandine for Si, Fe, and Al; grossular for Ca; biotite for Ti; and pyrope for Mg. For garnet they were grossular for Ca; spessartine-almandine for Mn, Fe, Al, and Si; and pyrope for Mg. All analyzed minerals were first marked on probe sections, then photographed. Analyzed minerals could then be studied petrographically after analysis.

Amphibole analyses were normalized to 23 oxygens, chlorite to 28, and garnet to 12. Amphibole and chlorite analyses were recalculated for Fe$^{+3}$. Amphibole compositions were corrected by summing all cations except Na, K, Ca, and Mn to 13. Ferric iron was calculated using a total cation charge of 46$^+$. Chlorite compositions were corrected by
summing all cations to 20 and obtaining $\text{Fe}^{3+}$ from a total cation charge of $56^+$. For a complete discussion of ferric iron corrections, see Jacobson (in press) and Laird and Albee (1981a).
MINERAL COMPOSITIONS

Garnet, chlorite, and amphibole compositions from the 21 mafic schists listed in Table 4 will be discussed individually, then mutual relations will be discussed. Aspects of muscovite compositions for both grayschists and mafic schists are presented for background and later discussion.

Muscovite

Muscovite in grayschists is moderately celadonitic, with Si between 3.3-3.5 atoms per 11 oxygens, calculated assuming all iron as FeO. In mafic schist, Si is lower, ranging between 3.2-3.3. Paragonite substitution is low, with Na between 0.0-0.1 pfu in grayschists and 0.0-0.2 pfu in metabasites.

Garnet

Garnet in the mafic samples is rich in FeO and CaO (Figure 17). Individual grains are zoned and generally have higher Mn in the cores. The proportions of Mg ranges from 0.1-0.2 pfu, Fe from 1.3-2.4 and Ca from 0.5-1.2. Mean Ca values for individual samples range from 0.7-1.0. This variation may be controlled in part by bulk composition.

Chlorite

Chlorite is aluminous, with Al between 4.5-5.3 pfu. The ratio Mg/Fe ranges between 0.4 and 0.6.
Figure 17. Mn - Mg+Fe$^{2+}$ - Ca diagram for garnet in mafic Orocopia Schist
Amphibole

Because the transition from actinolite to hornblende represents the facies change from greenschist to epidote amphibolite, amphibole compositions can be an indication of temperature change (Laird and Albee, 1981a). The range in composition of matrix amphiboles is shown in Figures 18-20.

Figure 18 shows the extent of the glaucophane substitution:
$$\text{Na}^4\text{Al}^6\text{Fe}^{3+} + \text{Ti} \rightleftharpoons \text{CaFe}^{2+} + \text{Mg} + \text{Mn}.$$  

The data indicate a glaucophanic substitution is present. Figure 19 shows $\text{Al}^6 + \text{Fe}^{3+} + \text{Ti}$ versus four-fold aluminum. Because these are the principal cations in the tschermak substitution:
$$\text{Al}^6 \text{Fe}^{3+} + \text{Ti} \rightleftharpoons \text{Fe}^{2+} + \text{Mg} + \text{Mn})\text{Si},$$

variation along the diagonal represents the transition from actinolite to hornblende. Amphibole analyses plot above the diagonal in this figure which indicates that glaucophanic substitution exceeds edenitic substitution. The axes of Figure 20 represent the reactants of the edenite substitution:
$$\text{Na} + \text{K} \text{Al}^4 \rightleftharpoons \text{Si}.$$  

The obvious feature seen in these diagrams is the presence of one distinct and continuous compositional trend or amphibole series. A more sodic series (actinolite to winchite and crossite) is not seen in the Orocopia amphiboles as is seen in the Rand Mountains and other localities inferred to have a subduction origin (Jacobson and Sorensen, 1986). Similar to the calcic series of the Catalina Schist, amphiboles in the Orocopia mafic schists show an increase of $\text{Na}^4$ from actinolite to
Figure 18. Na\textsuperscript{IV} vs. Al\textsuperscript{VI} + Fe\textsuperscript{3+} + Ti for amphibole from the OMS mafic schists
Figure 19. $\text{Al}^{\text{VI}} + \text{Fe}^{3+} \text{Ti}$ vs. $\text{Al}^{\text{IV}}$ for amphibole from the OMS mafic schists
Figure 20. $\text{Na}^+ + \text{K}$ vs. $\text{Al}^{IV}$ for amphibole from the OMS mafic schists
hornblende. The OMS amphiboles have less variation in Na\(^{M4}\) and Al than seen in the Rand amphiboles. The OMS amphiboles are also similar to those from metabasites of the Pelona Schist in the San Gabriel Mountains (Jacobson, unpublished data). When considering this comparison it should be noted that the analyses were collected from different microprobes, and a different Fe\(^{+3}\) correction was applied to the Rand amphiboles (that of Papike et al., 1974).

Garnet, chlorite, and amphibole compositions for the 21 mafic samples are plotted on an albite-epidote projection (Harte and Graham, 1975; Laird, 1980) in Figure 21-23. Each point represents the average of about nine analyses, and compositions for minerals in a single sample are connected by tie lines. Samples have been separated into two groups in Figures 22 and 23, so that Figure 22 shows chlorite-amphibole tie lines that are generally not overlapping, and Figure 23 shows the most discordant tie lines. As can be seen from Figure 22, in most samples, compositional tie lines are subparallel, implying that coexisting minerals are near equilibrium. These will be referred to as "group 1". Amphibole and chlorite coexisting with garnet are the most Fe-rich.

Most of the discordant tie lines have slopes less negative than group 1 tie-line slopes. This may indicate either that amphibole in these samples has lower Fe/Mg or that chlorite has higher Fe/Mg than in group 1 samples.
Figure 21. Epidote-albite projection (Laird, 1980) for mafic rocks from the Orocopia Mountains. $AF_{203} = Al_2O_3 + Fe_2O_3$; $FMO = FeO + MgO + MnO$
Figure 22. Epidote-albite projection of concordant "group 1" samples. Axes as in Figure 21.
Figure 23. Epidote-albite projection of discordant tie-line samples. Axes as in Figure 21.
Amphibole compositions in the metabasalts of the Orocopia Mountains are similar to calcic series amphiboles found in high-pressure terrains, e.g., the Catalina Schist and Rand Schist (Jacobson and Sorensen, 1986) and the Sanbagawa Metamorphic Belt and Franciscan Complex (Laird and Albee, 1981b). Sodic amphibole, indicative of high pressures, is present in the Rand Schist; however, sodic amphibole (other than riebeckite in metachert and barroisite in the San Gabriel Mountains) was not found in either the Orocopia Mountains or the San Gabriel Mountains, although relict crossite has been reported in the Sierra Pelona (Graham, 1975; Graham and Powell, 1984). Sodic amphibole in the OMS metachert is riebeckitic. Riebeckite is stable over a wide pressure range and is not indicative of high pressures. Additionally, epidote is not present, which further limits the use of riebeckite as an indicator of metamorphic grade. Although sodic-series amphibole was not observed in the Orocopia metabasalts, a subduction origin for the metamorphism of the Orocopia Schist is suggested by: the MORB-like nature of the mafic schists in the Orocopia Mountains (Chapter 1); the oceanic nature of the other lithologies; the similarity of calcic series amphibole to that in other high-pressure terranes; the high Si muscovites as found in other high-pressure terranes such as the Franciscan Complex and Sambagawa Metamorphic Belt (Ernst, 1963; Guidotti and Sassi, 1976; and others); the occurrence of proximate granitic plutons; and pressure estimates from the correlative schist of the Sierra Pelona (8-9 Kb, phengite and jadeite content of clinopyroxene: Graham and Powell, 1984).
Several aspects of the metamorphism in the OMS differ from that of other POR localities. These relate to metamorphic zonation, discordant tie-lines and retrograded mineral assemblages. In other POR localities, the metamorphic grade of the schist decreases downward from the contact with the upper plate, presumably due to underplating beneath a hot hanging wall (Peacock, in press). In the 4 km structural thickness of schist exposed in the San Gabriel Mountains, Ehlig (1958) found that grain size increased toward the fault. He also found that metamorphic grade in the Sierra Pelona changed from greenschist to lower amphibolite grade 500 meters beneath a one kilometer zone of mylonites. Graham and Powell (1984) estimated, using a hornblende-garnet geothermometer, that temperatures in the Sierra Pelona ranged from 480°C at the garnet isograd to between 620°C and 650°C adjacent to the upper plate. An inverted zonation has also been described for the Chocolate Mountains by Haxel and Dillon (1978), the San Gabriel Mountains by Jacobson (1980, 1983b), and the Rand Mountains by Postlethwaite and Jacobson (1987).

In the OMS mafic schist, an inverted metamorphic zonation was not found. In the more than 2 km of estimated exposed schist, garnet and rutile are present at all structural levels. The occurrence of either hornblende or actinolite does not seem to be related to distance from the thrust. Temperatures of the schist also appear uniform. Using the garnet-hornblende geothermometer (Graham and Powell, 1984), the composition of samples 29, 280A, and OR106 yielded temperatures of 495°C, 508°C, and 512°C respectively. Although it is difficult to follow the trace of the fault in the vicinity of sample OR106, this
sample is within at least 1 km of the thrust contact and is probably only 200-400 m structurally below the upper-plate. No pronounced increase of grain size was noted in the schist very close to the thrust, although no detailed petrographic studies were made for samples from that area.

In the QMS there is some textural evidence of retrograde metamorphism during the formation of style 2 folds. Examples include matrix garnet, generally idioblastic in albite porphyroblasts, retrograded to chlorite in fractures oriented perpendicular to schistosity; epidote altered to muscovite which is crenulated.

There are several possibilities for discordant tie lines: (1) The chlorite or amphibole compositions reflect recrystallization under conditions different than those of group 1. (2) Bulk composition plays some undetermined role. (3) Diffusion related to deformational stress gradients has modified mineral compositions. (4) The microprobe beam was overlapping grains of different compositions. Most of these samples contain either one or more of the following features: epidote with strongly retrograded rims; strained and undulatory albite; cross-cutting muscovite; bimodal grain size of quartz or chlorite; matrix grains oriented differently than inclusion orientations within albite porphyroblasts. These features suggest, but are not conclusive, that discordancy of tie lines is due to disequilibrium resulting from deformation or recrystallization.

In an attempt to quantify the physical conditions of matrix formation, Na content of amphibole was examined. Brown (1974, 1977) and
Trzcienski et al. (1984) believe that magnetite will buffer the reaction relation of NaM4 in calcic amphibole if coexisting with albite and chlorite in the reaction:

\[ \text{Ca-amphibole + iron oxide + albite + chlorite + H}_2\text{O (+ stilp, qtz)} \]

\[ \text{Crossite + epidote (± muscovite, qtz).} \]

This is based on a correlation between crossite component (NaM4) in Ca-amphibole and pressure of metamorphism. Experimentally, the relationship between glaucophanic activity or a "crossite component" and pressure has not been completely resolved. Additionally, the role of coupled magnetite/ilmenite buffering of oxygen fugacity in high-pressure metamorphism is still uncertain. Although the relation of pressure to amphibole composition has many uncertainties, this methodology has been applied to high pressure belts similar to the POR (e.g., Hosotani and Banno, 1986 study of sodic amphibole in the Sanbagawa Metamorphic Belt).

Of the 21 samples listed in Table 4, four contain magnetite. The approach of Holland and Richardson (1979) was taken to determine the "glaucophanic component" of calcic amphibole cores and rims in the OMS, based on assumptions of amphibole structure. Cations were assigned as follows: tetrahedral sites = Si then Al; M2 = remaining Al, Ti, Fe\(^{3+}\), then Fe\(^{2+}\) and Mg in proportion to make 2; M1-3 + remaining Fe\(^{2+}\) and Mg; M4 = excess Mn, Ca, then Na; A = excess Na and K.

The activities of glaucophane, tremolite, and edenite were calculated based on molar proportions of specific cations in specific amphibole sites and assuming ideal mixing.

The results were generally inconclusive, with similar glaucophanic
activities in both rim and core for three samples. A fourth sample (OR57A) has lower glaucophanic activity in the rim than the core.
CONCLUSIONS

The Orocopia Schist of the Orocopia Mountains has mineral assemblages and amphibole compositions which are similar to those from metabasites of the Pelona Schist of the San Gabriel Mountains and the Catalina Schist and are generally similar to those of the Rand Mountains, except with less variability of Al. Mineral assemblages and compositions resemble those of other high P/T metamorphic terranes (e.g., the Sanbagawa metamorphic belt of Japan). Limited recrystallization accompanied late deformation during which schistosity was folded. Several petrographic and metamorphic features suggest caution in adopting the assumption that late deformation occurred during "down-going" subduction. These include: the occurrence of crenulated muscovite overgrowing epidote, garnet retrograded to chlorite in fractures perpendicular to schistosity, and discordant tie lines in samples showing late stage deformation. The scarcity of mylonites and lack of evidence for an inverted zonation also support this conclusion.
REFERENCES


GENERAL SUMMARY

The Orocopia thrust of southern California is a segment of a regional system termed the Vincent-Chocolate Mountains thrust, which has been interpreted as a Late Cretaceous subduction zone. The Orocopia Schist of the lower plate is thought to represent an accretionary wedge. Element abundances indicate the mafic schists of the Orocopia Schist of the Orocopia Mountains (OMS) are tholeiitic and alkalic basalts generated from several sources. Most samples resemble normal to enriched mid-ocean ridge basalts (N-MORB to E-MORB). Variation of MORB-like samples can be accounted for by low-pressure crystal fractionation. A second group of samples resembles non-depleted basalts (P-type) and may have formed as an ocean island or seamount. A third group has elemental abundances similar to E-MORB, however, Y, Yb, Lu, and P are more depleted. Muscovite and amphibole compositions in the OMS metabasalts are similar to calcic series amphiboles found in high-pressure terranes. The OMS is metamorphosed to epidote amphibolite facies at temperatures of approximately 500°C. An inverted metamorphic zonation was not evident. The earliest deformational fabric of the OMS is preserved as a foliation of graphite in pseudomorphed porphyroblasts. This foliation (S1) has been transposed at least twice during deformation to produce the current matrix. Early foliations and relict crenulations in the matrix indicate that a process of foliation regeneration produced the fabric in the OMS. The youngest structures are open to tight folds ("style 2 folds") which fold schistosity and compositional layering. Style 2 folds, in some cases, overprint
isoclinal folds ("style 1 folds"). Both styles of folds, as well as stretching lineations, have variable orientations. Some samples show evidence of limited recrystallization and retrograde metamorphism during the formation of late folds. Examples include garnet retrograded to chlorite in fractures oriented perpendicular to schistosity and epidote altered to muscovite which is crenulated. Evidence from the Orocopia Mountains indicates that movement direction, in some areas, was oblique to an earlier shear strain and, at the time of style 2 folding, was generally eastward. However, the lack of mylonite and extensive mineralization at the fault contact suggest reactivation of the Orocopia thrust.
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