The structural geology of the Red Cloud thrust system, southern Eastern Transverse Ranges, California

Clay Edward Postlethwaite

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The structural geology of the Red Cloud thrust system, southern Eastern Transverse Ranges, California

Postlethwaite, Clay Edward, Ph.D.
Iowa State University, 1988
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The structural geology of the Red Cloud thrust system, southern Eastern Transverse Ranges, California

by

Clay Edward Postlethwaite

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

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Iowa State University
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INTRODUCTION

The Eastern Transverse Ranges are located in south-central California, east of the San Andreas fault. Unlike many of the other ranges in California, the Eastern Transverse Ranges are bounded by east-west striking, left-lateral strike-slip faults (Figure 1). The north-south borders of Eastern Transverse Ranges are the Pinto Mountains fault and the Aztec Mines Wash/Salton Wash fault, respectively. The eastern boundary is the northwest-striking Sheep Hole Mountains fault.

Approximately half of the Eastern Transverse Ranges are underlain by Jurassic and Cretaceous plutons. Powell (1981) found that these rocks are exposed in two overlapping belts. Jurassic plutons are most common in the northern and eastern portions of the Eastern Transverse Ranges where they define a northwest trending belt. The Cretaceous plutons are common throughout the rest of the Eastern Transverse Ranges.

Powell (1981) made the first comprehensive study of the crystalline rocks of the Eastern Transverse Ranges. He described a regional thrust system, called the Red Cloud thrust, separating two basement complexes. The generalized structure of the terranes prior to thrusting is shown in Figure 2, which is from Powell (1982). Powell called the hanging wall of the thrust the San Gabriel terrane, because many of its units had previously been described in the San Gabriel Mountains in the Western Transverse Ranges. The San Gabriel terrane is a complex of orthogneisses cross-cutting a volumetrically minor amount of paragneisses. The terrane is ultimately intruded by an undeformed 1.2
Figure 1. Simplified geologic map of the crystalline rocks of the Eastern Transverse Ranges. After Powell (1981). CM — Chuckwalla Mountains, CWM — Cottonwood Mountains, EM — Eagle Mountains, EOM — eastern Orocopia Mountains, HM — Hexie Mountains, LCM — Little Chuckwalla Mountains, LSBM — Little San Bernardino Mountains, PM — Pinto Mountains, WOM — western Orocopia Mountains, bw — Big Wash and cw — Cholla Wash
29 Palms
Meso. Intrusive
Desert Cenier
Saltan Wash/Mines Wash fault
Figure 2. Schematic block diagrams of the pre-thrusting San Gabriel and Joshua Tree terranes (from Powell, 1982). S1a - paragneiss, S1b - Soledad augen gneiss, S2 - Augustine gneiss, S3 - Syenite-mangerite-jotunite, J1 - Joshua Tree granite gneiss, J2 - Pinto quartzite, and J3a-c - metasediments of the Joshua Tree terrane
Ga syenite-mangerite-jotunite suite. The older units have been metamorphosed to upper amphibolite facies and apparently multiply deformed. Powell concluded that the San Gabriel terrane represents a mid-level section through the crust.

The footwall of the Red Cloud thrust was termed the Joshua Tree terrane because of its extensive exposures in Joshua Tree National Monument. The Joshua Tree terrane is a complex of paragneisses and quartzites in depositional contact upon a granite/granite gneiss. The lower portion of the metasedimentary section is quartzite rich. The upper portion is more pelitic.

The Red Cloud thrust, actually a system of thrusts, is thought to have had a complicated history (Figure 3; from Powell, 1981). The sequence of events that Powell proposed to explain the present thrust geometries is as follows:

1) Early thrusting of unknown direction, but presumably west directed.

2) Folding on a regional scale of the Red Cloud thrust surface.

3) Renewed thrusting. Powell illustrated three different geometries consistent with the field relationships. He favored the one in cross section 3c. An important implication of this interpretation is that some thrust surfaces are the Early Red Cloud thrust surface, some are only the Late Red Cloud thrust, and some surfaces are a combination of the two.
Figure 3. Cross sections showing the evolution of the Red Cloud thrust system and the relationship between the Early and Late Red Cloud thrusts (from Powell, 1981)
4) Continued folding into a west-vergent antiform-synform pair.

The purpose of this study was to test Powell's model through detailed structural analysis of the Red Cloud thrust system. The westward sense of thrusting on the Late Red Cloud thrust is constrained by offset metamorphic isograds. Powell (1981) suggested an offset of approximately 30 km. The sense of motion on the Early surface is not so constrained, but is inferred to be west-directed. By Powell's model there are no units in common to both plates, which implies a minimum offset of 80 km based on the width of overlap of the hanging-wall and foot-wall blocks. Therefore, it is this thrust surface and its associated deformations that are the primary focus of this study.

Tentative radiometric dating since Powell's dissertation has suggested that the Red Cloud thrust deformation was early Mesozoic in age (Powell, 1986; K. Howard, U.S.G.S., Menlo Park, CA, personal communication, 1988). The tectonic history of southern California at this time is poorly understood. There is some evidence that a nascent continental arc developed in the Triassic, apparently just after a truncation of the continental margin (Burchfiel and Davis, 1981). A well defined continental arc, however, did not form until Jurassic time.

Recently Powell (1986) has amended his original model and attempted to place the Red Cloud thrust event into a regional framework. Powell believes that the Transverse Ranges underwent a regional 50 km by 300 km antiformal uplift in the Late Cretaceous which raised the crystalline
rocks of this study to shallow crustal levels. He argued that a similar uplift occurred in the early Mesozoic, along with episodic decollement along the roofs of some plutonic units.

This study involved detailed mapping of selected areas critical to Powell's interpretation. Particular attention was paid to fold development and to any kinematic indicators that might allow interpretation of the sense of movement and the conditions of deformation. Samples were collected for microscopic examination and selected chemical analyses. Mylonitic rocks were studied for shear indicators and quartz-rich rocks were used for c-axis quartz fabric analysis.

The results are presented in four chapters, the first three of which describe individual geographic areas studied in detail (for locations see Figure 1). The first deals with a segment of the Early Red Cloud thrust is exposed on the limbs of a south plunging synform in the southeast Eagle Mountains. There is evidence of reactivation of a portion of this surface.

The second area of interest is in the western Hexie-southern Little San Bernardino Mountains, where a mylonite zone is present at the base of a section of mylonitic gneisses. Powell's model predicts the mylonite zone here to be the combined Early and Late thrusts.

The third chapter describes the western Chuckwalla Mountains and the southern Eagle Mountains. Here, segments of the Early and Late Red Cloud thrusts are exposed, as well as a migmatite complex that was involved in syn-magmatic deformation.
A final section deals with a set of north-northwest trending mylonite zones. These zones are relatively narrow, right-lateral shears that are found throughout the Eastern Transverse Ranges. They postdate the Red Cloud thrust deformation and play an as yet uncertain role in the present distribution of lithologies in the Eastern Transverse Ranges.
BIG WASH, SE EAGLE MOUNTAINS

Introduction

Powell (1981) mapped a segment of the Red Cloud thrust on both limbs of a south-plunging synform in the southeastern Eagle Mountains. Here, the thrust places sillimanite-grade orthogneisses and plutonic rocks upon andalusite- and sillimanite-bearing metasediments (Figure 4).

The metasediments are composed of two lithosomes (Powell, 1981) (Figure 5). Greater than 90 percent of the sequence is the Pinto quartzite. Two textures were recognized in the field: a bluish-gray to tan quartzite with rare relict quartz-pebble conglomerate lenses and a coarse-grained white, vitreous quartzite. The Pinto quartzite is interlayered with the Black Eagle schist. It is a gray to greenish weathering quartz-aluminosilicate-muscovite schist. The white quartzite everywhere occurs between it and the bluish-gray to tan quartzites. This, and the absence of isoclinal folds in the metasediments has led Powell to interpret the sequence as primary, not the result of structural interlayering.

The Pinto quartzite overlies the Joshua Tree granite gneiss. Separating the two is an aluminous schist with quartz augen that Powell interpreted as a paleoweathered zone, indicating that the protolith of the quartzites was deposited upon the protolith of the granite gneiss. L. T. Silver (personal communication in Powell (1981)) reported a 1650 Ma age for the protolith of the Joshua Tree granite gneiss.
Figure 4. Aerial Photo of Big Wash area with geologic contacts
Image I.D. LBSS000130234 frame 235. Scale = 1:33,000 (approximate). Jgd = Jurassic (?) gabbro-diorite, SMJ = syenite-mangerite-jotunite suite, gg = granitic gneiss, Pq = Pinto quartzite, JTgg = Joshua Tree granite gneiss
Figure 5. Stratigraphic section measured through metasediments above the Joshua Tree granite gneiss in western Big Wash. Section measured in meters.
A syenite-mangerite-jotunite (SMJ) suite and lesser amounts of banded granitic gneiss form the hanging wall of the Red Cloud thrust. The SMJ suite is a complex of pyroxene-bearing igneous rocks of dioritic to syenitic composition. L. T. Silver (1971; personal communication reported in Powell, 1981) reported a U/Pb lead age of 1200 Ma on correlative rocks in the Western Transverse Ranges.

Eastern Limb of the Big Wash Synform

The footwall of the Red Cloud thrust on the east limb of the synform is composed of a thickness of approximately 100 m of strongly fractured quartzite and metaconglomerate resting upon Joshua Tree granite gneiss. Very little of the actual fault contact is exposed in lower Big Wash (Figure 4). An elongate gabbroic body intruded along, or adjacent to, the northern portion of the fault trace (Figures 4, 6a). This intrusive is inferred to be one of the regionally extensive plutons of Jurassic age by Powell. The fault lies beneath alluvium near Big Wash. The best exposure on the east limb is where the thrust cuts across the toe of a quartzite spur southeast of the pluton. Here, the contact between the banded granitic gneiss and the Pinto quartzite is quite sharp. There is little evidence of ductile shear along the contact. There are no clear textural differences in the appearances of hand-specimens of quartzite from directly beneath or farther from the thrust.

Small, decimeter-scale open to tight folds are present in the quartzites. They are generally west-vergent. No isoclinal folds were recognized. Strained metaconglomerate clasts have aspect ratios of
Figure 6a. View looking southwest along the Red Cloud thrust in lower Big Wash. Jgd - Jurassic gabbro-diorite, gg - granitic gneiss, Pq - Pinto quartzite. The Joshua Tree granite gneiss-Pinto quartzite contact is on the ridge in the near distance. Desert Center is partially visible in the Chuckwalla Valley in the far distance.

Figure 6b. Exaggerated grain-growth microtextures from the Pinto quartzite. All areas marked "A" are portions of the same grain that is presumed continuous outside the the plane of the thin section. Horizontal field of view - 4.8 mm.

Figure 6c. Sawtooth grain boundaries in the Pinto Quartzite. Horizontal field of view is 2.8 mm.

Figure 6d. Well developed subgrains in quartz grain from Pinto quartzite. Horizontal field of view is 2.8 mm.
19:6:1 indicating a flattening finite strain ($k=0.38$). The long axes of the pebbles parallel the stretching lineation in the quartzites and overlying granitic gneisses.

These quartzites possess an unusual microtexture. Most samples are composed of large (>1mm), interpenetrating grains (Figure 6b). Individual grains have very irregular habits, so much so that non-adjacent grains with the same optical orientation are believed to be portions of the same grain continuous outside the plane of the thin section. Minor phases in the quartzites (muscovite and sillimanite) have little effect on the grain shapes of the quartz crystals. In most cases they are completely overgrown.

This sort of microtexture is reported to be the product of grain boundary migration (Wilson, 1973; Bouchez and Pecher, 1981; Schmid and Casey, 1986). It is most common in quartzites that have been deformed or annealed at a high temperature. "Sawtooth patterns" are seen along some grain boundaries (Figure 6c). These apparently represent portions of a migrating grain boundary that have been pinned or impeded by dislocations or inclusions.

Individual quartz grains have complicated internal structures. Most grains show strong subgrain development and undulatory extinction (Figure 6d). Similar textures are described from Mount Isa, Australia by Wilson (1973, Figure 13, p. 57) for samples that had undergone a later, lower temperature deformation.
Western Thrust Contact

On the west limb of the synform, the shear zone places the SMJ suite directly upon the schist-quartzite sequence. At the field trip stop 5 of Powell (1982), the shear zone is a fine-grained chloritic mylonite with layers of brecciated quartzite (Figure 7a). Only the lowest few decimeters of the SMJ suite is strongly affected by the shear zone, although a faint lineation is present in these rocks throughout the lowest hundred meters. The uppermost quartzites and schists have a mylonitic aspect (Figure 5). The contact is intruded by a mafic dike for part of its outcrop length. The mafic dike has a crude foliation (Figure 7b) along the shear zone indicating that some movement occurred during or after the dike's emplacement. Samples taken in and near the shear zone are strongly chloritized and extremely fine grained. Much of the chloritic material along the contact is apparently derived from the mafic dikes.

The north-northwest trending dikes visible in the syenite-mangerite suite on Figure 4 are members of a quartz latite dike swarm. These are very common in the syenite-mangerite suite, but less so in the quartzites. The quartz latites are older than the mafic dikes previously described and are truncated by the fault separating the quartzites from the SMJ suite (Figure 7c).

There are some significant differences between the mylonitized dike rock and the mylonitic schists and quartzites at the top of the
Figure 7a. Chloritic quartzite-clast breccia from beneath the syenite-mangerite-jotunite suite in upper Big Wash

Figure 7b. Chloritic mylonite derived in part from mafic dikes beneath syenite-mangerite suite in upper Big Wash

Figure 7c. View looking southeast at the reactivated Red Cloud thrust near its termination against a high-angle fault in upper Big Wash. Jql = quartz latite dikes, SMJ = syenite-mangerite-jotunite suite, Pq = Pinto quartzite

Figure 7d. East-vergent shear bands in mylonitic schist along the reactivated Red Cloud thrust
footwall. First, the schists and quartzites are cut by a variety of dikes. Second, the schists have undergone recovery processes that have allowed quartz to develop blocky grain boundaries. No ribbon grains were noted. Grain boundaries indicate grain boundary migration was active to a minor degree. Third, along the limited exposure of footwall mylonite, the lineation trends due east (Figure 8). Although the average lineation trend in the underlying quartzites is about N64E, the lineation becomes more easterly in the upper 80 m of the section. It is not clear if this is a result of overprinting of the northeast lineation by an easterly one or simply a gradual variation in lineation orientation across the section. Northeasterly trending lineations in the SMJ suite above the mylonites suggests an overprinting of the northeast-trending lineation by the east-west lineation, but there are too few measurements to be certain. Slight differences are also evident in the foliation attitudes. The mylonites have a more northeasterly strike than the underlying quartzites.

A variety of mesoscopic kinematic indicators are present within the mylonitized schists at the SMJ/Pinto quartzite contact. Small, asymmetric folds, shear bands (7d), and foliation "fish" similar to those described by Hanmer (1982) (Figure 9a) all indicate a top-to-the-east sense of shear.

The footwall

From the Joshua Tree granite gneiss at the base to the thrust, over 350 m of metasediments is exposed in upper Big Wash. Powell (1981,
Figure 8. Structural data for the Pinto quartzite and mylonitic schist in western Big Wash. + = quartzite foliation poles. X = quartzite lineations. △ = mylonite lineations
Figure 9a. A lens with an oblique foliation in the mylonitic schist (foliation fish of Hanmer (1982))

Figure 9b. Relict conglomerate lenses in western Big Wash. Viewed approximately down lineation

Figure 9c. Undeformed conglomerates in the Pinto Mountains

Figure 9d. Possible relict sand grain near the base of the Pinto quartzite in western Big Wash. Horizontal field of view is 2.8 mm

Figure 9e. Boudinaged sillimanite pods in the upper Pinto quartzite, western Big Wash. Horizontal field of view is 4.8 mm

Figure 9f. Partially recrystallized quartzite (sample BW 34). Horizontal field of view is 2.8 mm
1982) interpreted an aluminous schist along the gneiss-quartzite contact as a metamorphosed soil horizon, and that the protolith of the quartzite was deposited directly upon the gneiss protolith.

The lower portion of the quartzite section contains numerous conglomerate lenses (Figure 9b). Correlable undeformed quartz-pebble conglomerates occur in the Pinto Mountains (Figure 9c). The pebbles are composed predominately of white vein quartz and black quartz rich in hematite (?) in a nearly pure quartz matrix (Powell, 1982; fieldtrip stop 11). No clasts of the underlying gneiss were found in the quartzite. Outcrop measurements of clast shapes indicate a constrictive total finite strain (Figure 10), in marked contrast to the clast shapes measured on the eastern limb of the synform. The long axes of the pebbles are parallel to the stretching lineation. Although the shape measurements were made over 300 m from the thrust contact at the top of the quartzite section, they record relatively high strains (average axial ratios=13.5:2.5:1, k=2.9). Evidence of similar, though finer grained conglomerate lenses was found higher in the section, where small black clasts were recognized, but shape analysis was not possible because of the high strains, smaller grain size, and recrystallization of the quartzites.

The western quartzites are relatively coarse grained, although not as coarse as those on the east limb. Average grains sizes for the west-limb quartzites range from around 0.4 mm at the base of the section to 0.7 mm near the top. The quartzites higher in the section are purer. This may account for larger grain size. Unlike the quartzites from the
Figure 10. Flinn (1962) diagram of pebble shapes from various exposures of the metaconglomerates in the Pinto quartzite.
east limb, the grain shapes and sizes are strongly controlled by pods of fibrous sillimanite. Figure 9d is from a sample taken from the lower part of the section. It displays an unusual texture in which the foliation wraps around "knots" of pure quartz, with pressure shadow-like growths of sillimanite at the ends of the knots. Bouchez and Pecher (1981) described a texturally similar structure from upper-greenschist facies quartzites beneath the Main Central thrust of the Himalayas. They interpreted the knots as being relict after individual sand grains. These textures are not present higher in the section. Rather, there the sillimanite pods are more evenly distributed and show progressively better developed boudinage (Figure 9e) toward the top of the section. This indicates that the sillimanite grew before or during the ductile deformation of these quartzites and implies that the finite strain since the formation of the sillimanite increases up-section.

C-axis Quartz fabrics

Schmid and Casey (1986) argued that the dominant process occurring in polycrystalline quartz during ductile simple shear deformation by intracrystalline glide is rotation of the <a> glide direction of the active slip system into parallelism with the shear direction. Activity of the basal <a> glide causes the high population densities of c-axes at a high angle to the foliation. Maxima at moderate angles to the foliation indicate activity on the rhombohedral glides, and c-axis maxima in the foliation, perpendicular to the shear direction, indicate activity of the prism <a> glide systems.
The Pinto quartzites are >90 quartz with minor sillimanite and a dusting of a fine opaque mineral (hematite?). Other phases are andalusite, muscovite, plagioclase, biotite, and zircon. C-axis quartz fabric analysis was conducted in order to try and determine the shear sense during the Red Cloud thrust deformation.

**Eastern Limb of the Big Wash Synform**

Figure 11 shows contoured, equal-area, lower-hemisphere projections of optically measured quartz c-axes. In each case, the foliation of the sample is horizontal and the lineation lies in the page with the westerly end on the left.

Samples BW 45 and BW 45b were taken directly beneath the thrust contact. Their fabrics are classified as incomplete type I cross girdles of Lister (1977), although sample BW 45 has very weak girdle development. Samples BW 44 and BW 46 were collected further from the thrust. Sample BW 44 has a single girdle fabric. BW 46 has an elongate maximum oblique to the foliation with no clear relation to the finite strain reference. Although all four samples have microtextures typical of rocks that have undergone exaggerated grain growth, a high temperature (amphibolite grade) phenomenon, the c-axis fabrics of BW 45 and BW45b are similar to those found in greenschist facies rocks. Based on the analysis of Bouchez and Pecher (1981) and Schmid and Casey (1986), three samples (BW 46 excluded) show evidence of strong activity of the basal \(<a>\) glide system, a characteristic of relatively low-temperature deformations (Tullis *et al.*, 1973). Quartz-rich tectonites
Figure 11. Contoured c-axis quartz fabrics from the Pinto quartzites eastern Big Wash. The foliation of the sample is shown bisecting the stereonet. The lineation of each sample is oriented such that the west end intersects the net at the left perimeter and east end at the right. Approximately 200 grains were measured in each sample. Contouring is by the Kamb method.
deformed under amphibolite facies conditions typically display elongate point maxima parallel to the Y (intermediate) strain axis (Allison et al., 1978; Schmid and Casey, 1986).

Sample BW 34 was taken from a quartzite near the thrust at the nose of the synform. The coarse grains in this sample are partially recrystallized (Figure 9f). The c-axis fabric diagram for BW 34 in Figure 11 includes measurements of the large grains only. These grains contain prismatic deformation bands.

The c-axis fabric for large grains suggests fabric development during west-vergent shear. The incompleteness of the girdle may imply that the sample underwent a nonplane strain or that recrystallization has selectively removed portions of the fabric.

Based on a study of quartz ribbons in a high-grade granite gneiss, Culshaw and Fyson (1984) showed that grain-boundary migration can alter c-axis fabrics. In their case, the larger grains defined a better girdle than the smaller grains, indicating that the grain boundary migration actually enhanced the fabric. This suggests that the larger grains are better oriented for slip. Wilson (1973) and Bouchez and Pecher (1981) found that this process caused a degeneration of the crystallographic fabric.

The small grains have an essentially random fabric (Figure 12). They have polygonal habits which suggests that they are the product of a post-deformational event. The host grains do exert some control over the orientation of the new grains. Figure 13 is a histogram of the c-axis host/subgrain angle for sample BW 34 (cf. Ransom, 1971). The
curve for a random population is included on the plot. As found by Ransom (1971) and Wilson (1973), there are fewer new grains at angles of 60° or greater from the host orientation than would be expected for a random population.

Western Limb of the Big Wash Synform

Quartzites from the western limb of the Big Wash synform were also used in c-axis fabric analysis. Their c-axis fabrics are shown in Figure 14, along with their height in the stratigraphic column in Figure 5.

There is a general increase in fabric intensity higher in the section to sample BW 10. Sample BW 3 is from the base of the section. It has a very low intensity fabric that has no relationship to the foliation and lineation.

Sample BW 7 has an asymmetrically filled single girdle inclined at about 45° counterclockwise to the foliation. The maxima at the perimeter and center indicate that the basal and prism glides were active.

BW 8 has a moderately well developed girdle inclined eastward. The two maxima close to the foliation suggest that the rhombs were the dominant glide system. From the inclination of the central portion of girdle, this fabric suggests counterclockwise shear (cf. Lister and Williams, 1979) and Behrmann and Platt, 1982). The orientation of the whole girdle implies clockwise shear.
Figure 12. C-axis fabrics of the large (old) and small (new) grains in sample BW 34
Figure 13. Histogram of the difference in c-axis orientation of host grains and adjacent neoblasts. After Ransom (1971)
Figure 14. Contoured c-axis fabrics of quartzites in western Big Wash. Sample locations on the section shown in Figure 5 given for each sample in meters. Projections as for Figure 11.
Although the central portion of the girdle of BW 10 is oriented almost perpendicular to the foliation, the girdle as a whole has a slight clockwise vergence. Rhombohedral glide planes were the most active.

The c-axis fabric of BW 25 bears little relationship to the mesoscopic fabric of the rock. There is very weak girdle development. This sample contains >10 percent plagioclase and muscovite. An effort was made to avoid areas rich in minor phases, but they may have affected the fabric development.

Sample BW 47 was taken from quartzites just beneath the mylonitic schists at the top of the metasedimentary section. BW 47 has a very poorly defined pattern with perhaps two poorly developed girdles intersecting at Y.

The samples from the base to the upper part of the section display an upward increase in fabric intensity. However, those at the top tend to have irregular fabrics.

Discussion

There are two points to be discussed in the following section. First, there are distinct differences in the structures present at the Red Cloud "thrust" contacts on either limb of the Big Wash synform. What are the tectonic implications of these differences and how do they affect Powell's model?

The contact exposed on the east limb of the Big Wash synform is clearly a segment of the Early Red Cloud thrust as Powell (1981) defined
The quartzite microtextures suggest a high temperature deformation, which has annealed most of the strain indicators such as any textures reflecting high strains along the granitic gneiss/quartzite contact. The quartz crystallographic fabrics and a few asymmetric folds suggest west-vergent thrusting, but the data are not beyond reproach.

The contact on the west limb is clearly not the original Red Cloud thrust surface. The absence of the granitic gneiss on the west limb, the presence of chloritic breccias, and the mylonitization of the schists at the top of the Pinto quartzite are all atypical of the Red Cloud thrust. The low-grade mylonites and the low intensity quartz crystallographic fabrics at the top of the section suggest an overprinting of deformations.

The fact that dikes are cut by the western fault suggests a reactivation of the fault surface. James (1987) reported a 147±1 Ma U/Pb emplacement age for the dike swarm. Terres (1984) reports K/Ar ages as old as 100 Ma on the same dikes. James concluded that the quartz latites, as well as most of the mafic dikes in the eastern Eagle Mountains, are the southernmost known exposure of the Independence dike swarm. If this is true, then there is a conflict between the geochronology and crosscutting relationships in the Eastern Transverse Ranges. These dikes cut undeformed granitic rocks south of the area portrayed in Figure 4 that have been correlated with granitic rocks of late Cretaceous age in the Chuckwalla Mountains (Bob Powell, U.S.G.S., Spokane, WA, personal communication, 1987). The fault is also cut by the undeformed granites that are intruded by the dikes. This
causes conflicting cross-cutting relationships with the fault cutting the dike, the dike cutting the granite, and the granite cutting the fault. Additional fieldwork should clarify this contradiction.

The offset on the low-temperature shear zone cutting the dikes need not be great. Quartz latite dikes are present on either side of the shear zone. It may die-out along strike and be absent to the south.

Because the east-vergent fabrics in the mylonitic schists are cut by the dikes, they are perhaps more difficult to explain. Although it predates the intrusion of the dike swarm, this east-vergent deformation is believed to be a retrograde event as well. The lack of strong quartz fabrics at the top of a sequence of samples with increasing fabric intensities seems unusual and may argue that the upper part of the section has been affected by a later event. Strong west-vergent quartz fabrics are present in Cholla Wash quartzites, just 8 km west.

Figure 15 shows a schematic section across the Big Wash synform. The reactivation of the west limb fault and truncation of the granitic gneisses are shown, but the geometries at depth are uncertain.
Figure 15. Schematic cross section of the Big Wash area
THE SOUTHWEST HEXIE MOUNTAINS

Introduction

Powell (1981) mapped an east-dipping mylonite zone in the western Hexie and Cottonwood Mountains, along the southwestern boundary of Joshua Tree National Monument. He believed that this shear zone was the westernmost exposure of the Red Cloud thrust system in the Eastern Transverse Ranges.

A structural map of the southwest Hexie and southeast Little San Bernardino Mountains is shown in Figure 16. The map area has been divided into three domains. The domains are separated from one another by distinct structural boundaries.

Descriptions of structural domains

Eastern Domain

The eastern domain is composed of a north-northwest trending antiform of migmatitic pelitic gneisses that Powell (1981) termed Hexie gneiss. These gneisses possess a lineation that parallels the axis of the associated antiform (Figure 17a). The lineation is defined by the axes of pytgmatically folded migmatitic layers. The kinematic significance of this structure is unclear.

The pelitic gneisses have undergone a pervasive greenschist facies retrograde metamorphism. Garnet appears green in hand specimen due to nearly complete alteration to chlorite. Other clumps of pale chlorite
Figure 16. Geologic map of the southwestern Hexie Mountains and southeastern Little San Bernardino Mountains

EXPLANATION

Kg  Cretaceous granitic plutons
mg  Mylonitic gneiss
lg  Leucocratic orthogneisses
pg  Paragneisses (Lost Horse lithosome of
     of the Joshua Tree terrane)
hg  Hexie gneiss

JTNM  Joshua Tree National Monument
Figure 17a. Foliation poles (open squares) and lineations (filled circles) for the Eastern Domain

Figure 17b. Foliation poles (open squares) and lineations (filled circles) for the Central Domain

Figure 17c. Fold axes for the Central Domain. Solid squares are "S" folds, open squares are "Z" folds, small diamonds are symmetric folds

Figure 17d. Foliation poles (open squares) and lineations (filled circles) for the Western Domain
chlorite are also present. It is unclear what mineral this chlorite might be replacing. Plagioclase has a cloudy appearance. In spite of the retrograde effects, these gneisses have textures similar to the higher grade rocks found else where in the Eastern Transverse Ranges. Quartz tends to occur as single, undulatory grains. Plagioclase is xenomorphic. Biotite tends to occur as discrete grains.

Central Domain

The central domain is separated from the eastern domain by a relatively abrupt change in structural style. There is a distinct change in lineation trend, from north-northwest to northeast, and foliation attitude, from steeply west dipping to east dipping (Figures 16, 17b). In the easternmost portion of the central domain, a crenulation lineation is present. The crenulation is apparently a product of the overprinting of the east dipping foliation of the central domain on the foliation of the eastern domain.

The gneisses of the central domain have a better developed foliation than the gneisses in the eastern domain. The foliation is parallel to the compositional layering in the rock. The central domain is more heterogeneous lithologically than the eastern domain. Quartzofeldspathic gneisses are more common.

There are west-vergent folds present in the central domain (Figure 17c). An interesting feature of the central domain is the asymmetric boudins (Figure 18a). Unlike most boudin structures, the boundaries between individual boudins are not perpendicular to the foliation.
Figure 18a. Asymmetric boudins in the central domain

Figure 18b. Relict andalusite porphyroblasts in the pelitic lithosome of the Lost Horse granofels

Figure 18c. Late syn-tectonic dike cutting mylonitic foliations

Figure 18d. Altered igneous plagioclases. Large epidote crystals are evident within the highly twinned feldspars. Horizontal field of view is 2.8 mm
Instead they form an acute angle with the foliation toward the west in all cases. The long axes of the boudins parallel the local fold axes and are roughly perpendicular to lineation. In fact, these boudins may act as nucleation sites for folds. If these boudins formed with the plane of their pull-aparts perpendicular to the foliation, then they may suggest west-vergent shear. Assuming that the foliation approximates the flattening plane during a noncoaxial deformation deformation, a material line that was parallel to the pole of the foliation before shearing will rotate more rapidly than foliation itself. This will lead to a consistent shear-related asymmetry. Some boudins appear as if they have been squeezed back together, indicating changes in the kinematic framework during the deformation.

The base of the central domain is a mylonite zone. This is the structure that Powell mapped as a portion of the Red Cloud thrust. Lineations in the mylonites generally parallel those of the central domain and are included on Figure 17b. The mylonites are predominantly quartzofeldspathic in composition with type 2a and 2b quartz ribbons of Boullier and Bouchez (1978). Little evidence of strain is preserved in the microtextures of these mylonites. Shear indicators are uncommon but consistently west-vergent.

**Western Domain**

The western domain is composed of the Lost Horse granofels of Powell (1981) and a suite of orthogneisses whose protoliths intruded the metasediments. The Lost Horse granofels is a complex assemblage of
paragneisses and granofelses. It ranges from siliceous granofels to a
distinctive pelitic schist with what Powell (1981) described as bearing
relict andalusite porphyroblasts (Figure 18b). The orthogneisses whose
protoliths intruded the Lost Horse granofels range from amphibolites to
relatively leucocratic granitic gneisses. The more felsic orthogneisses
cross-cut the more mafic ones. Table 1 contains three analyses of
orthogneisses from the Little San Bernardino Mountains as well as one of
the quartzofeldspathic gneisses from the central domain (HM 47). The
orthogneisses are highly deformed over a kilometer away from the shear
zone. Closer to the shear zone the orthogneisses are finer grained with
feldspar porphyroclasts.

Foliation and lineation measurements are shown in Figure 17d. As with
the central domain, the foliation generally dips eastward. The
lineations are distributed in a great circle coincident with the average
foliation orientation. They show a greater scatter than the lineations
from the central domain (Figure 17b).

Powell interpreted the mylonite zone as a major tectonic contact.
The orthogneisses of the western domain are not found east of the
mylonite zone. However, lenses of the pelitic gneiss with the unusual
relict texture shown in Figure 18b are found in the western portion of
the central domain.

Interpretation

The central domain contains a number of shear indicators consistent
with west-directed thrusting. It has a consistent lineation
Table 1. Compositions* of orthogneisses in the southeastern Little San Bernardino Mountains and southwestern Hexie Mountains

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</tr>
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</table>

| Sc       | 23.35 | 33.46 | 10.17 | 15.58 |
| Cr       | 23.9  | 84.0  | 25.1  | 22.2  |
| Co       | 5.7   | 32.0  | 8.1   | 6.2   |
| Ni       | 7.8   | 39.0  | 13.7  | 10.3  |
| Zn       | 106.0 | 50.1  | 32.8  | 344.0 |
| Rb       | 78.1  | 27.0  | 113.2 | 69.1  |
| Sr       | 48.3  | 111.5 | 107.2 | 76.1  |
| Y        | 60    | 74    | 77    | 49    |
| Zr       | 263   | 245   | 429   | 249   |
| Nb       | 10    | 14    | 15    | 9     |
| Cs       | 0.5   | 0.7   | 0.5   | 0.6   |
| Ba       | 1396  | 325   | 727   | 1210  |
| Hf       | 8.0   | 8.4   | 12.7  | 7.9   |
| Ta       | 0.56  | 1.31  | 0.76  | 0.42  |
| Th       | 19.5  | 2.5   | 30.1  | 19.7  |
| U        | 5.5   | 1.5   | 2.6   | 3.0   |
| La       | 46.0  | 36.1  | 68.2  | 53.2  |
| Ce       | 92.15 | 85.06 | 132.7 | 101.6 |
| Sm       | 8.94  | 11.27 | 12.23 | 9.76  |
| Eu       | 1.35  | 1.95  | 1.49  | 1.48  |
| Tm       | 6.31  | 7.13  | 7.65  | 4.79  |
| Lu       | 1.017 | 1.14  | 1.23  | 0.803 |

*XRF analyses on fusion discs, pressed powder pellets and loose powders using a WDS Siemens XRF spectrometer and EDS Kevex XRF at ISU. Rare earths, Sc, Co, Cr, Cs, Hf, Ta, Th, and U determined by INAA at Columbia (MO) Research Reactor. Loss on ignition (LOI) at 850°C.
orientation, and its foliation apparently overprints the foliation of the eastern domain.

The presence of the pelitic units of the Lost Horse lithosome east of the mylonite zone implies that it is not a major tectonic contact. The structural data suggest that the entire central domain functioned as a broad zone of heterogeneous shear. The mylonite simply served as the locus for the last increments of deformation. Further evidence for the mylonite serving as a late stage movement zone is provided by the dike shown in Figure 18c. This porphyritic dike crosscuts the foliation in the upper part of the mylonite zone. It is less deformed than the adjacent units. Figure 18d shows a partially altered igneous plagioclase in thin section. The dike has been deformed into a Type I S-C mylonite of Lister and Snoke (1984). The fabric in the dike is a result of west-vergent shear. In the eastern portion of the central domain, similar dikes are undeformed. This suggests that the later movements in the central domain were along the western boundary of the central domain.

Structural data from the western domain are similar to central domain data, but have more scatter. These relations seem to imply that the fabric in the western domain is from the same event as the central domain fabric.

An interpretive section is shown in Figure 19. The central domain is inferred to be a shear zone approximately a kilometer wide. The section shows the general relationship between the central domain and the eastern domain. It is important to note that the change in foliation
Figure 19. Schematic cross section through the southwestern Hexie and southeastern Little San Bernardino Mountains.
attitude across the boundary is not the result of a synformal fold hinge along the contact. The foliation does not swing through horizontal.

The general parallelism of structures in the western and central domains suggests that the deformation of the western domain was related to that of the central domain. The broader spread in lineations in the western domain may indicate a pre-existing fabric. The general scarcity of shear-related fabrics in the western domain may suggest that the deformation was of a more coaxial nature, although that relationship cannot be quantified.
CHUCKWALLA MOUNTAINS

Introduction

Although the northern and eastern Chuckwalla Mountains are composed largely of Cretaceous plutonic rocks, pre-batholithic units involved in the Red Cloud thrust deformation crop out over much of the rest of the range. On the generalized geologic map of the Chuckwalla Mountains (Figure 20), the most obvious structural feature is the apparent antiform-synform defined by the Joshua Tree granite gneiss. The Joshua Tree granite gneiss is entirely fault-bounded in the Chuckwalla Mountains. The other pre-batholithic units are elements of the San Gabriel terrane.

The western portion of the Chuckwalla Mountains was chosen for detailed study because both the Early and Late Red Cloud thrust faults are exposed there, and because contacts between various units of the San Gabriel terrane crop out in the area. The various lithosomes of the western Chuckwalla Mountains and their contact relationships are described in the following section. This is followed by a discussion of the mesoscopic and microscopic structures and discussion of possible tectonic interpretations. The Early Red Cloud thrust is offset 11 km westward by the Chiriaco fault (Powell, 1975), north of the Chuckwalla Mountains. It continues in the southern Eagle Mountains in the vicinity of Cholla Wash (Figure 1). This area will be discussed following the western Chuckwalla Mountains section.
Figure 20. Geologic map of the Chuckwalla Mountains (after Powell, 1981; Powell et al., 1984)
Lithologic units of the Chuckwalla Mountains

**Joshua Tree granite gneiss**

The Joshua Tree granite gneiss was named informally by Powell (1981). In the Chuckwalla Mountains, it is a homogenous, well foliated augen gneiss consisting of microcline augen in a quartz-plagioclase-microcline-biotite matrix. The foliation is defined by discontinuous, biotite-rich layers. The Joshua Tree granite gneiss tends to be well lineated. Chevron-style folds are the most common fold type in this unit. Silver (personal communication in Powell, 1981) reported a minimum U/Pb age of 1.65 Ga for the Joshua Tree granite gneiss.

**Biotite gneiss and pelitic gneiss**

The biotite gneiss and pelitic gneiss comprise Powell's Hexie gneiss unit. The biotite gneiss contains the assemblage: quartz + plagioclase + biotite + muscovite ± orthoclase ± garnet. It is a relatively dark colored unit with ptygmatite quartz-feldspar veins. Millimeter-scale gneissic layering is present in some samples. Whole-rock analyses of the biotite gneiss are shown in Table 2.

The pelitic gneiss contains the assemblage:

quartz + muscovite + biotite + plagioclase + orthoclase + chlorite ± sillimanite ± garnet.

The sillimanite has largely been replaced by clots of muscovite and chlorite. The garnet is generally partially replaced by chlorite. An analysis of a pelitic gneiss is also shown in Table 2.
Table 2. Compositions of biotite and pelitic gneisses from the western Chuckwalla Mountains and Cholla Wash in the Eagle Mountains

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<td>99.79</td>
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<td>98.64</td>
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Sc      | 21.10 | 6.29  | 13.65 | ND    |
Cr      | 2.4   | 11.2  | 9.1   | 47.4b |
Co      | 6.2   | 4.3   | 7.0   | ND    |
Ni      | 9.8   | 7.8   | 6.4   | 22.9  |
Cu      | 11.0  | 7.0   | 51.8  | 7.0   |
Zn      | 103   | 65    | 100   | 98    |
Rb      | 192.0 | 138.3 | 122.9 | 206.3 |
Sr      | 100.5 | 153.2 | 194.0 | 113.4 |
Y       | 154   | 34    | 38    | 26    |
Zr      | 404   | 221   | 186   | 274   |
Nb      | 49    | 15    | 14    | 14    |
Cs      | 11.3  | 9.0   | 8.6   | ND    |
Ba      | 311   | 1740  | 576   | 1003  |
Hf      | 12.7  | 7.0   | 5.6   | ND    |
Ta      | 16.05 | 1.43  | 1.02  | ND    |
Th      | 22.4  | 15.7  | 11.3  | ND    |
U       | 14.1  | 5.6   | 5.1   | ND    |
La      | 57.4  | 35.2  | 33.1  | 74.3b |
Ce      | 121.5 | 75.51 | 64.7  | 136.5b|
Sm      | 12.74 | 6.52  | 6.63  | ND    |
Eu      | 1.16  | 0.85  | 1.13  | ND    |
Yb      | 18.30 | 3.47  | 3.69  | ND    |
Lu      | 2.890 | 0.569 | 0.658 | ND    |

aND = not determined.
bValues determined by XRF, not INAA as in the other samples.
The biotite gneiss-pelitic gneiss boundary is inferred to be a stratigraphic one in Red Cloud Canyon. The contact can generally be located to within a meter or two, and no signs of high ductile strains are evident.

**Soledad augen gneiss**

The Soledad augen gneiss is a dark colored augen gneiss that Powell described as granodioritic to monzogranitic in composition. It consists of 1-5 cm perthitic microcline augen in a quartz-biotite-plagioclase matrix. Feldspars appear fractured in thin section. Quartz occurs as undulose grains with sutured boundaries.

In Red Cloud Canyon, the contact between the Soledad augen gneiss and the biotite gneiss is a meta-intrusive one (Figure 21a). The Soledad augen gneiss was not found in contact with the pelitic gneiss. The Soledad augen gneiss is typically well foliated and lineated in the Red Cloud Canyon area, but textures range from mylonitic to only slightly foliated. The augen gneiss locally contains chevron folds. Silver (1971) reported an age for this unit of 1.65 Ga.

**Granitic gneiss suite**

Powell (1981) used the name "Augustine gneiss" to describe a heterogeneous suite of orthogneisses. He named the unit for exposures in Augustine Pass in the southwestern Chuckwalla Mountains. He believed that portions of the gneiss had textures indicative of retrograde metamorphism of granulite facies assemblages and that portions of this
Figure 21a. Biotite gneiss/Soledad augen gneiss contact in Red Cloud Canyon, Chuckwalla Mountains

Figure 21b. Example of banded textures in the granitic gneiss

Figure 21c. Photograph of banded gneiss cut by an unbanded, poorly foliated unit of the granitic gneiss suite

Figure 21d. Block of Soledad augen gneiss in granitic gneiss
unit were correlable with the Mendenhall Gneiss of the Western Transverse Ranges. Powell (personal communication, 1987) currently believes that little, if any, of the orthogneiss he mapped as Augustine gneiss in the central and western Chuckwalla Mountains has been retrograded from granulite facies. These leucocratic granitic gneisses crop out over much of the southern and western Chuckwalla Mountains, eastern Orocopia Mountains, and, as will be discussed below, in the Little Chuckwalla Mountains as well. They have also been recognized in the eastern Hexie and southwestern Pinto Mountains. Although there is a continuum, the leucocratic orthogneisses have two general textural types. They are either massive, lepidoblastic-granular with a weak foliation defined by individual biotite grains, or they possess a centimeter-scale gneissic banding caused by discrete biotite-rich layers (Figure 21b). The first textural type is generally coarser and cross-cuts the second (Figure 21c). Uncommon coarse-grained veins of pegmatitic material are found cross-cutting all of the above.

Locally these gneisses contain lenses of amphibolite. Biotite schlieren are common. In the Chuckwalla Mountains, xenoliths of the Soledad Augen gneiss are most common (Figure 21d). These range in size from individual microcline crystals to blocks of mappable scale. They are apparently free-floating blocks within the complex. The granitic gneisses are composed of approximately equal amounts of perthitic microcline, plagioclase and quartz. Biotite generally comprises less than ten percent of any sample. Centimeter-scale ellipsoidal segregations of leucocratic minerals cored by magnetite are found in
laminated granitic gneisses near the Soledad augen gneiss. This is similar to the flecked gneisses described by Trumbull (1988) in anatectic gneisses in Colorado. Primary igneous textures are not preserved in any samples collected. In thin section, quartz typically forms single ovoid grains with slight to moderate undulatory extinction. Plagioclase is unzoned and variably altered. Microcline and microperthite and rare orthoclase are generally fresh. Myrmekite is present in most sections along potash feldspar/plagioclase boundaries. Feldspars are xenomorphlc and embayed by quartz in some sections (Figure 22a). This texture was noted by Crowell and Walker (1962) in both pre-1220 Ma (Silver, 1971) gneisses and in their blue-quartz granites in the western Orocoipa Mountains.

Geochemistry of the Chuckwalla Granitic Gneiss Suite

Because the origin of the granitic gneisses is uncertain, a reconnaissance study of whole-rock chemistry was undertaken. Five samples of leucocratic gneiss and one sample of amphibolite (CM83) from the western Chuckwalla Mountains were analyzed along with two samples of granitic gneiss from the west side of Gulliday Wash in the southern Chuckwalla Mountains and one sample from the Little Chuckwalla Mountains. The results are presented in Table 3.

**Major Elements**

Two samples have more than 8 wt% K2O (CM1, CM91) though apparently as a result of different processes. compared to the average leucogneiss
Figure 22a. Photomicrograph of a granitic gneiss sample showing embayed microcline and ovoid quartz (Q). Horizontal field of view is 2.8 mm

Figure 22b. Complex textures in the granitic gneiss complex

Figure 22c. Evidence of multiple injections of magma in the protolith of the granitic gneiss

Figure 22d. Biotite gneiss/Soledad granodiorite contact in Fried Liver Wash, Hexie Mountains
Table 3. Compositions of orthogneisses from granitic gneiss complex

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Cr 123  ND 13.3  8.9  7.1  4.3  
Co 47.5  ND 4.6  2.2  2.1  1.9  
Ni 6.4  5.4  5.9  5.4  6.9  6.9  
Cu 37.8  7.0  7.4  9.0  6.2  12.6  
Zn 141  90  67  45  44  86  
Rb 17.1  267.9  113.2  146.9  162.3  234.2  
Sr 125.9  43.9  217.0  121.6  194.0  109.6  
Y 36  69  67  24  18  12  
Zr 90  266  283  275  157  156  
Nb 6  16  18  14  6  18  
Cs 0.9  ND 2.1  1.4  1.5  1.4  
Ba 351  1713  2066  1118  1177  628  
Hf 3.0  ND 9.7  7.8  6.0  5.1  
Ta 0.63  ND 0.79  0.69  0.24  1.05  
Th 2.4  ND 53.3  29.7  44.7  38.7  
U 1.9  ND 6.1  5.0  4.38  9.6  

La 13.9  36.5ᵃ  118.6  67.3  67.4  29.1  
Ce 32.82  136.0ᵃ  223.52  131.32  126.91  58.07  
Sm 5.59  ND 14.2  9.06  10.61  3.88  
Eu 1.85  ND 1.64  1.55  1.29  0.61  
Tb 0.96  ND  ND  ND  ND  ND  
Yb 3.815  ND  5.21  1.67  0.99  1.03  
Lu 0.595  ND  0.828  0.253  0.199  0.30  

ᵃND = not determined. Dashes mean below detection.
ᵇDetermined by XRF instead of INAA.
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(as well as CM91), CM1 is low in Al₂O₃, CaO, Na₂O, and Sr. It is high in Fe₂O₃, MgO, MnO, Zn, Rb, and Y. However, CM91 is high in Al₂O₃ and low in Fe₂O₃, MgO, MnO, CaO, Sc, Co, Zn, Sr, Zr, Nb, and Ba.

Table 4 shows the general range in compositions for granitic rocks based on their inferred source material. The I- and S-type granites were first defined by Chappell and White (1974). These granitic rocks had igneous and sedimentary sources respectively. The data in the table are from the Koskoisko Batholith in southeast Australia. The A-type granite data are from Collins et al. (1982). A-type granites are inferred to have been derived by dry melting of felsic granulite.

The leucogranites and migmatites are thought to be the products of melting of crustal, generally metasedimentary material. A continuum of increasing melting can be inferred from the compositional changes from migmatites through leucogranites to S-type granite. The leucogranites described in the table are the Hercynian syn-metamorphic intrusions of Didier et al. (1982) and Strong and Hanmer (1981). Some information is from the average world-wide 2-mica granite as listed in Le Fort (1981). The migmatites are the Arvika migmatites described by Henkes and Johannes (1981) from Sweden. These migmatites were derived from partial melting of paragneisses of arenitic composition.

The leucogneisses of this study have relatively high K₂O and low color index and CaO. Most of the major-element characteristics are most similar to the Arvika migmatites and leucogranites. The highly potassic nature of these rocks, their most unusual major-element feature, is most similar to that of the migmatites. Figure 23 is modified from Johannes.
### TABLE 4. Differences in composition of granites from different sources

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<th>K₂O (mean)</th>
<th>K₂O (range)</th>
<th>CaO</th>
<th>CaO (mean)</th>
<th>CaO (range)</th>
<th>Zr</th>
<th>Zr (mean)</th>
<th>Zr (range)</th>
<th>Ba</th>
<th>Ba (mean)</th>
<th>Ba (range)</th>
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Notes: Data for I- and S-type granites are from Hine et al. (1978). The data for the A-type granites are from Collins et al. (1982). The leucogranite data is from Didier et al. (1982), Strong and Hanmer (1981), and Le Fort (1981). The migmatite data is extrapolated from graphs in Johannes (1983).

*A Rock types as follows: gd = granodiorite, qd = quartz diorite, md = monzodiorite, mg = monzogranite, sg = syenogranite.*
Figure 23. Ternary normative albite (Ab), quartz (Qz), and orthoclase (Or) plot for granitic gneisses in the western Chuckwalla Mountains with fields for various leucosomes of Johannes (1983)
(1983) and shows the normative compositions of the Chuckwalla leucogneisses compared to the compositional ranges of the various migmatites studied by Johannes. Except for the most potassic sample, the Chuckwalla leucogneisses have similar compositions. Johannes interpreted his rocks to have a cotectic composition.

**Trace elements**

The trace element abundances for the Chuckwalla leucogneisses are also presented in Table 4. Variation diagrams (Figure 24) show that Rb and Sr concentrations lack a good correlation with total iron. Alderton et al. (1986) demonstrated that these traces are easily mobilized during alteration. Other less mobile trace elements are related to total iron in a more linear fashion.

In spite of the major element similarities with Hercynian and Himalayan leucogranites, the trace element abundances are quite different. The Chuckwalla leucogneisses have much higher Th, Zr, and especially Ba. In these respects they are more similar to the A-type granites of Collins et al. (1982). Harris et al. (1986) employed trace element discrimination diagrams as indicators of tectonic environment and source material. In Figure 25 it is evident that the Chuckwalla leucogneisses plot in the Volcanic Arc Granite field, regardless of questionable Rb values. This is largely the result of low Ta values, which seem unusually low when plotted on the "spider" diagrams of Pearce et al. (1984). Pearce et al. (1984) used Rb as the primary discriminant of volcanic arc and syn-collisional settings. Most of the leucogneisses
Figure 24. Trace element variation diagrams for the granitic gneisses. Sample CM1 is plotted as a triangle.
Figure 25. Hf-Rb-Ta discrimination diagram from Harris et al., (1986). VAG = Volcanic arc granites. STG = syntectonic granites, L-POG = Late-postorogenic granites, WPG = Within plate granites.
plot along the boundary between the two, but as mentioned above, the Rb abundances of Chuckwalla samples are probably not representative of the original magma. Discrimination diagrams are difficult to apply to granitic rocks because of the complexities involved in the melting, movement, and emplacement of granitic magmas.

Similarities in chemistry of the leucogneisses with leucogranites and migmatites suggest a crustal source and an association with orogenesis. Textures in these gneisses indicate syntectonic magmatism. This is based on the decreased deformation in the later portions of the complex and the complex foliation relationships. There is insufficient data to define a source for the melts, so it is not clear if they formed locally, or at deeper levels. The most likely model for the granitic gneisses is melting of crustal material and emplacement of the resulting magma during the high-temperature Red Cloud thrust deformation.

**Age of the granitic gneisses**

Foliated plutonic rocks east of Graham Pass in the Little Chuckwalla Mountains have been correlated with the Mount Lowe Intrusion (as defined by Barth, 1985) of the western Transverse Ranges (R. Tosdal, 1986, his Figure 1; R. Powell (U.S.G.S., Spokane, WA) and R. Tosdal (U.S.G.S., Menlo Park, CA) personal communication, 1987). The Mount Lowe Intrusion has a well constrained age of emplacement of 220 ± 10 Ma (Silver, 1971). In the Little Chuckwalla Mountains, the Mount Lowe correlative is cross-cut and intruded by leucocratic granitic rocks that share the same foliation as the Mount Lowe. These foliated
leucogranites are very similar texturally and chemically to portions of the granitic gneiss suite (See Table 4; sample LCM 7 is a foliated leucogranite from the Little Chuckwalla Mountains). If these granitic gneisses are members of the same generation, then western Chuckwalla granitic gneisses can be no older than Late Triassic in age. Future geochemical investigations may improve this correlation.

Structural geology of the Western Chuckwalla Mountains

The general structure of the Chuckwalla Mountains is shown in Figure 20. The western Chuckwalla Mountains were studied in detail and a map of that area is shown in Figure 26. The various lithosomes are exposed in north to northwest striking bands or lenses. At the northeastern corner of the area (Figure 26) a northeast dipping mylonite zone separates granitic gneisses and Soledad augen gneiss from an overlying plate of Joshua Tree granite gneiss. As on Powell's map, this is correlated with the Late Red Cloud thrust. The granitic gneisses and Soledad augen gneiss form the core of Powell's antiform. To the west these units are truncated by vertical to steeply west dipping mylonite zone that Powell defined as the Late Red Cloud thrust. The sheet of Joshua Tree granite gneiss that lies west of the Late Red Cloud thrust is bounded on the west by the Early Red Cloud thrust. The Early and Late thrusts were named here for the Red Cloud Mine which lies between these two shear zones in upper Red Cloud Canyon. The biotite and pelitic paragneisses crop out west of the Early Red Cloud thrust. A continuous sheet of Soledad augen gneiss dips steeply to moderately east
Figure 26. Geologic map of the western Chuckwalla Mountains
beneath the paragneisses. The westernmost Chuckwalla Mountains is composed of the migmatitic granitic gneiss suite with sheets and lenses of Soledad augen gneiss inclusions.

The easternmost Orocopia Mountains, too, are composed of granitic gneisses. The Early Red Cloud thrust is not exposed here as shown in Powell (1981). Inclusions of Soledad gneiss are not found, but near the western limit of exposure limit of prebatholithic rocks, blocks of siliceous paragneisses become abundant.

**Foliations**

With the exception of the crenulation cleavages described below in the section on folding, all of the gneisses in the Chuckwalla Mountains display a single penetrative foliation. Although the granitic gneisses have some very complicated cross-cutting fabrics, cross-cutting foliations are not present (Figure 22b).

In Red Cloud Canyon, the Soledad augen gneiss and the biotite gneiss have the same foliation (Figure 21a). In Fried Liver Wash, in the central Hexie Mountains, undeformed Soledad porphyritic granodiorite intrudes the biotite gneiss (Figure 22d). Therefore, the foliation of the biotite gneiss in the Hexie Mountains is older than the foliation present in the Chuckwalla Mountains.

The foliation is generally steep throughout the area (Figure 26, 27a-d). It is mostly east-dipping, except in the paragneisses adjacent to the Early Red Cloud thrust (Figure 27a), where dips steeply west and in the westernmost Chuckwalla Mountains in the core of the
Figure 27a. Orientations of foliation poles (open squares) and lineations (filled circles) in the paragneisses

Figure 27b. Orientations of foliation poles (open squares) and lineations (filled circles) in the Soledad augen gneiss west of the Early Red Cloud thrust

Figure 27c. Orientations of foliation poles (open squares) and lineations (filled circles) in the granitic gneisses in the Western Chuckwalla Mountains west of the Early Red Cloud thrust

Figure 27d. Orientations of foliation poles (open squares) and lineations (filled circles) in the granitic gneisses of the eastern Orocopia Mountains
granitic gneiss complex where foliation attitudes become less consistent (Figure 27c). Westerly dips are also seen in the Orocopia Mountains in the westernmost exposures (Figure 27d).

Lineations

Most of the gneisses possess a single lineation. It is best defined by pressure shadows in the augen gneisses. The lineation in the Red Cloud Canyon area trends east-northeast (Figure 27a-b). In the westernmost Chuckwalla Mountains and the eastern Orocopia Mountains, lineation trends are more scattered, but are similar (Figure 27c-d).

Folds

Folds have been recognized in all the gneisses of the Chuckwalla Mountains. There appears to be a single set of folds associated with the foliation and lineation present in the various units. Most fold hinges lie at high angles to the local stretching lineation (Figures 28a-c, 29a). Shapes range from open to isoclinal. Folds in the paragneisses tend to be more cylindrical than those in the granitic gneisses. Most folds are roughly similar and affect only a limited thickness of gneiss, but folds in augen gneisses tend to be chevrons that are quite continuous perpendicular to the layering. Some complex fold shapes were recognized (Figure 29b). A few sheath-like folds were found in the orthogneisses. The majority of the asymmetric hinges measured in the western Chuckwalla Mountains are west-vergent (Figure 28a-c). Powell's 1981 model predicts this area to be the exposed overturned limb of a
Figure 28. Fold data from the western Chuckwalla Mountains. Filled squares = "S" folds, open squares = "Z" folds, small diamonds = symmetric folds. (a) paragneisses (b) Soledad augen gneiss west of Red Cloud Canyon (c) granitic gneiss suite
Figure 29a. F2 fold with highly appressed limbs. Note the hinge lies at a high angle to the lineation

Figure 29b. Complex fold in the granitic gneiss complex

Figure 29c. Incipient F3 fold with associated foliation

Figure 29d. Outcrop of F3 folds complexly folded by F3 folds
west-vergent antiform (Figure 3). The folds measured are roughly parallel to the antiformal axis. If they were related to the west-vergent antiform, folds of the opposite vergence might be expected on the overturned limb (Figure 30, from Ramsay, 1967; page 355). The sense of asymmetry of these folds indicate that they predate the macroscopic structure.

The folds display a wide range of interlimb angles and orientations. On a plot of lineation/hinge angle versus interlimb angle a definite relationship can be seen (Figure 31). Folds with large lineation/hinge angles have a wide range of interlimb angles, whereas folds with small lineation/hinge angles have only small interlimb angles.

The fact that these folds fold the foliation implies that they postdate the deformation that formed the foliation. However, the relatively consistent lineation trend indicates that if the folding is not related to foliation development, then it was at least formed in a similar kinematic framework.

There are isoclinal folds that parallel the lineation. The foliation is axial planar to these isoclines. In traditional structural analysis, these isoclines would be considered a separate fold generation (F1) because of their geometric relationship to the foliation and lineation. The dominant fold set would then be termed F2. However, the relationship between the folds and their associated lineation in Figure 31 suggests that at least some of the isoclines could have been generated during the same event that formed the more open folds.
Figure 30. Schematic fold showing first, second, and third order structures (from Ramsay, 1967)
Figure 31. F2 fold data from the western Chuckwalla Mountains plotted on a lineation/hinge angle vs. interlimb angle diagram.
At first glance, Figure 31 implies generation and rotation of folds during a single deformation as suggested by Bryant and Reed (1969) and Escher and Watterson (1974). However, pre-existing folds would tend to fall within the same field on Figure 31. Folds that postdate the foliation/lineation forming deformation would not necessarily do so. Sanderson (1973) tested a pure-shear model for the rotation of fold hinges due to stretching in the XY plane. He used a lineation/hinge angle (theta) histogram of field data to test his model. His data show an asymmetric population with a single peak (Figure 32a from Sanderson, 1973; Figure 8). Chuckwalla Mountains data displayed in the same manner yield a bimodal distribution (Figure 32b). This may suggest multiple fold events, reflect a sampling bias, or indicate that the Sanderson model is inappropriate for the folds in the western Chuckwalla Mountains.

F3 folds are a locally intensely developed fold set. They are best developed on the ridge west of Powell's (1982) field trip stop #1 at the southwestern corner of the Chuckwalla Mountains where they are found in the pelitic gneiss adjacent to the Early Red Cloud thrust. These folds are characterized by a strong crenulation foliation and lineation (Figure 29c, d). The F3 folds trend southwest and have nearly horizontal axial planes (Figure 29d). Where the asymmetry of individual hinges is clear or where medial surfaces can be used, these folds generally verge northwest.
Figure 32a. Histogram of lineation/hinge angle from Sanderson (1973, Figure 8)

Figure 32b. Histogram of lineation/hinge angle of F2 folds in the western Chuckwalla Mountains
Figure 33. Geologic map of the southwestern Chuckwalla Mountains

EXPLANATION

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<td>Pinto quartzite</td>
</tr>
<tr>
<td>JTgg</td>
<td>Joshua Tree granite gneiss</td>
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Red Cloud thrust

0 km
Early Red Cloud Thrust

Powell called the surface that separates the Joshua Tree granite gneiss from the paragneisses in Red Cloud Canyon the Early Red Cloud thrust for its exposure at the Red Cloud Mine. The contact is vertical to steeply west-dipping in Red Cloud Canyon. It does not truncate the foliation. Lineations are approximately parallel on either side of the contact. The most interesting feature of the Early Red Cloud thrust is the apparent lack of thrust-related strain in the adjacent paragneisses. The Joshua Tree granite gneiss is strongly laminated and augen-poor in the 10-50 m closest to the contact, suggesting that some higher strain is recorded in that unit closer to the thrust. Segments of the contact are intruded by coarse-grained orthogneiss similar to portions of the granitic gneiss complex. In these orthogneisses, as in the granitic gneiss suite, feldspars are xenomorphic. Perthites are common, and plagioclase is moderately sericitized. Quartz occurs either in single ovoid grains or irregular clumps of interpenetrating grains. It is slightly undulose.

Some portions of the paragneiss-Joshua Tree granite gneiss contact are marked by a narrow, dark band of mylonite, indicating that the original Early Red Cloud thrust has been locally reactivated.

As described by Powell (1982) at field trip stop #1, the Early Red Cloud thrust dips moderately westward at the southwestern corner of the Chuckwalla Mountains (Figures 33, 34). It is not clear if this change in attitude is a result of folding of the thrust around the regional antiform-synform structure, or if it is a result of tilting associated
Figure 34. Structural data from the southwestern Chuckwalla Mountains. (A) foliation poles (open squares) and lineations (filled circles) (B) generally west-vergent F2 folds (diamonds) and generally northwest vergent F3 folds (filled triangles)
with the western termination of the Ship Creek fault. As shown on Figure 33, a thin lens of Pinto quartzite is present along the thrust above the Joshua Tree granite gneiss.

Late Red Cloud Thrust

The Late Red Cloud thrust as described by Powell (1981) is exposed across the central Chuckwalla Mountains (Figure 20). The Late Red Cloud thrust truncates units of the San Gabriel terrane east of Red Cloud Canyon. Samples taken from along the contact are mylonitic, with a steeply plunging lineation. Feldspars are fractured, not uncommonly by a series of en echelon microfaults (Figure 35b). Simpson and Schmid (1983) state that faulted porphyroclasts can be used as reliable shear indicators. Microfaults with recognizable offsets of feldspars were seen in each of three thin sections cut parallel to the XZ plane (Figure 36). Two roughly similar populations of microfaults with opposing dips show no tendency to asymmetric populations or shear sense. All but one of the microfractures have normal offsets.

Quartz in these samples consists of ribbons showing varying degrees of recrystallization (Figure 35b). At one extreme are ribbons formed by a single old grain that is highly undulose with distinct deformation bands. At the other extreme are polycrystalline ribbons of small, poligonal grains (type 2a ribbons of Boullier and Bouchez, 1978). Where recrystallization of the old, monocrystalline ribbons is complete, the neoblasts all have similar crystallographic orientations.
Figure 35a. Photomicrograph of feldspar in the Late Red Cloud thrust zone. Width of field of view = 4.8 mm

Figure 35b. Photomicrograph of Late Red Cloud thrust mylonite Width of field of view = 4.8 mm

Figure 35c. Portion of fold train in right-lateral mylonite in Cholla Wash

Figure 35d. Narrow shear zones near Early Red Cloud thrust in Cholla Wash. RCT = Red Cloud thrust. Unlabeled arrows point at vertical shear zones
Figure 36. Histogram of feldspar microfault orientations in mylonitic gneisses adjacent to the Late Red Cloud thrust. Normal faults are stippled, reverse faults are black.
West-vergent shear indicators were found in the Late Red Cloud thrust mylonites on the east side of Red Cloud Canyon. No unequivocal shear indicators were found at other locations. The steep lineation in the Late Red Cloud thrust mylonites is similar to that in the adjacent units, suggesting a relationship between this structure and the regional deformation. But the truncation of the Soledad and granitic gneisses as shown in Figure 27 as well as the feldspar and quartz microtextures suggest that this is indeed a retrograde structure.

Introduction to the Cholla Wash Area

The Early Red Cloud thrust is exposed in and along the east slope of Cholla Wash along the southern edge of the Eagle Mountains. As in Red Cloud Canyon, the thrust separates Joshua Tree granite gneiss from an overlying complex of pelitic, biotite, Soledad, and banded granitic gneisses (Figure 37).

In the upper reaches of Cholla Wash, a mylonite zone with right-lateral displacement cuts up section through the Joshua Tree granite gneiss and merges with the Red Cloud thrust. The mylonite places Pinto quartzite in contact with the upper plate rocks. This shear zone and others related to it will be discussed in a later section.

In lower Cholla Wash, the foliation dips steeply west with steeply plunging lineations (Figures 37, 38a). Parallelism of stretched conglomerate pebbles in the Pinto quartzite with the gneissic lineation in Cholla Wash (Figure 38a) supports a stretching origin for the lineation. Toward the top of the wash, the foliation dips more
Figure 37. Geologic map of Cholla Wash, southern Eagle Mountains

EXPLANATION

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JTNM Joshua Tree National Monument
Figure 38. Structural data from Cholla Wash. (A) foliations in the gneisses (open squares), lineations in the gneisses (filled circles), Pinto quartzite foliations (open triangles), and Pinto quartzite lineations (open circles) (B) "S" folds (filled squares), "Z" folds (open squares), and symmetric folds (small diamonds)
moderately to the west, particularly in the Pinto quartzite. Folds in the upper plate gneisses have northward plunging axes (Figure 38b). Although more hinges verge west than east, the majority of the folds measured are symmetrical.

**Pinto Quartzite**

The Pinto quartzite is well exposed on the east slope of Cholla Wash. The depositional contact between the Joshua Tree granite gneiss and the metasediments, as described in the Big Wash area, is present. Pebbles in the metaconglomerates are oblate (Figure 10), in contrast to pebbles in western Big Wash.

The quartz is finer grained in the Cholla Wash samples than in the Big Wash samples. Grain diameters were measured in seven Cholla Wash samples and in all cases averaged around 0.3 mm. There is less evidence of annealing and grain boundary migration than in the Big Wash samples. The samples generally contain a few ribbon grains, but most of the grains are small and polygonal.

**C-axis Quartz Fabrics**

C-axis quartz fabrics were measured on 6 samples of Pinto quartzite (Figure 39) collected within about 500 m of the thrust. Most samples show a single girdle or modified type I cross girdles. Four samples have west vergent fabrics (CW3, CW5, CW6, CW48); CW 45 has an east-vergent type I cross girdle. Samples CW 5, 6, 45, and to a lesser extent CW 3 all show strong activity on the prism glide and rhombs.
Figure 39. C-axis quartz fabrics from the Pinto quartzite in Cholla Wash. Projections as in Figure 11.
Sample CW 48 has maxima near the XZ plane indicating activity on the basal glide.

**Evidence for dynamic recrystallization**

Microtextures indicate dynamic recrystallization was an active process in the deformation of these quartzites. The process of dynamic recrystallization implies grains nucleating and growing during deformation. As the grains grow, they are deformed, presumably by intracrystalline glide. When the grains have accumulated enough dislocations they are no longer stable and new, small grains nucleate through subgrain rotation or grain boundary migration. If the rocks have not annealed since deformation, then certain relationships might be expected in the microtexture of the rock.

1). Since the grains grow as they are deformed, the larger grains should be more elliptical.

2). If the quartz grains are undergoing intracrystalline glide, the grains may be rotating toward some preferred crystallographic orientation allowing easy slip on the most active glide planes. There should then be some relationship between c-axis orientation and size.

3). There should also be some relationship between c-axis orientation and ellipticity.

Initial grain-shape measurements were made on seven samples on sections cut parallel to the lineation and perpendicular to the hand-specimen foliation (X-Z plane). The intensity of the shape fabric is shown in Figure 40. These are plots of ellipticity (R) versus the
Figure 40. $R_f/\phi$ plots of quartz grains from the Pinto quartzite
Figure 40. continued
angle between foliation (as defined by lenses of muscovite and aluminosilicates) and the long axis of the quartz grain (0). Samples CW 3, 5, 45, and 48 were selected for further study based on their strong shape-fabric intensity and their well developed c-axis fabrics. C-axis orientation and grain size, shape, and orientation measurements were made on approximately 50 grains from each sample. An effort was made to sample a variety of grain sizes and shapes.

Figure 41 shows the relationship of grain size (equivalent circular diameter) to ellipticity for the four samples. Because the plane of the thin section does not perfectly bisect any individual grain, the maximum cross section of the grains is not generally encountered. This may account for much of the scatter in grain size. Based on only a few samples, Behrmann (1985) found a tendency for larger grains to be more elliptical in a study of finer-grained, greenschist-facies rocks. A similar relationship is suggested by samples CW 5 and CW 45. Sample CW 3 shows evidence of secondary grain growth which has produced the unusually large grains evident in Figure 41.

Upon scanning thin sections cut parallel to the XZ plane, it was noted that most of the ribboned grains have their c-axes oriented within about 20° of the Y direction. If these ribbons are older grains that have undergone the most strain, perhaps this is the orientation to which the c-axes are rotating toward during intracrystalline glide. In order to test for any relation between c-axis orientation and grain size, the angle between the c-axis and the foliation pole (Z) was calculated. I have called this angle "gamma" in the following discussion. Size versus
Figure 41. Size (equivalent circular diameter) vs. Ellipticity (R) plot of quartz grains from four samples of Pinto quartzite.
plots of all four samples (Figure 42) show little evidence of a relationship between grain diameter and crystallographic orientation. This is a result of two factors. First, as mentioned above, the thin section takes a random slice through the grains measured. This will tend to result in artificially small size measurements. Second, even with strong host control over neoblast orientation, new grains are generated over a range of orientations (Ransom, 1971) and start rotating from different orientations. This is more evident in the next figure.

Ellipticity versus gamma plots for the four samples are shown in Figure 43. The ellipticity of a grain should be more or less insensitive to where the section cuts the grain, so it is probably a better indicator of grain evolution than area measurements. The least elliptical grains have gamma values from near 0° to about 60°. This suggests that new grains do not generally form with their c-axes at low angles to the foliation, but that they evolve in that direction as they are strained. The highest ellipticities measured were around six in grains with gamma values greater than 60°. This may represent the maximum strain that the crystal can undergo before it recrystallizes and erases all evidence of the previous strain. This relationship implies that as quartz grains grow and deform, their c-axes rotate toward parallelism with the Y axis of strain, resulting in a progressive rotation of the prism \(<a>\) glides into their easy glide orientation. In this orientation, at least one of the prism planes will be oriented nearly parallel with the shear plane and the \(<a>\) direction parallel to
Figure 42. Size (equivalent circular diameter) vs. gamma (angle between c-axis and foliation pole) plot of quartz grains from four samples of Pinto quartzite.
Figure 43. Ellipticity (R) vs. gamma (angle between c-axis and 112 foliation pole) plot of quartz grains from four samples of Pinto quartzite.
the shear direction (Schmid and Casey, 1986). It is interesting to note that although samples CW 3, 5, and 45 have c-axis fabrics that imply activity on the prisms, CW 48 does not (see Figure 39); yet it still follows the same trends on Figures 41 and 43.
In the other two chapters, the final figure has been a geological cross section showing the significant structural relations. This is extremely difficult in the Chuckwalla Mountains/Orocopia Mountains area because almost all of the exposed rocks are granitic orthogneisses and plutons which tend not to show consistent stratigraphic relationships. This inhibits interpretation beyond the surface exposures and limits the effectiveness of methods such as down-structure projections (Mackin, 1950). It is particularly difficult in the Red Cloud Canyon area because the foliation is so steeply dipping. It is presumed that the Early and Late thrusts exposed in Red Cloud Canyon flatten with depth, but the direction is not clear if Powell's antiform is missing. South of the Ship Creek fault, the Early thrust dips westward and overlies a thin lens of quartzite (Figure 33). Powell interpreted this to have regional significance, but these relations are found south of the Ship Creek fault near its western termination. This has clearly affected the strike of the fault in this area, so the dip is suspect as well.

One disturbing feature is the lack of strain preserved in the rocks along the Early Red Cloud thrust. The only thing differentiating the thrust contact from a primary lithologic contact is a decrease in size and number of microcline porphyroblasts in the Joshua Tree granite gneiss near the contact. There is no strain preserved in the microtextures of samples from along the contact due to the high temperatures associated with this deformation. This allows the
possibility of unrecognized "thrusts" being present in the Chuckwalla Mountains. Even if these have been obliterated by the granitic gneiss suite, major structural discontinuities are not likely to be present in the western Chuckwalla Mountains because of the persistence of Soledad gneiss blocks within the granitic gneiss across the area. The absence of these blocks in the eastern Orocopia Mountains may argue for a shear zone beneath the intervening valley fill, but they could simply be absent because of the distance to a source of the blocks. The granitic gneisses are injected into metasedimentary rocks in the Orocopia Mountains at the eastern limit of prebatholithic exposure. It is possible, then, that the granitic gneisses have intruded more-or-less along the contact between the metasediments and the Soledad augen gneiss. The consistent northeast dipping foliation in the granitic gneiss complex with associated west-vergent folds is interpreted be the product of distributed shear across the complex.

The Antiform Problem

A regional antiformal structure is strongly implied in the southern Eagle Mountains, where the east-dipping Pinto quartzite in western Big Wash correlates with west-dipping Pinto quartzite in upper Cholla Wash. There is a problem with Powell's proposed north-plunging antiform in the Chuckwalla Mountains east of Red Cloud Canyon. Mapping in the area between the two exposures of Joshua Tree granite gneiss on Figure 20 has failed to define the hinge of such a structure. The orthogneisses exposed there seem to have a continuous, steep to moderately northeast
dipping foliation, and the Late Red Cloud thrust cannot be traced around a closure. If the antiform is not present east of Red Cloud Canyon, then the two contacts mapped as the Late Red Cloud thrust on Figure 20 may in fact be two different structures.
RIGHT-LATERAL MYLONITES

A set of right-lateral, north-northwest trending mylonite zones cuts across the Eastern Transverse Ranges. With one exception, the zones are generally less than a meter or two across. The mylonites are dark, gray to greenish quartz-rich ultramylonites. The most quartz-rich layers are poorly lineated and flinty. Shear sense was determined by the geometries of shear bands and from asymmetric pressure shadows. In no case is the absolute displacement known.

The mylonites contain greenschist facies assemblages of:

quartz + biotite + muscovite (sericite) + chlorite.

They also contain porphyroclasts of altered feldspars and opaques. No garnet or amphibole was observed. Quartz tends to form ribbons of small, polygonal grains (type 2a ribbons of Boullier and Bouchez, 1978). Even the most siliceous samples contain more than 20 percent micas in the matrix. This may explain the small quartz grain size.

Their age is poorly constrained. They clearly post-date the Early Red Cloud thrust and apparently the Late Red Cloud thrust as well. They cut a correlative of Triassic Mount Lowe Intrusion in the Little Chuckwalla Mountains but are cut by granodiorites of inferred Cretaceous age in the eastern Chuckwalla Mountains. The relationship between the mylonites and the Jurassic plutons is unknown.

The widest shear zone of the group is shown on Powell's (1981) geologic map, as well as in Powell et al., 1984. The southernmost exposure of this shear zone in the Chuckwalla Mountains is just west of
Aztec mines along the southeastern edge of the range (Figure 20). The shear zone as Powell mapped it is actually an anastomosing group of shear zones. To the northwest the various shear zones appear to coalesce into a single zone. The foliation is near vertical and the lineation plunges gently to moderately northward. The mylonite continues northward to Ship Creek, where just north of its offset on the Ship Creek fault it is cut by a pluton of Cretaceous (?) granodiorite. Shear indicators near Ship Creek indicate right-lateral shear sense.

The shear zone separates a western block composed of the 1200 Ma syenite-mangerite suite from an eastern block of granitic gneiss and syenite. Just east of Graham Pass in the Little Chuckwalla Mountains, another large right-lateral shear zone separates a western block of syenite-mangerite from an eastern block of Mount Lowe related diorite (?) and granitic gneiss. Powell (1981) proposed a left-lateral offset of 8 km on the intervening Aztec Mines Wash fault, based on an offset plutonic contact. If the shear zones also correlate across the Aztec Mines Wash fault, they indicate an offset of approximately 8-9 km on the fault, depending on the strike used to project beneath the alluvium.

Another probable example of this set of shears is shown on the map of Cholla Wash (Figure 37). There, a shear zone separates moderately dipping quartzite from steeply dipping Joshua Tree granite gneiss. In the upper portion of Cholla Wash, near its intersection with the Red Cloud thrust, the shear zone is vertical. Farther south the shear zone dips to the northeast. Trains of folds (Figure 35c) within the shear zone indicate right-lateral motion.
Smaller shear zones, a few centimeters across are found in the Joshua Tree granite gneiss near the Red Cloud thrust (Figure 35d). The sense of motion is uncertain on these shear zones, but a near horizontal lineation is present. Similar small shear zones are present in the Joshua Tree granite gneiss in Red Cloud Canyon above the Red Cloud Mine. Other high-angle shear zones were found east of the Late Red Cloud thrust near Red Cloud Canyon, and west of Gulliday Wash in the southern Chuckwalla Mountains. Near Gulliday Wash a 1-2 m shear zone striking just east of north was traced for 2 km. At one location left-lateral shear indicators were found. Outside of the Chuckwalla Mountains, a right-lateral shear (N7W 90) was found between a body of Soledad augen gneiss and granitic gneiss south of the Golden Bee Mine in the northern Hexie Mountains.

These shears obviously postdate the peak ductile deformations because of their metamorphic grade and planar nature. They roughly parallel the foliation in gneisses so finding offset markers is difficult. In addition, since their displacements are unknown, it is difficult to determine their effect on the distribution of the prebatholithic rocks in the Eastern Transverse Ranges.
DISCUSSION

The results of this investigation, based on detailed study of selected areas do not lend themselves to a single summary of events. The interrelationship of the different areas is still not clear, so specific problems or points will be addressed individually.

Age of the Red Cloud Thrust Deformation

At the time this project was undertaken, the Red Cloud thrust deformation was known to have occurred between 1195 Ma and 160 Ma based on cross-cutting relations with dated plutons, but was suspected to be Precambrian.

The oldest undeformed igneous rocks in the Eastern Transverse Ranges are belong to Powell's Jurassic plutonic suite. They are probable correlatives of Tosdal's (1986) Kitt Peak-Trigo Peak Supergroup. The oldest dates on these plutons cluster around 165 Ma.

A foliated porphyritic quartz monzonite at Twentynine Palms Mountain in the Pinto Mountains gives a amphibole Rb/Sr age of approximately 210 Ma and a tentative U/Pb age of 220 Ma (Keith Howard, U.S.G.S., Menlo Park, CA, personal communication, 1986; 1988). Similarities between quartz monzonite and the 215 Ma porphyritic monzogranite of Frizzell et al. (1986) also suggest a late Triassic age for the Twentynine Palms quartz monzonite. This unit is well foliated on Twentynine Palms Mountain and possesses a west-northwest lineation. These are apparently Red Cloud thrust-related structures, and the tentative Late Triassic age
may date the Red Cloud thrust deformation. However Bob Powell (U.S.G.S., Spokane, WA, personal communication, 1988) suggested that much of the fabric in the Twentynine Palms pluton may be protoclastic.

Tosdal (1986, 1988) demonstrated Triassic syn-plutonic deformation in the Mule Mountains, which lie just east of the Little Chuckwalla Mountains. This deformation, which is associated with Mount Lowe-related rocks, has an E-W lineation, similar to that in the western Chuckwalla Mountains. As discussed in the Chuckwalla Mountains section, if the granitic gneisses are correlable with granitic rocks in the Little Chuckwalla Mountains, then a Triassic or Early Jurassic age is implied for the Red Cloud thrust deformation.

**Conditions of the Red Cloud Thrust Deformation**

Amphibole geobarometry indicates that the intrusives in the Mule Mountains correlative of the Mount Lowe Intrusion were emplaced at depths of approximately 15 km (Barth et al., 1988). Powell (1981) estimated the peak metamorphism of the Joshua Tree metasediments to have occurred at pressures of 3.5-4 kb, which implies similar depths.

Powell (1981) also estimated the peak metamorphic temperatures reached in the pelitic units to be 500 to 550°C. Even the essentially undeformed conglomerates (Figure 9c) above the undeformed Joshua Tree granite in the Pinto Mountains possess high-temperature mineral assemblages. Powell also argued that both the Early and Late Red Cloud thrusts postdate the peak conditions. The Early thrust was clearly active under high temperature. Temperatures were high enough during and
after the Early thrust to allow the recovery of all strain-induced microstructures. The Late Red Cloud thrust was active at lower temperatures based on the quartz microtextures.

Regional Structural Relations

Significance of the granitic gneiss suite

The presence of pegmatites similar to the granitic gneiss along the Early thrust in the Chuckwalla Mountains and in eastern Big Wash suggests an association between the granitic gneisses and major shear zones. The granitic gneisses are found in meta-intrusive contact in units of both the San Gabriel and Joshua Tree terranes. Powell (1986) suggested that the regional Red Cloud thrust deformation involved decollement along the tops of plutons. Leucocratic granitic gneisses believed to be correlative with the granitic gneiss complex crop out in a belt extending northward from the Chuckwalla and Orocopia Mountains through the eastern and northern Hexie Mountains (they are well exposed in the wash below Hidden Tank and Stirrup Tank) and southwestern Pinto Mountains. Perhaps the Chuckwalla/Pinto Mountain belt of granitic gneiss is an important boundary. Most of the pelitic gneisses and granofeloses lie west of this belt. Powell (1981) correlated the pelitic units in the eastern domain of the Hexie Mountains section with the pelitic lithosome of his Hexie paragneiss in Red Cloud Canyon. They differ greatly in texture, but this may be a product of differing deformation intensities. It is unclear how many separate pelitic
gneisses there are. The only certain correlative across the granitic gneiss belt is the Soledad augen gneiss which is exposed over a large area of the central Hexie Mountains, as well as in the Western Transverse Ranges.

In Powell's original mapping scheme, most orthogneisses without augen were lumped into his Augustine gneiss unit, which were in part believed to be retrograded granulites of Proterozoic age. The possibility that some portions, (perhaps most?) of this heterogeneous suite are probably Early Mesozoic in age indicates that these gneisses can be divided further with detailed study. Geochronological work would obviously be a great help in securely dating the deformation as well as in differentiating syn- and pre-TECTONIC units within the Augustine gneiss. Geochemical studies may aid correlation as well.

The Late Red Cloud thrust

The two best exposures of the Late Red Cloud thrust fault are in the Chuckwalla Mountains and in the vicinity of Pinto Mountain. In the Chuckwalla Mountains the Late Red Cloud thrust separates Joshua Tree granite gneiss from interlayered granitic gneisses and Soledad augen gneiss. There is a mylonite developed along the contact, and the Joshua Tree granite gneiss appears unusually fine grained and lineated several tens of meters away from the thrust. The thrust is fairly straight, except for the fold in the northeast corner of Figure 26.

In the Pinto Mountains the thrust is largely within the Pinto quartzite, although a lens of Joshua Tree granite gneiss is thrust above
quartzite along a portion of the contact. The fault has a highly irregular trace, and is easily missed. No mylonites were found along the contact, and pebbles 10-20 m into the lower plate are scarcely deformed. The fault appears to be a brittle structure. The whole footwall shows little deformation, and the deformation of the hanging is generally modest. If the two surfaces exposed in the Pinto and Chuckwalla Mountains are part of the same structure, it has very different characteristics in these two places.

Evidence for pre-Red Cloud Thrust Deformations

The north-northwest lineation in the Eastern Domain in the southwest Hexie Mountains is admitted to be of uncertain origin. In Washington Wash, in the central Hexie Mountains, there is broad spread of lineation attitudes in generally west dipping units. The lineations, some of which are clearly stretching lineations, range about N70W TO N30W, athwart the general east-west to northeast-southwest attitudes in the rest of the Eastern Transverse Ranges. This, combined with the foliated biotite xenoliths in the undeformed Soledad augen gneiss in the Hexie Mountains may indicate that vestiges of an older (Proterozoic?) deformations are preserved there.

Implications of the Right-Lateral Mylonites

Geological differences between lithologies exposed in the Eastern and Western Transverse Ranges and the rest of the southwestern United States have been commented on by several authors (Silver, 1971; Powell,
argued for a major left-lateral wrench fault, the Mojave-Sonora Megashear, to cut northwesterly across southern California in the general vicinity of the Eastern Transverse Ranges. Based on correlations of Proterozoic and Paleozoic rocks between the Inyo Mountains in eastern California and the Caborca area of Mexico, the Megashear is thought to have up to 800 km of offset. Anderson and Schmidt (1983) suggest that the movement on the Megashear was related to the opening of the Gulf of Mexico in Jurassic time, and is necessary for plate reconstructions. Tracing the Megashear into southern California has not be done with a great deal of success due to the limited outcrop of pre-Cretaceous age units.

Sm/Nd isotopes suggest that the basement rocks of the Eastern Transverse Ranges are related to the southwest North American craton (Bennett and DePaolo, 1986) and are thus neither far-traveled terranes nor probably transform-faulted great distances. Wright et al. (1987) determined Proterozoic plutonic events in the eastern Mojave that strongly correlate with emplacement ages of orthogneiss protoliths in the Transverse Ranges. Stone and Stevens (1987) argued that Paleozoic sediments in the Caborca area may be autochthonous.

Tosdal (1988) proposed a wrench fault with uncertain shear sense and an offset of less than a few hundred kilometers northeast of the Eastern Transverse Ranges. This fault would have been active after the Red Cloud thrust event and before the Jurassic plutonism which pins the rocks to North America.
Because of the region-wide distribution of the right-lateral shear zones in the Eastern Transverse Ranges, it is tempting to correlate them with some regional strike-slip event. Their orientation and general age make them possible correlatives of the Mojave-Sonora Megashear, but they record right-lateral shear (generally), as opposed to the left-lateral movement proposed for the Megashear. Perhaps these shear zones are related to some as yet undiscovered strike-slip system.
CONCLUSIONS

1. In the southern Eastern Transverse Ranges, the Red Cloud thrust deformation involved regional west-directed shear. This is inferred from asymmetric folds in the central domain of the southwest Hexie Mountains and in the western Chuckwalla Mountains. In the Hexie Mountains the west-vergent folds are inferred to be related to the mylonitic foliation because of the coaxiality of the folds and the boudinage structures in the central domain. In the Chuckwalla Mountains, the west-vergent F2 folds are the dominant fold set. Although they fold the foliation, they are thought to have developed roughly simultaneously with it, because of their geometric relationship to the lineation.

C-axis quartz fabrics in Cholla Wash and eastern Big Wash in the Eagle Mountains indicate a west-directed shear during the Red Cloud thrust deformation. The Cholla Wash samples preserve optical evidence of dynamic recrystallization, whereas the eastern Big Wash samples have textures indicative of secondary recrystallization (exaggerated grain growth). Western Big Wash data are ambiguous.

2. The granitic gneiss suite of the western Chuckwalla Mountains is interpreted to be the result of syn-kinematic intrusion of material derived from melting of crustal material. These same gneisses are found in the Eagle, Hexie, and Pinto Mountains and may be associated with shear zones.
3. The deformation in the western Chuckwalla Mountains is tentatively assigned a Middle to Late Triassic age based on correlation of granitic gneisses with similar gneisses in the Little Chuckwalla Mountains which intrude a Mount Lowe Intrusion correlative. The Mount Lowe intrusions are dated at 210-230 Ma. Preliminary dating in the Pinto Mountains also support a Triassic-Early Jurassic age.

4. North-northwest striking, generally right-lateral mylonite zones cut the prebatholithic units in the Little Chuckwalla Mountains, the Chuckwalla Mountains, the southern Eagle Mountains and in the central Hexie Mountains. They are generally planar, and have greenschist facies assemblages. They are cut by Cretaceous (?) granodiorites, but were not found in contact with Jurassic plutons. Offsets were not determined on any shear zones.

5. The west limb of the Big Wash synform shows signs of reactivation in a top-to-the-east sense. This apparently occurred after 148 Ma.
REFERENCES


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