GEOLOGY AND GEOCHRONOLOGY

OF THE

AVAWATZ MOUNTAINS, SAN BERNARDINO

COUNTY, CALIFORNIA

by

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ABSTRACT

Geologic field mapping, and potassium-argon and rubidium-strontium geochronologic studies in the Avawatz Mountains, southeastern California, indicates that this range is primarily a single large fault block of Mesozoic plutonic rocks, flanked to the southwest and northeast by the Ceonzoic Arrastre Spring and Mule Spring faults, respectively. Several roof pendants of pre-Mesozoic metasedimentary rocks occur within the plutonic complex.

Several moderate size and many small pendants of Paleozoic and Precambrian metasedimentary rocks occur within the Mesozoic intrusive complex. Some of the pendants contain sequences of Precambrian metasedimentary rocks that are not easily correlated with any known formations in the area. Two of the largest pendants in the southern half of the range together contain an almost complete stratigraphic sequence extending from the upper part of the upper Precambrian Johnnie Formation to the Upper Devonian Crystal Pass member of the Sultan Limestone. Field mapping and section measurement indicates that this sequence is characterized by: 1) a moderately well-developed sequence of upper Precambrian to lower Cambrian terrigenous rocks typical of other southwestern Great Basin miogeoclinal sequences, and 2) a thin, platformal carbonate sequence in which the Middle and Upper Devonian Sultan Limestone rests disconformably on the upper Cambrian Nopah Formation. Determination of the stratigraphic sequence in these rocks adds a significant new control point for the location of isopach lines in the southern Cordilleran miogeocline. The zero isopach of all Ordovician and Silurian formations must lie north of the Avawatz Mountains, and these isopachs project southwestward directly toward Paleozoic eugeoclinal rocks in the northwestern Mojave region. This adds significant support to the hypothesis that the western Mojave eugeoclinal rocks have been tectonically juxtaposed against the truncated margin of the southern Cordilleran miogeocline in Permian to Triassic time.

Most of the Avawatz Mountains are underlain by Mesozoic plutonic rocks. The Avawatz quartz monzodiorite complex is by far the most abundant rock type, and is exposed throughout the range. This plutonic body was not entirely homogenized with respect to strontium isotopes at the time of emplacement. It intrudes older leucocratic granite, and is intruded by a granodiorite stock. A body of quartz monzodiorite in the northwestern part of the range contains less than 10% quartz and has a modal mineral composition similar to alkalic plutons at several other localities in the

southern Cordilleran orogen. Folding, boudinage, and foliation development in pre-Mesozoic metasedimentary rocks occurred prior to emplacement of the Avawatz quartz monzodiorite complex. In addition, the quartz monzodiorite is penetratively deformed in the northwestern part of the range: Potassium-argon and rubidium-strontium dating, and regional stratigraphic relationships indicate that the quartz monzodiorite complex is 200 + 20 m.y. old. Potassium-argon dating also indicates that plutonic rocks in the Avawatz Mountains have undergone a long and complex thermal history probably characterized by multiple heating events related to magmatism, and cooling events related to uplift during Mesozoic and Cenozoic time.

The Cenozoic Arrastre Spring fault juxtaposes Tertiary sedimentary rocks and Mesozoic volcanic and plutonic rocks to the southwest with deeply eroded Mesozoic plutonic and pre-Mesozoic metasedimentary rocks to the northeast. Distinctive, pink, medium- to coarse-grained granite crops out on both sides of the fault and limits total lateral offset to less than about 10km. The Tertiary Avawatz Formation is a thick sequence of Tertiary clastic sedimentary rock deposited across topographic relief produced by southwest-side-down normal movement on the Arrastre Spring fault. Potassium-argon dates of an ash bed within the Avawatz Formation indicate that the lower part of the formation, and most of the movement on the Arrastre Spring fault, occurred in Eocene to Paleocene time. Because the northern Arrastre Spring fault curves westward to merge with the Garlock fault, it is interpreted as a Garlock related structure. This new evidence supports the hypothesis that early Tertiary movement occurred on a proto-Garlock fault.

Quartz monzodiorite occurs on both sides of the Mormon Spring fault (new name), limiting its offset to less than about three kilometers in a left-lateral sense and about 20km in a right-lateral sense. Based on offset constraints on the Mormon Spring fault, and geometric considerations, direct correlation of the Mormon Spring fault with the Mule Spring fault is rejected. Distinctive quartz monzonite occurs on both sides of the Leach Lake fault, limiting its offset to less than a few kilometers.

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"The form of the object studied is essential to understanding its
essence."

                            Robert A. Schumann
                            19th century German composer and
                            music critic

You're going to have to rewrite all the geology books about this area.
You see these mountains around here? They're all meteorites! That's
right, every one of them. They came down in Biblical times. First
the Avawatz came in, then the Panamints came in from the east...

                            John Loscott, 1979

INTRODUCTION

The research described in this dissertation is based primarily on a field geologic investigation of the Avawatz Mountains, northeastern Mojave Desert region, California (see Fig. 1 for location). The results of this study are improved understanding of the geologic evolution of the Avawatz Mountains and its relavence to the regional development of the southern Great Basin and Mojave Desert region. This research involved seven months of geologic field mapping and included K-Ar and Rb-Sr isotope geochronological investigations. The most important contributions of this research are determination of: 1) detailed stratigraphy and regional relationships of a late Precambrian to Devonian age sequence of metasedimentary rocks, 2) the distribution, petrography, and sequence of Mesozoic intrusive rocks, with some constraints on their absolute age and cooling history, and 3) the age and nature of the Cenozoic Avawatz Formation and Arrastre Spring fault.

Field mapping was done during 1979, 1980, and 1981 on GS-CW series aerial photographs (1:25,000 approximate scale). Field data was transferred to a topographic base map (1:15,000 scale) made from a photographic enlargement of parts of four USGS 1:62,500 scale topographic maps (Avawatz Pass 15', Red Pass Lake 15', Baker 15', Silurian Hills 15'). The resultant geologic map is enclosed as Plate 1A. A map key including indentification of map units and structural symbols, and cross sections, is enclosed as Plate 1B. Plate 2 is a compilation geologic map (1:62,500 scale) of the Avawatz Mountains and immediately surrounding areas, and includes recently completed mapping of the northeast flank of the Avawatz Mountains by Troxel and Butler (1979).

The geologic map (Plate 1A) represents a body of geologic data, with interpretive aspects minimized or eliminated as much as possible. All geologic features were located on the map with exceptional care, and the greatest uncertainty in location of geologic features is probably the result of imprecise alignment of the topographic base map and geologic map during map reproduction. This is a systematic error, however, and can be estimated by observation of the magnitude of

Fig. 1. Schematic geologic map of Mojave-southwestern Great Basin region showing location of study area. Major Cenozoic faults are shown by heavy lines. SAF = San Andreas fault; GF = Garlock fault; SDVF = southern Death Valley fault zone; FCF = Furnace Creek fault. All Paleozoic rocks are outlined. Paleozoic eugeoclinal rocks are designated by horizontal lined pattern. SB = San Bernardino Mountain; SV = Sidewinder Mountain, Victorville area; EP = El Paso Mountains; SM = Soda Mountains; HH = Halloran Hills; SH = Silurian Hills; P = Providence Mountains; C = Clark Mountains; S = Spring Mountains; LV = Las Vegas; F = Frenchman Mountain.

Fig. 1.

mis-alignment of registration marks. The geologic map forms the primary data base for the interpretations and conclusions made in this study.

This dissertation is divided into three sections, with each section representing a coherent body of research, independent of the other sections except all are based on the enclosed geologic maps. This organization is intended to facilitate publication of this dissertation as three separate reports plus a geologic map.

ACKNOWLEDGEMENTS

PALEOZOIC AND PRECAMBRIAN ROCKS

INTRODUCTION

Regional Setting

The Cordilleran miogeocline is a broad, northwestward thickening
wedge of sedimentary rocks deposited on a stable continental margin
during late Precambrian and Paleozoic time. Isopach and facies lines
generally trend in a south-southwesterly direction through western
Utah and eastern and southern Nevada to the Mojave Desert region of
California where pre-Mesozoic rocks are largely obliterated by the
plutons of the south-southeasterly trending Mesozoic batholith belt
(Fig. 1). Within the Mojave region, miogeoclinal rocks are preserved
only as small, widely spaced, metamorphosed and internally deformed
pendants. Continuity of the miogeocline across the Mojave region was
first demonstrated by Stewart and Poole (1975), Tyler (1975) and Miller
(1977).

The Cordilleran miogeocline is composed of upper Precambrian to
Lower Cambrian terrigenous rocks and Paleozoic carbonate rocks. West-
ward thickening occurs by three mechanisms: 1) by addition of a thick
sequence of upper Precambrian terrigenous rocks, 2) by westward thicken-
ing of all units, and 3) by progressive addition of Ordovician, Silurian,
and Early Devonian units beneath a regional, sub-Middle Devonian uncon-
formity (Burchfiel and Davis, 1981). The platformal or cratonal sequence
east of the miogeocline is exposed at Frenchman Mountain east of Las
Vegas, where a thin sequence of Lower Cambrian terrigenous rocks rest
directly on older, Precambrian crystalline basement, and within the
Paleozoic rocks, Upper Devonian carbonates rest directly on Upper
Cambrian carbonates (see Fig. 2 for location). These two features are
characteristic of platformal or cratonal sequences near the transition
zone with the Cordilleran miogeocline. Progressive westward thickening
of the miogeocline by the mechanisms discussed above is well documented
in the Spring Mountains west of Las Vegas (Burchfiel and others, 1974).

Miogeoclinal sequences in the Mojave region are somewhat anomalous
in that relatively cratonal Paleozoic sequences, with Middle or Upper
Devonian carbonates resting disconformably on Upper Cambrian carbonates,

Figure 1. Schematic tectonic map of California and Nevada during Late Triassic to Middle Jurassic time (modified from Schweickert, 1976, and Burchfiel and Davis, 1972).

Fig. 2. Schematic geologic map of Mojave-southwestern Great Basin region showing location of study area. Major Cenozoic faults are shown by heavy lines. SAF = San Andreas fault; GF = Garlock fault; SDVF = southern Death Valley fault zone; FCF = Furnace Creek fault. All Paleozoic rocks are outlined. Paleozoic eugeoclinal rocks are designated by horizontal lined pattern. SB = San Bernardino Mountain; SV = Sidewinder Mountain, Victorville area; EP = El Paso Mountains; SM = Soda Mountains; HH = Halloran Hills; SH = Silurian Hills; P = Providence Mountains; C = Clark Mountains; S = Spring Mountains; LV = Las Vegas; F = Frenchman Mountain.

Fig. 2.

overlie moderately thick miogeoclinal sequences of late Precambrian to Lower Cambrian terrigenous rocks. Some of the isopachs of upper Precambrian-Lower Cambrian terrigenous rocks trend north-south through the eastern Mojave region whereas, in contrast, rapid lateral thickening of Paleozoic carbonate sequences occurs along an east-west axis through the northern Mojave-southern Great Basin region (Burchfiel and Davis, 1981). The reason for this divergence of isopach and facies trends is not known, but apparently the subsidence rate of the Mojave region decreased greatly in Cambrian time.

Over much of the eastern Great Basin and Mojave region, the Cordilleran miogeocline rests on older Precambrian crystalline rocks. In the southern Death Valley region, however, a thick sequence of Precambrian metasedimentary rocks known as the Pahrump Group rests on older Precambrian crystalline basement (Hewett, 1956; Wright and others, 1976). Strata of the Cordilleran miogeocline rest with moderate angular unconformity on the Pahrump Group in all areas except the Silurian Hills located about ten miles east of the Avawatz Mountains (Wright and Troxel, 1967; Kupfer, 1960). In the Silurian Hills, the Pahrump Group is conformable or disconformable beneath miogeoclinal rocks, suggesting that it is not much older. However, lower parts of the Pahrump Group are intruded by diabase that is lithologically similar to diabase dikes in Arizona dated 1.15 to 1.20 b.y. (Silver, 1960; Wrucke and Shride, 1972).

Precambrian crystalline rocks are exposed at a few scattered localities in the Mojave and Death Valley regions. Radiometric age determinations generally give ages ranging from 1400 to 1800 million years (Wasserberg and others, 1959; Lanphere and others, 1963; Lanphere, 1964). These high grade metamorphic rocks form the basement upon which the Pahrump Group and younger rocks were deposited.

Previous Geologic Studies

Precambrian and Paleozoic sedimentary rocks are well exposed east of the Avawatz Mountains in the southern Great Basin where they have not been obliterated by Mesozoic plutonism. Not surprisingly, these are some of the areas where strata of the southern Cordilleran miogeocline were first studied and described: studies by Nolan (1929), Hazzard (1937), and Hewett (1931, 1956), among others, were instrumental in

determining the basic stratigraphy and distribution of the miogeocline
and older strata of the Pahrump Group. More recent studies of upper
Precambrian to Lower Cambrian strata have led to a better understanding
of patterns of sedimentation in the Death Valley region (Wright and
Troxel, 1966; Stewart, 1970). It has not been possible, at least un-
til very recently, to extend isopach and facies lines across the Ava-
watz Mountains and across the Mesozoic granitic terrane of the Mojave
region. Stewart and Poole (1975) and Tyler (1975) recognized rocks
in the Victorville-San Bernardino Mountains area of the western
Mojave region that are correlative with upper Precambrian-Lower
Cambrian miogeoclinal sequences in the southern Great Basin. However,
constraints on the configuration of isopach and facies lines across
the Mojave region are poor because of the large distances between con-
trol points.

Our knowledge of the regional stratigraphy of the Paleozoic,
largely carbonate strata of the southwesternmost part of the Cordilleran
miogeocline has advanced greatly during the past ten years as a result
of studies by B. C. Burchfiel and students at Rice University. Gans
(1974) determined the internal stratigraphy of the Goodsprings Dolomite,
a Cambrian to Devonian carbonate sequence mapped over large areas of
the Spring Mountains and Clark Mountains by Hewett (1931, 1956). De-
scriptions of Paleozoic carbonate sections in the New York Mountains
(Burchfiel and Davis, 1977), Cowhole Mountains (Novitsky-Evans, 1978)
and Sidewinder Mountain (Miller, 1977) demonstrate continuity of a
Paleozoic platformal sequence across the Mojave region.

Past geologic studies in and adjacent to the Avawatz Mountains
described a number of Precambrian to Lower Cambrian sections, but no
Paleozoic carbonate sequences have been reported. Reconnaissance studies
by Jahns and others (1971) first located the central pendant in the
Avawatz Mountains. They identified stratigraphic units extending from
the Johnnie Formation to the Bonanza King Formation, but did not
report on younger parts of the Paleozoic section. Wright (1968) studied
a small area of well exposed Pahrump Group at Sheep Creek in the north-
ern Avawatz Mountains. Scattered fault blocks of Pahrump Group strata,
including one mapped by Wright (1968), were mapped by Troxel and
Butler (1979) north of the Mule Spring fault on the northeast flank of
the range. Geologic field studies of the adjacent southern Salt Spring

Hills (Troxel, 1967) and Silurian Hills (Kupfer, 1960) resolved
stratigraphic details in the Pahrump Group and upper Precambrian-Lower
Cambrian terrigenous strata, but did not describe the stratigraphy
of Paleozoic carbonate rocks in the area.

Study Area

Paleozoic and Precambrian rocks in the Avawatz Mountains are
present in two areas: 1) as fault blocks northeast of the Mule Spring
fault, and 2) as pendants and inclusions within the Mesozoic plutonic
complex which underlies most of the range (Figs. 3, 4, 5). This
study describes rocks from only the latter areas. All pre-Mesozoic
rocks in the Avawatz Mountains have undergone metamorphism and defor-
mation, and as a result, proper stratigraphic assignment of units is
difficult or impossible in some areas.

Two major pendants in the Avawatz Mountains, the central and
southern pendants, contain west dipping sequences of upper Precambrian
to Upper Devonian age metasedimentary rocks. The most complete se-
quence is present in the central pendant, and extends from the upper
part of the upper Precambrian Johnnie Formation to the Upper Devonian
Crystal Pass Member of the Sultan Limestone. Using exposures in both
pendants, an almost complete section across this stratigraphic interval
was located and measured (Appendix A). This rock sequence is similar
to other sequences described from the Mojave region; it consists of
moderately thick, miogeoclinal, upper Precambrian-Lower Cambrian ter-
rigenous strata that are overlain by relatively thin cratonal Paleo-
zoic strata in which Middle or Upper Devonian rocks rest disconformably
on Upper Cambrian rocks.

Older Precambrian metasedimentary and metaigneous(?) rocks are
exposed at scattered localities in the southern Avawatz Mountains. The
metasedimentary rocks are probably correlative with the lower part of
the Johnnie Formation, Noonday Dolomite, or Pahrump Group. Metasedi-
mentary rocks exposed at two localities may be even older. A Precambrian
migmatitic gneiss appears to intrude one of these metasedimentary bodies.

Upper Paleozoic metacarbonates are exposed as scattered, highly
irregular pendants and inclusions in the northwestern part of the range.

Fig. 3. Simplified geologic map of the Avawatz Mountains.
pС = Precambrian metasedimentary rock, includes upper Wood
Canyon Formation and Zabriskie Quartzite of Early Cambrian age.
Pz = Paleozoic metasedimentary. Mzi = Mesozoic intrusive.
Tav = Tertiary Avawatz Formation. Location of Figures 4 and 5
shown by dashed outline.

Fig. 4. Simplified geologic map of central pendant and surrounding areas. See text for unit identification and description. Circled "s" indicates stromatoporoid occurence.

Fig. 4.

Fig. 5. Simplified geologic map of south pendant and surrounding areas. See text for unit identification and description. Circled "s" indicates stromatoporoid occurence.

Fig. 5.

MAP UNITS

The following are descriptions of all pre-Mesozoic rock units in the part of the Avawatz Mountains within the mapped area (Plate 1A).

Gneiss (p€gn)

Gneiss of Precambrian age is present in only one area at the southeastern end of the Avawatz Mountains. Where best exposed along the edge of the adjacent wash, it is a lineated, fine-grained, biotite gneiss with medium-grained aplite dikes 1cm to 1m thick. Thin dikes (1cm thick) locally eminate from thicker (10-50cm) dikes, and the thinner dikes exhibit pervasive ptygmatic folding. The aplite is composed of about 50% K-spar, 25% plagioclase, and 25% quartz with sparse biotite and muscovite. Ptygmatic folds and aplite dikes are numerous within the gneiss, giving it a migmatitic appearance. Ptygmatic fold axial planes are co-planar with the foliation, and fold axes are co-linear with the lineation. However, the orientation of the lineation and foliation vary within a single outcrop by perhaps $20°-30°$. Unaltered gneiss, forming the southernmost portion of this body, is flanked to the north and west by a broad zone of fine-grained, epidote rich brown and green metamorphic rocks which are gradational with the gneiss.

In thin section, the gneiss is a fine-grained, interlocking network of anhedral quartz, oligoclase, orthoclase, biotite, and magnetite(?). Plagioclase is slightly sericitized, and quartz grains exhibit undulatory extinction. Biotite grains are usually oriented sub-parallel to foliation, but no other minerals show preferential orientation. Clots of magnetite surrounded by biotite are locally present in the biotitic gneiss.

Age and correlation: Two leucosome-melanosome sample pairs were collected from the gneiss and analyzed for rubidium and strontium content and isotope composition. Rocks within each pair were collected from sites 1m apart, while the two pairs were collected from sites about 100m apart. Sample locations are shown on Plate 1A, and the isotopic analyses are shown on figure 6. Analyses of one pair of samples define a two-point isochron of 1.387×10^9 years (Fig. 6,

Fig. 6. Rubidium-strontium isochron diagram for samples from Precambrian gneiss (p€gn). Samples 166A and 167A are biotite gneiss forming most of the rock body. Samples 166B and 167B are aplite dike rock. Two point isochron through 167A and 167B defines an age of 1.387×10^9 years and $(^{87}Sr/^{86}Sr)_0$ = .726 (assuming $\lambda^{87}Rb = 1.42 \times 10^{-11}$/yr.). Model isochron through 167A and $^{87}Sr/^{86}Sr_0$ = .704.

Precambrian gneiss

Sample*	Rb(ppm)	Sr(ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$** $\pm$ 2σ
166A M	198.858	95.247	6.12355	.847950 $\pm$ .000063
166B L	53.334	97.150	1.59615	.757867 $\pm$ .000053
167A M	138.333	34.922	11.5400	.779243 $\pm$ .000078
167B L	46.240	69.3615	1.943	.783583 $\pm$ .000055

*M = melanosome, L = leucosome
** Sr isotopic compositions normalized to $^{86}Sr/^{88}Sr$ = 0.1194, reported relative to E and A $SrCO_3$ = .70800.

Table 1: Analytical data for Precambrian gneiss.

Table 1). This age falls well within the range of ages determined for other Precambrian rocks in surrounding areas. The other pair of samples yielded a negative two-point isochron. The samples are not significantly altered, and the reason for this negative isochron is not clearly understood.

A fundamental assumption in Rb-Sr dating of leucosome-melanosome pairs is that both rock types were homogenized with respect to strontium isotopes at the time of crystallization of the leucosome. Field evidence from the Avawatz gneiss supports this assumption. The rock appears to have underwent distributed, in situ partial melting, followed by migration of the partial melt into veins and pods within the host rock and subsequent crystallization. The highly ductile nature of associated deformation, producing pervasive ptygmatic folds, lineations, and foliations, with minimal fabric development, suggests that this was a high temperature event, quite likely occurring synchronously with partial melting. Perhaps partial melting decreased the strength of the rock, resulting in the onset of deformation in a constant stress regime. In any case, the pervasive migmatitic and highy deformed character of the rock suggests that it was hot, very ductile, and contained some partial melt, all conditions favorable to strontium isotope homogenization.

A model isochron through 167A, assuming a very low $^{87}Sr/^{86}Sr$ initial ratio of .704 yields an age of 468 m.y.. There is no evidence for Paleozoic thermal or magmatic activity in the Mojave region, and the gneiss has yielded a K-Ar age (biotite) of 95 m.y.. Therefore, sample 167A must have exchanged strontium and/or rubidium with surrounding rocks of different composition during Mesozoic time. It may have had a composition similar to leucosome sample 166B until its rubidium content was increased six-fold by diffusion of rubidium during Mesozoic heating events.

A Precambrian age for the gneiss seems certain, especially considering the high $^{87}Sr/^{86}Sr$ ratios of all samples. The isotope systematics of this rock are incompletely understood, but some open system behavior must have occurred during Mesozoic time.

Metamorphic and igneous rocks forming the basement of the Pahrump

Group and younger Precambrian sediments in the southern Great Basin and eastern Mojave region generally fall in the age range 1400 to 1800 m.y.. Rubidium-strontium ages of 1570 to 1700 m.y. were obtained from muscovite separates from a pegmatite that cuts an older schist in the southern Panamint Mountains. Muscovite from the schist yielded an age of 1480 m.y. (Wasserburg and others, 1959). Uranium-lead and lead-lead ages of 1720 to 1780 m.y. were obtained from zircon separates from metavolcanics and granite gneisses in the southern Panamints (Silver and others, 1961). Potassium-argon and rubidium-strontium dating of biotite and feldspar from granitic rocks in the Marble Mountains yielded ages of about 1200 to 1400 m.y. (Lanphere, 1964). Rubidium-strontium analyses of the Avawatz gneiss indicates that its age probably falls in the range of other radiometric dates from surrounding areas.

Quartzite, undifferentiated (p€q)

Undifferentiated quartzite is present as faulted, folded, metamorphosed, and brecciated quartzite which crops out over small areas and can not be confidently correlated with any formation. It consists of white-, orange- or brown-weathering, thin- to thick-bedded, and medium- to fine-grained quartzite.

The largest exposed body of undifferentiated quartzite is east of the central pendant and below a Tertiary low-angle fault (Fig. 4). This body of quartzite is lithologically distinct from the other quartzites because it is composed primarily of white, massive, medium-grained orthoquartzite. In thin section, the quartzite is composed of medium- to fine-grained quartz crystals which are thoroughly intergrown and show no evidence of detrital origin such as overgrowths on rounded grains. Micaceous, black to gray quartzite and tan dolomite are interbedded with the white quartzite, and thicker (5-30m) dolomite beds are mapped separately as Precambrian dolomite (p€d).

Age and correlation: Many of these small outcrops of quartzite are probably correlative with lithologically similar quartzites in the Wood Canyon Formation, Stirling Quartzite, Johnnie Formation, and possibly the Noonday Dolomite, and Kingston Peak and Crystal Spring

Formations of the Pahrump Group (Stewart, 1970; Roberts, 1976; Williams and others, 1976).

The large body of predominantly white quartzite may be correlative with orthoquartzites in the Halloran Hills that are intruded by Precambrian plutons dated at 1400 to 1500 m.y. (DeWitt, 1980; Lauren Wright, personal communcation, 1981). Unlike the quartzites in the Halloran Hills, however, these quartzites contain dolomite beds. They warrant further study as their mature composition may indicate a period of tectonic quiescence between periods of Precambrian orogenesis.

Dolomite, may include Noonday Dolomite (p€d)

Precambrian dolomite occurs at several localities within and around the central pendant, and at most localities it is interbedded with or adjacent to quartzite. The dolomite is generally tan and internally brecciated, showing little or no bedding. Most of the dolomite and associated quartzite are lithologically similar to the Johnnie Formation, with which they are tentatively correlated.

At two localities, however, a correlation with the Johnnie Formation can not be made. The first locality is the large exposure of predominantly Precambrian quartzite (p€q) below a thrust fault at the east edge of the central pendant (Fig. 4). Dolomite is a minor constituent of these metasedimentary rocks, and both are older than the Johnnie Formation (see section on quartzite (p€q)).

The second locality is along a ridge at the northeastern edge of the central pendant where dolomite locally contains possible algal structures. These structures are defined by thin layers and laminations of protruding, slightly siliceous dolomite which form broad (1-3m) convex upward mounds perhaps 10-30cm high. The narrow (10-20cm) gap between mounds contain concave upward layers. Climbing ripples with 1-2cm amplitude also occur within the dolomite. This dolomite overlies black, phyllitic shale (p€q on Plate 1), and the two are tentatively correlated with the basin facies of the Noonday Dolomite (Williams and others, 1976), although they could be correlative with the Crystal Spring Formation (Roberts, 1976).

Quartzite, phyllite, dolomite and limestone, undifferentiated (p€u)

This map unit is restricted to outcrops in three small ridge top pendants just east of the central pendant that contain an east dipping sequence of metasedimentary rocks (Fig. 7). Evidence from cross-stratified beds suggest that the rocks are right-side up.

Age and correlation: This sequence is unlike the Johnnie Formation because it contains white quartzite (orthoquartzite?) beds and calcite marble beds, and therefore it is probably older. Both the Crystal Spring Formation and basinal facies of the Noonday Dolomite contain arkosic sandstone, limestone, and dolomite, but relatively pure, white quartzite is nowhere reported from these formations (Hewett, 1956; Roberts, 1976; Williams and others, 1976). Even so, I tentatively correlate these rocks with one of these two formations. Further work is needed to establish a basis for a more restrictive stratigraphic correlation.

Johnnie Formation (pЄj)

The Johnnie Formation is recognizable only in one location on the east edge of the central pendant (Fig. 4). At this location, the upper 111.4m of the Johnnie Formation is in a continuous stratigraphic sequence with the overlying Stirling Quartzite. Stratigraphically lower parts of the Johnnie Formation are faulted, folded, and brecciated, but are lithologically similar to the upper Johnnie. An east dipping sequence of lithologically similar metasedimentary rock 1-2 miles to the northwest was mapped as Johnnie Formation(?) on plate 1A, although these rocks could be older.

Lithology: The most characteristic feature of the Johnnie Formation is its lithologic diversity (Appendix A). It is composed primarily of fine- to coarse-grained, orange-, maroon-, or brown-weathering quartzite. Tan dolomite marble, greenish-gray to brown calc-silicate rocks, and gray phyllite beds also occur throughout the Johnnie. Tan dolomite beds up to 5-10m thick are most abundant in the basal part of the Johnnie Formation. The thickest of these beds, which extend for several hundred meters along strike, is mapped separately as Precambrian dolomite (pЄd). It was mapped because the dolomite

APPROXIMATE STRATIGRAPHIC COLUMN

Precambrian quartzite, phyllite, dolomite,

and limestone, undifferentiated (p€u)

Interbedded white, gray, and tan calcite marble.

Interbedded argillite, dark, fine grain micaceous quartzite, and sparse beds of medium grain white quartzite and white calcite marble.

Interbedded dolomite and calcite marble. Quartzite, fine grain, brown weathering.

Quartzite, medium to coarse grain, white, massive appearing.

Quartzite, fine grain, brown weathering.

Quartzite, medium to coarse grain, white, massive appearing.

Dolomite marble, tan.

Quartzite, very fine grain, micaceous, black-gray, faintly bedded with local cross beds.

Argillite, very fine grain, micaceous, brown weathering, breaks on bedding planes

Dolomite marble, tan.

100

METERS

0

Figure 7. Schematic stratigraphic column of metasediments designated p€u. Thicknesses are approximate.

bed occurs near the base of the Johnnie Formation, and it is pos-
sible that it is correlative with the Noonday Dolomite.

Age and correlation: Because the Johnnie Formation is Precambrian in
age and contains no fossils, stratigraphic correlations must be based
solely on lithologic criteria and stratigraphic position. The most
distinctive feature of rocks assigned to the Johnnie Formation is
their lithologic diversity. This is a characteristic feature in
the southern Death Valley-Kingston Range area (Stewart, 1970; Wright
and Troxel, 1966), and in areas to the northeast including the type
area in the northwest Spring Mountains (Nolan, 1929). The distinctive
Johnnie oolite, an oolitic dolomite bed generally 2-4m thick, occurs
within the upper Rainstorm Member of the Johnnie Formation throughout
much of the southern Death Valley-Kingston Range area and northward,
but was not located in the Avawatz Mountains. Failure to locate the
oolite is attributed to the strong metamorphic overprint and to the
thinness of the oolite in this area as indicated by exposures in the
Salt Spring Hills just northeast of the Avawatz Mountains (Troxel,
1967).

Correlation of rocks assigned to the Johnnie Formation in the
Avawatz Mountains with the Johnnie Formation in surrounding areas and
in its type area is based on lithologic similarity and on stratigra-
phic position. In the Avawatz Mountains the Johnnie Formation clearly
underlies the sequence Stirling Quartzite-Wood Canyon Formation-
Zabriskie Quartzite-Carrara Formation-Bonanza King Formation. The
base of the Johnnie Formation could not be defined due to extensive
deformation.

Stirling Quartzite (p€s)

Stirling Quartzite is well-exposed in the southern and central
pendants of the Avawatz Mountains (Figs. 4, 5). Metamorphism and
deformation have strongly affected the Stirling Quartzite in the
southern pendant, where it is typically bleached white and highly
folded, faulted, and brecciated. It was not possible to divide the
Stirling Quartzite into members in the southern pendant, but an almost
complete, intact section of Stirling Quartzite is exposed in the

central Avawatz Mountains (Fig. 4). The Stirling Quartzite in the central pendant can be divided into three members: a lower quartzose member (63.7 meters thick), a middle pelitic member (51.6 meters thick), and an upper quartzose member (108.8 meters minimum thickness). The 224.1m measured thickness of the Stirling Quartzite is a minimum figure, as the top of the third member is faulted, and a complete section is not present in the map area.

Lithology: The lower member of the Stirling Quartzite is composed of maroon, orange, brown, and white fine- to medium-grained quartzite. It varies from thin- to thick-bedded, and is moderately resistant. Distinctive features are its consistent quartzose character and dark maroon- to orange-weathering color. The base is placed at the top of the highest dolomite or calc-silicate bed in the Johnnie Formation.

The middle member is generally composed of black to dark orange, fine-grained, thinly bedded to laminated, variably argillaceous quartzite. Several tan-weathering dolomite beds occur within the member. The lower contact of the middle member is placed at the base of the lowest argillaceous quartzite bed greater than 1m thick above quartzites of the lower member.

The upper member is composed of fine- to coarse-grained, relatively pure quartzite. Color generally varies from white to reddish-brown or orange. Beds are typically 5 to 40cm thick, locally varying from 1m to 0.5cm thick. Low-angle cross-stratification and pebbly beds with well-rounded quartzite pebbles up to 1cm diameter are present locally. The base of the upper member is placed at the top of the highest dolomite, calc-silicate, or argillaceous quartzite bed greater than .5m thick in the upper part of the middle member. The top of the upper member is faulted in the central pendant, but very similar quartzites, although more highly metamorphosed, occur depositionally below the Wood Canyon Formation in the southern pendant.

Age and correlation: The distinctive quartzite-siltstone-quartzite sequence of the three members of the Stirling Quartzite is widely recognized in the southern Great Basin (Nolan, 1929; Wright and Troxel, 1966; Stewart, 1970), and is clearly present in the Avawatz

Mountains. Correlation of this sequence in the Avawatz Mountains
with similar sections elsewhere in the southern Great Basin is
based on lithologic similarities and on stratigraphic position. A
Precambrian age assignment is based on the lack of fossils within
the Stirling Quartzite and the presence of Early Cambrian fossils
in overlying formations (Stewart, 1970).

Wood Canyon Formation (€wc)

The Wood Canyon Formation is exposed discontinuously in a re-
markably linear west-dipping belt extending for perhaps five miles
from the central to the southern Avawatz Mountains (Plate 1A). Al-
though tectonic attenuation and metamorphism are greater in the
southern pendant, faulting and incomplete exposures in the central
pendant made this section unsuitable for measurement. Total measured
thickness in the southern pendant is 139.2 meters.

Lithology: The Wood Canyon Formation is composed primarily of medium-
to fine-grained, gray- to brown-weathering micaceous quartzite. It
varies from laminated to medium-bedded, and generally weathers into
beds 2-20cm thick. Some beds contain considerable silt-sized material
and are poorly indurated. Indurated quartzites locally contain mi-
caceous parting surfaces and biotite rich laminations. Cross-strati-
fied beds 2 to 7cm thick and pebbly beds with pebbles up to .5cm
diameter occur locally. Although generally gray to brown, color
varies from black to orange, red, or pale yellow. Sparse dolomite
marble and calc-silicate beds are tan or chocolate-brown, and green,
red, yellow or gray, respectively. The lower contact of the Wood
Canyon Formation is placed at the base of the lowest dark chocolate-
brown dolomite bed above the pure quartzites of the upper member of
the Stirling Quartzite. This contact can be mapped even where the
dolomite beds are missing since it corresponds to a color change from
orange quartzite of the Stirling Quartzite to brown quartzite of the
Wood Canyon Formation.

Age and correlation: The brown to gray-brown silty quartzite assigned

to the Wood Canyon Formation in the Avawatz Mountains is clearly distinguishable from relatively pure, orange quartzite of the underlying upper member of the Stirling Quartzite, and from pink, massive orthoquartzite of the overlying Zabriskie Quartzite. In addition, the silty quartzite is lithologically similar to rocks assigned to the Wood Canyon Formation in adjacent areas (Stewart, 1970; Wright and Troxel, 1966; Nolan, 1929). No fossils were found in the Wood Canyon Formation in the Avawatz Mountains. Stewart (1970) reports Early Cambrian olenellid trilobites and archeocyathids in the upper Wood Canyon Formation in other areas, and considers the Wood Canyon Formation Precambrian to Early Cambrian in age.

Zabriskie Quartzite (€z)

The Zabriskie Quartzite is a distinctive orthoquartzite that is exposed in both the central and southern pendants. The best exposed section is in the southern pendant, where the thickness was measured as 50.9 meters.

Lithology: The Zabriskie Quartzite is composed of medium(?)-grained, pink to white, generally massive to faintly thick-bedded orthoquartzite. It contains very sparse micaceous laminations, and tends to weather light brown near the top and base. The base of the Zabriskie Quartzite is placed at the top of the highest argillaceous, calcareous, or crumbly, dark colored quartzite bed in the Wood Canyon Formation.

Age and correlation: The distinctive lithologic characteristic of the Zabriskie Quartzite, plus its stratigraphic position between the silty quartzite of the Wood Canyon Formation and the siltstone and limestone of the Carrara Formation, permit unambiguous correlation of this formation (Stewart, 1970; Hazzard, 1937). Vertical worm burrows (Scolithus) are common in the Zabriskie, although none were found in the Avawatz Mountains. At other localities, the Zabriskie is both overlain and underlain by strata containing Early Cambrian fossils, indicating an Early Cambrian age for this formation (Stewart, 1970).

Carrara Formation (€c)

The Carrara Formation is exposed in both the south and central pendants. It is composed of three members: a lower argillaceous member (26.2m thick), a middle calcite marble member (34.0m thick) and an upper argillaceous member (36.9m thick). The lower and upper members are slope forming, whereas the middle member forms a cliff. Changes in the thickness due to deformation in the Carrara are moderate to extreme. The total measured thickness of 97.1 meters, determined in a section across the central pendant, is probably greater than at most localities. However, north of this transect thicknesses may be even greater.

Lithology: The lower member of the Carrara Formation is composed of fine-grained, black, gray or brown, argillaceous quartzite with local beds of fine-grained, orange quartzite. The quartzite is cleaved, weathers easily on micaceous or argillaceous bedding planes to form chips or plates, and is locally laminated. The lower con-tact of the Carrara is placed at the base of the lowest argillaceous quartzite bed greater than .5m thick.

The middle member is composed of medium- to light-gray, diffusely laminated and layered or massive calcite marble. It is generally slight-ly siliceous. Its lower contact is placed at the base of the lowest calcite marble bed greater than 1m thick.

The upper member is composed of quartzite, argillaceous quartz-ite, and calcite marble. Argillaceous quartzite, the dominant rock type, is generally black to brown, micaceous, fine-grained to very fine-grained, and with local greenish calc-silicate layers. Fine-grained quartzite beds are generally gray with rare calc-silicate layers. Calcite marble layers typically have a greenish tint indi-cating the presence of calc-silicate minerals. The lower contact is palced at the top of the highest calcite marble bed greater than 2 meters thick.

Age and correlation: The three members of the Carrara Formation in the Avawatz Mountains are probably correlative with the Latham Shale, Chambless Limestone, and Cadiz Formation (Hazzard and Mason, 1936;

Hazzard, 1954). These three formations are correlative with the
Carrara Formation as defined by Cornwall and Kleinhampl (1961)
(Stewart, 1970; Palmer and Halley, 1979). The lower member (Latham
Shale) is late Early Cambrian and earliest Middle Cambrian in age,
while the middle (Chambless Limestone) and upper (Cadiz Formation)
members are Middle Cambrian in age (Palmer and Halley, 1979). No
fossils were found in the Avawatz section, but correlation on
lithologic criteria and stratigraphic position is unambiguous.

Bonanza King Formation (€bk)

The Bonanza King Formation is composed almost exclusively of
calcite and dolomite marble. A complete section was carefully mapped
and pieced together, and in the process, the formation was divided
into ten lithologically distinctive units. Recognition of a calc-
silicate unit (unit #6) allows division of the Bonanza King Forma-
tion into the lower Papoose Lake Member (172.1m thick) and upper
Banded Mountain Member (235.9m thick). The total thickness of the
formation is 408 meters. A complete section of Bonanza King Forma-
tion is recognized only in the central pendant, and Bonanza King is
used as a map unit only in this area (Fig. 4). In other areas, the
obliterative effects of metamorphism and deformation make separation
of the Bonanza King Formation from younger Paleozoic carbonate rocks
difficult to impossible. The base of the Bonanza King Formation is
placed at the top of the highest argillite or fine-grained quartzite
bed greater than 20cm thick.

Lithology: The Bonanza King Formation is divided into the Papoose
Lake and Banded Mountain Members, and each member is divided into
five units. All units, except unit #6 which forms the base of the
Banded Mountain Member, are composed of calcite or dolomite marble.
Units 1 through 4 are composed predominantly of gray calcite marble
and form high cliffs. Dolomite marble, the predominant rock type
higher in the section, is less resistant to erosion and is more
internally brecciated. One isolated body of fine-grained, white
quartzite is present in unit #7(?).

The following is a unit by unit description of the Bonanza King Formation.

Papoose Lake Member (€bp)
Unit #1 (thickness 23.7m)

Unit #1 is composed of variably silt-bearing calcite marble. Silty layers are crumbly or fissile, locally with calcite crystals suspended in a silt matrix. Purer calcite marble layers protrude from less resistant silty layers, and locally weather into plates. Layering is defined by color, texture, and weathering contrasts, with layers typically .5 to 5cm thick. Silt-bearing layers are light gray to dusty gray-brown, while less silty calcite marble varies from medium to light gray and tan with gray and white laminated marble occurring in the upper part. Bedding is well-defined by the same criteria that define layering, with individual beds containing discrete layered packages that range from .2 - 2m thick. The base of unit #1 is placed at the top of the highest fine-grained quartzite or argillite bed greater than 20cm thick. The transition from slope forming siliceous rocks to cliff forming carbonate rocks typical of the Papoose Lake Member occurs within unit #1.

Unit #2 (thickness 48.4m)

Unit #2 is composed primarily of well-laminated to thin-bedded, gray and white calcite marble. Beds are laminated and thinly bedded, and range from .3 to 3cm thick. They commonly are variably mottled. White layers are composed of relatively pure calcite marble, whereas darker layers are slightly dolomitic. The dark layers vary from blue-gray to tan-gray on weathered surfaces, and are consistently medium to dark gray or blue-gray on fresh surfaces. Weathered surfaces of the dark layers are much rougher than those of the white calcite layers, which may be due to differential weathering of dolomite and calcite crystals. This gives the dark layers a grainy appearance. The sharply defined gray and white laminations, or "zebra stripe" appearance, is characteristic of this unit. Unit #2 consistently forms steep cliffs throughout the map area, and is the most resistant unit to weathering in the Bonanza King Formation. Its base is placed

at the top of the highest silty carbonate bed in Unit #1.

Unit #3 (thickness 13.9m)

The lower 8.5 meters of unit #3 is composed primarily of massive, tan-weathering, coarsely crystalline, slope-forming dolomite marble, which is white on fresh surfaces. The basal 2 meters is platy weathering and contains a single, 10cm-thick, white calcite marble bed. The upper 5.4 meters of unit #3 is composed of very coarsely crystalline calcite marble (crystals up to .5cm), which is white on both fresh and weathered surfaces. Unit #3 is a distinctive marker unit which can be recognized from a distance of a kilometer or more.

Unit #4 (thickness 38.0m)

Unit #4, a relatively pure calcite marble, is similar to unit #2. It is generally medium-gray to blue-gray, with light brown or gray, grainy, stringy or mottled laminations up to 2cm thick that may run together into layers as much as 30cm thick. Mottling is variable and more pervasive than in unit #2. Widely spaced silty calcite marble layers, .2 to 1m thick, weather to form recesses in an otherwise cliff-forming unit. The top 1.4 meters of unit #4 is a distinctive marker unit, and is composed of white calcite marble with tan, calcareous dolomite stringers and layers up to 8cm thick.

Unit #5 (thickness 48.1m)

Unit #5 is composed predominantly of light tan- or light gray-weathering calcite marble. Mottled stringers and laminations are sparse, diffuse, and faint. Bedding is best defined by cracks and recesses produced by preferential weathering along thin, less resistant horizons. Sparse, crumbly, silty beds occur in the unit, and the slope forming character of the entire unit suggests that it may be slightly silty throughout. A 6 meter thick, massive, tan-weathering dolomite bed occurs at the base.

Banded Mountain Member (€bb)

Unit #6 (thickness approximately 19.6m)

Unit #6 is composed of silty calcite marble and calc-silicate rock. Laminated to diffusely bedded calc-silicate rocks weather in negative relief. They are crumbly to platy weathering, and are orange, green, tan, or gray. Small (less than .2cm), individual calcite crystals are suspended in a crumbly, silty matrix. Gray to reddish-brown calcite-marble layers weather in positive relief. They are 1-5cm thick and are locally interbedded with the calc-silicate beds. Calcite marble and calcareous dolomite marble beds are slightly to moderately silt-bearing. Preferential weathering occurs on thin, siltier layers, producing blocky, moderately resistant outcrops. Brown-weathering siliceous laminations occur locally and weather in positive relief.

Silty carbonate rocks such as those in unit #6 weather easily and do not crop out well. Although unit #6 was measured in two different places, it was still inadequately described. A third, isolated section was measured in order to provide a more complete description (Section 6, Appendix A). Unit #6 is well-exposed in this third area, although underlying and overlying units were not suitable for measurement due to deformation, poor exposures, or vertical cliffs. The total thickness of silty carbonate or calc-silicate rocks at this location is about 8 meters, which is considerably less than the 19.6 meter thickness measured to the south at section 5A. At section 5A, the top of unit #6 was placed at the top of a calcite marble bed containing light green- or brown-weathering chert nodules or bands. Failure to recognize this cherty bed at section 6 may have led to the difference in thickness. This is a resonable supposition because the upper half of unit #6 at section 5A is more calcareous than the lower half, and its equivalent may have been included in unit #7 at section 6. Differential structural attenuation may also be a factor.

Unit #7 (total thickness 35.1m)

Unit #7 is composed of tan-orange-weathering, massive dolomite marble. Faint laminations and brown, siliceous stringers occur

locally. Slightly variable shades of orange define beds more than
1 meter thick near the top of the unit. Internal brecciation is
common in this and other massive dolomites.

Unit #8 (total thickness 24.8m)

Unit #8 is composed of tan-gray to white, massive to faintly
laminated or splotchy dolomite marble. It is less resistant than
the underlying unit, and weathers on bedding planes to form blocky
talus. The base of unit #8 is placed at a wide and distinctive
ledge at the top of the last orange dolomite bed of unit #7.

Unit #9 (total thickness 89.6m)

Unit #9 is composed primarily of calcite marble with minor inter-
beds of dolomite marble, calcareous dolomite marble, and dolomitic
calcite marble. The calcite marble ranges from gray to tan to white,
and is variably laminated and mottled. This unit can be divided into
three subunits. The lower and upper subunits are composed of both
dolomite and calcite marble. Dolomite is typically massive and tan-
weathering, whereas calcite-bearing rocks range from massive to lami-
nated. Sparse, thin (1-3cm) dark brown and green calc-silicate layers
occur locally in the upper subunit. The middle subunit is composed
exclusively of massive to laminated calcite marble, and is more
resistant and cliff-forming than the upper and lower subunits. A
distinctive marker bed composed of thinly laminated calcite marble
occurs within the middle subunit in both measured section 5B and 7.

Unit #10 (total thickness 66.8m)

Unit #10 is composed of massive, tan-weathering, coarse-grained
dolomite marble. It is white on fresh surfaces, and contains very
sparse calc-silicate stringers and nodules. The base of unit #10
is placed at the top of the highest calcite marble bed greater than
.5m thick in unit #9.

Age and correlation: The Bonanza King Formation was named and de-
scribed by Hazzard and Mason (1936) from exposures in the Providence
Mountains. Palmer and Hazzard (1956) report fossil evidence for a

Middle to Late Cambrian age for the Bonanza King Formation, and
correlate the formation from the Providence Mountains to the Nopah
Range and Goodsprings District of Nevada. Barnes and Palmer (1961)
formally subdivided the Bonanza King Formation into the lower
Papoose Lake Member and upper Banded Mountain Member. A silty
dolomite unit forms the base of the upper Banded Mountain Member,
and is readily identifiable as the rest of the Bonanza King Forma-
tion is composed of dolomite and calcite marble (Gans, 1974). On
the basis of lithologic similarity, and its stratigraphic position
above the Carrara Formation and below calc-silicate rocks assigned
to the Dunderberg Shale Member of the Nopah Formation, the carbonate
sequence described above is correlated with the Bonanza King Formation.

Nopah Formation (€n)

Rocks assigned to the Nopah Formation are exposed only in one
small area at the northwest edge of the central pendant (Fig. 4). At
this locality, a 5m thick sequence of thin-bedded green, brown and
maroon calc-silicate rocks and gray calcite marble can be traced for
about 300m. This sequence overlies massive, tan dolomite marble
of the uppermost Bonanza King Formation, and is overlain by massive,
tan dolomite marble. Approximately 30m above the calc-silicate se-
quence, Devonian stromatoporoids first occur within the dolomite mar-
ble. A disconformity between Devonian, stromatoporoid-bearing dolomite
and the upper Cambrian Nopah Formation must occur within the 30 meters
of massive dolomite marble above the calc-silicate unit.

Age and correlation: The Nopah Formation was named and described by
Hazzard (1937) for exposures in the Nopah Range, and assigned a Late
Cambrian age based on fossil evidence. The basal Dunderberg Shale
Member is the most distinctive unit within the Nopah Formation, and
is recognized over a broad area of the southern Great Basin and Mojave
region. Correlation of calc-silicate rocks in the Avawatz Mountains
with the Dunderberg Shale Member is based on: 1) lithologic similarity,
2) the fact that it is the second calc-silicate unit to occur above
the Carrara Formation, 3) it is both overlain and underlain by mas-
sive dolomite, apparently a characteristic of more cratonal sections

(Gans, 1974; B. C. Burchfiel, personal communication, 1980), and
4) the immediately overlying dolomite is in turn overlain discon-
formably by Devonian stromatoporoid-bearing dolomite of the Sultan
Limestone, a characteristic of platformal sections in nearby areas
(Novitsky-Evans, 1978; Burchfiel and Davis, 1977).

Ironside-Valentine Members, Sultan Limestone, undifferentiated (Dsiv)

Rocks assigned to the Ironside and Valentine Members of the Sul-
tan Limestone are clearly identifiable only in one area along the
west side of the central pendant. They are composed of massive, tan
and gray-weathering, coarsely crystalline dolomite marble (Fig. 4).
Stromatoporoids occur throughout the Ironside-Valentine, but are
most abundant in the lower part of the sequence. At two localities
on the ridge crest, stromatoporoids are so abundant they form a
significant percentage of the rock. The stromatoporoids are con-
centrically laminated, elliptical to circular, 2-20cm diameter,
and are composed of tan calc-silicate rock, white calcite marble,
and tan dolomitic marble. These are clearly distinguishable from
algal stramatolites or algal concretions because they contain septa
between concentric laminations that divide the inter-lamination
area into chambers. The lower contact of the formation was mapped at
the base of the lowest stromatoporoid-bearing beds, although it
could actually be as much as 30 meters lower. Total thickness of
these rocks is poorly constrained but is estimated as 100m.

Age and correlation: The Sultan Limestone was named and described
by Hewett (1931) from localities in the southern Spring Mountains,
Nevada. Hewett (1931) divided the Sultan into three members: 1)
the basal Ironside Dolomite, 2) the Valentine Limestone, and 3) the
upper Crystal Pass Limestone. The Ironside Dolomite Member, where
studied by Hewett (1931, 1956), is generally a dark, smokey gray
dolomite. The Valentine Limestone Member commonly shows partial
or complete alteration to dolomite, whereas the Crystal Pass Limestone
Member consistently retains its pure limestone composition, even where
both overlying and underlying strata are altered entirely to dolomite.
Stromatoporoids are common in the Ironside and Valentine Members, but

are rare in the Crystal Pass Member (Hewett, 1956).

Correlation of stromatoporoid-bearing dolomites in the Avawatz Mountains with the Ironside and Valentine Members of the Sultan Limestone is based on: 1) fossil evidence, and 2) lithologic similarity. Metamorphism and tectonic disruption prohibit separation of the Ironside and Valentine Members in the Avawatz Mountains. The presence of stromatoporoids provide the best evidence for a Middle to Late Devonian age, as indicated elsewhere on the basis of paleontological data (Hewett, 1956; Bereskin, 1976).

Crystal Pass Member, Sultan Limestone (Dsc)

The Crystal Pass Member is exposed on the west flank of both the central and southern pendants (Figs. 4, 5), and is composed almost exclusively of coarsely crystalline (up to 1cm crystals), light gray to white calcite marble. The rock is generally massive, although bedding is locally defined by faint color banding and by weathering on bedding planes that may produce remarkably smooth surfaces of several square meters. Laterally discontinuous fine-grained quartzite beds (1-3m thick) crop out at three localities within the calcite marble. Interbedded calcite and dolomite marble forms a 10m-thick transition zone at the base of the Crystal Pass Member. The base of the Crystal Pass Member was mapped at the top of the highest tan or gray dolomite marble bed greater than 1m thick within the transition zone. The top of the Crystal Pass Member has been obliterated by Mesozoic granitic intrusions.

Age and correlation: Based on its pure calcite marble composition and stratigraphic position above stromatoporoid-bearing dolomites, this unit is correlated with the Crystal Pass Limestone Member of the Devonian Sultan Limestone, as defined by Hewett (1931).

Bonanza King Formation, Nopah Formation, and Sultan Limestone, undifferentiated (€Du)

This designation was used for a single large body in the south pendant composed primarily of dolomite marble (Fig. 5). It rests conformably on the Carrara Formation and is conformably overlain by

coarse, calcite marble of the Crystal Pass Member of the Sultan
Limestone. The carbonate rocks in this body are so strongly brec-
ciated, metamorphosed, and deformed that it was not possible to
divide them into formations. One low-angle fault visible on the
east face of the pendant appears to attenuate the section. Stroma-
toporoids were found at one locality above this fault.

These metasediments are correlated with the Bonanza King Forma-
tion, Nopah Formation, and Ironside-Valentine Members of the Sultan
Limestone, on the basis of stratigraphic position, lithology, and
fossils.

Bird Spring Formation and Monte Cristo Formation, undifferentiated (PCu)

These formations crop out primarily in two areas along the northern
part of the Arrastre Spring fault: along upper Sheep Creek and north-
west of Arrastre Spring (Plates 1A, 2). At all localities, these
rocks are either cut by faults or intruded by plutons, and are not
in depositional contact with underlying or overlying formations.
They are composed of gray and blue-gray calcite marble interbedded
with tan-orange or tan-brown calc-silicate rocks. Elongate chert
nodules 2-10cm thick are present locally. Individual calcite marble
beds 5-20cm thick locally contain abundant fossil fragments and are
internally graded. Single calcite crystals, forming isolated lenses,
are probably crinoid columnals. Fusilinids were located 1/2 mile
N55W from Arrastre Spring.

Age and correlation: The Monte Cristo Limestone and Bird Spring
Formations are composed primarily of limestone and dolomite, with
sparse to abundant beds of calc-silicate rocks, fine-grained sand-
stone, chert, and shale. Correlation of the metasedimentary rocks
in the Avawatz Mountains with these formations is based on lithologic
similarity, and on the presence of fusilinids of Pennsylvanian or
Permian age. Fossils from the Monte Cristo Formation in the Good-
springs area indicate an Early Mississippian age, whereas those from
the Bird Spring Formation are of Late Mississippian, Pennsylvanian,
and Permian(?) age (Hewett, 1931, 1956).

Bonanza King Formation through Bird Spring Formation, undifferentiated
(Pc)

Pendants and faulted blocks of Paleozoic metacarbonate rocks
occur over a large area along the northern half of the Arrastre
Spring fault and within and adjacent to the south pendant. In the
northern area, these form a half dozen elongate, highly irregularly
shaped pendants, and several smaller, satellitic pendants. They
contain variably metamorphosed, brecciated, and deformed calcareous
metasedimentary rocks which generally resemble rocks designated as
Bird Spring Formation and Monte Cristo Formation, but locally include
rocks that contain stromatoporoids(?), and white, coarsely crystal-
line calcite marble. In the southern area, these rocks occur in two
locations: 1) at the north end of the south pendant, and 2) in a
klippe just east of the north end of the south pendant (Fig. 5).
The carbonate rocks within the south pendant contain several calc-
silicate beds up to 5-8m thick, one laterally discontinuous quartz-
ite bed, and a large amount of massive dolomite and calcite marble.
These rocks are not clearly correlative with any of the previously
described formations. The klippe contains similar rocks including,
at one locality, concentric, eliptical calc-silicate bodies 5-10cm
in diameter that resemble stromatoporoids. Also within the klippe
are coarsely crystalline tan dolomite beds that grade upward over
10-30cm into bright maroon-red calc-silicate rocks.

Age and correlation: These rocks are of uncertain specific age and
correlation, but contain sufficient lithologic similarity to ad-
jacent Paleozoic metacarbonate rocks that they are confidently cor-
related with them. They almost certainly fall within the sequence
Cambrian Bonanza King Formation-Pennsylvanian-Permian Bird Spring
Formation.

Pre-Mesozoic metacarbonate rocks (mc)

Pre-Mesozoic metacarbonate rocks include brecciated, metamorphosed,
coarse-grained dolomite or calcite marble of uncertain age. Most of
these metacarbonate bodies are small and are surrounded by younger
intrusive rocks. However, one group of metacarbonate rocks at the

southeast flank of the Avawatz Mountains included two moderate
size bodies (1/4 mi.^{2}) that appear to be intruded by Precambrian
gneiss (pꞓg). This intrusive relationship is strongly suggested
by the irregular nature of the contact between the two, but can not
be positively confirmed because of pervasive alteration across a
broad zone adjacent to the contact. If the gneiss is indeed intru-
sive into the carbonate, it indicates that the carbonate is older
than about 1400 m.y..

Pre-Cenozoic plutonic-metamorphic rocks, undifferentiated (pm)

Rocks assigned to this undifferentiated unit are so hopelessly
altered by metamorphism, brecciation and deformation that their
protolith is difficult or impossible to determine in the field, even
with excellent exposure. They occur in four areas: 1) at the south-
east end of the range, adjacent to the Precambrian gneiss, is a
brecciated and faulted mass of rocks that include coarse-grained,
prophyritic pink biotite granite(?), dark, fine-grained, biotite
quartz diorite(?), breccia of chlorite-biotite-feldspar(?)-quartz(?)
clasts in a quartz-feldspar matrix (brecciation during intrusion?),
calcite and dolomite marble, dikes of pink granite(?) intruding
intermediate plutonic rock, and altered, fine-grained to aphanitic
green, gray, and yellow rocks of unknown protolith. These rocks
occur within the east branch of the Arrastre Spring fault zone and
have undoubtedly suffered their most recent abuses as a result.
2) At the pass between the south pendant and the klippe of Paleozoic
metacarbonate rocks is an area of fault gouge of uncertain but proba-
bly plutonic protolith. 3) Within the diorite complex and adjacent
to several small bodies of metacarbonate rock, just east of the cen-
tral pendant and below a major low-angle fault, is an area of highly
altered rock whose protolith is probably diorite and partially as-
similated carbonate. 4) A mass of highly brecciated metasedimentary
and metaigneous(?) rock lies above a low-angle fault at the east edge
of the central pendant. These rocks may include sedimentary as well
as tectonic breccias.

REGIONAL SIGNIFICANCE OF LATE PRECAMBRIAN-
PALEOZOIC STRATIGRAPHIC SECTION IN THE AVAWATZ MOUNTAINS

Northwestward thickening of the Cordilleran miogeocline occurs
by three mechanisms: 1) thickening of all units, 2) successive addi-
tion of Ordovician, Silurian, and lower Devonian strata beneath a sub-
Middle Devonian unconformity, and 3) addition of a thick sequence of
upper Precambrian terrigenous rocks. Determination of stratigraphic
sequence and thickness within a particular area thus allows approxi-
mate placement of that area within a transect across the miogeocline.
However, overall thickness determinations are only approximately
accurate in areas that have undergone deformation, plutonism, and meta-
morphism because of potentially severe tectonic thickness modifica-
tion. As a result, determination of stratigraphic sequence is most
useful in resolving the configuration of the Cordilleran miogeocline
in the Mojave region.

The Avawatz upper Precambrian-Paleozoic section exhibits the
following features: 1) a fully developed sequence of upper Precambrian-
Lower Cambrian terrigenous strata, and 2) a platformal Paleozoic car-
bonate sequence in which Devonian strata rest disconformably on upper
Cambrian strata. Similar sequences are found in the Providence and
Cowhole Mountains to the southeast (Stewart, 1970; Novitsky-Evans,
1978), and Victorville area 115km to the southwest (Miller, 1977).
In contrast, the New York Mountains, located about 90km to the east,
contain a comparable thickness and sequence of Paleozoic carbonate
rocks resting on the Lower Cambrian Tapeats Sandstone and Middle
Cambrian Bright Angel Shale (Fig. 8; Burchfiel and Davis, 1977). This
thin sandstone and shale sequence rests directly on Precambrian crystal-
line basement. Thus, it appears that westward to northwestward thick-
ening of the miogeocline, by the addition and thickening of upper
Precambrian terrigenous rocks, occurs within the Mojave region. In
contrast, the Paleozoic carbonate sections of all these areas contain
a Devonian-Cambrian disconformity and thus maintain their platformal
or cratonal affinity across the Mojave region (Fig. 8). This indicates
that the subsidence rate of the Mojave region was moderate during late

Fig. 8. Stratigraphic correlation chart for upper Precambrian
Paleozoic rocks across the Mojave region. See Figure 9 for
location. Note that Devonian rocks rest on Cambrian rocks in
all sections except possibly the Providence Mountains where the
exact stratigraphy of this part of the section is uncertain.

Fig. 9. Schematic map of southern Great Basin-Mojave region.
All Paleozoic rocks southwest of the Las Vegas shear zone are
shown. Reconstruction of Tertiary faulting based on Davis and
Burchfiel (1973) and Wright and Troxel (1967). Zero isopachs
shown for Ordovician Eureka Quartzite, Ordovician Pogonip Group,
and all Silurian rocks. Note angular discordance between these
isopachs and boundary with eugeoclinal rocks. Control points
for isopach location are shown by numbered circles: 1 = Talc
City Hills, Hall and MacKevett (1962); 2 = Argus Range, Streitz
and Stinson (1974); 3 = Nopah Range, Burchfiel and others, in
preparation; 4 = Spring Mountains, Burchfiel and others (1974);
5 = Avawatz Mountains, this report; 6 = Cowhole Mountains,
Novitsky-Evans (1978); 7 = Providence Mountains, Stewart (1970)
and Hazzard (1954); 8 = New York Mountains, Burchfiel and Davis
(1977); 9 = Spring Mountains, Gans (1974); 10 = Sidewinder Moun-
tain, Miller (1977).

COWHOLE MTS.

NOVITSKY EVANS (1978)

PROVIDENCE MTS.

HAZZARD (1954)

STEWART (1970)

NEW YORK MTS.

BURCHFIEL AND DAVIS (1977)

SIDEWINDER MT.

MILLER (1977)

AVAWATZ MTS.

SPENCER (1981)

BIRD SPRING FM

BIRD SPRING FM

BIRD SPRING FM

MONTE CHISTO LMSTN

SULTAN LMSTN

NOPAH FM

BANDED MOUNTAIN MBR

BANDED MOUNTAIN MBR

PAPOOSE LAKE MBR

CARRARA FM

ZABRISKIE W C FM

CRYSTAL PS
IRONSIDE-VALENTINE UNDIF
NOPAH FM

BANDED MOUNTAIN MBR

PAPOOSE LAKE MBR

CARRARA FM

ZABRISKIE

WOOD CANYON FM

STIRLING QTZ

STIRLING QTZ

JOHNNIE FM

MONTE CHISTO LS

CRYSTAL PASS

VALENTINE

NOPAI FM

BANDE MTN MBR

PAPOOSE LAKE MBR

CARRARA FM

ZABRISKIE QTZ

WOOD CANYON FM

MONTE CRISTO LMSTN

CRYSTAL PASS

VALENTINE

IRONSIDE

BONANZA KING FM

CARRARA FM

ZABRISKIE

WOOD CANYON FM

STIRLING QTZ

JOHNNIE

MONTE CRISTO LMSTN

CRYSTAL PS
VALENTINE MEMBER

NOPAH FM

BANDED MOUNTAIN MBR

SILTY UNIT

PAPOOSE LAKE MBR

BRIGHT ANGLE

TAPEATS

DISCONFORMITY

DISCONFORMITY

DISCONFORMITY

DISCONFORMITY

DUNDERBERG SHALE MBR

DUNDERBERG SHALE MBR

DUNDERBERG SHALE MBR

DUNDERBERG SHALE MBR

SILTY UNIT

SILTY UNIT

FLAT UNIT

PENN.

MISS.

DEV.

CAMBRIAN

PRECAMBRIAN

51

Fig. 8.

Fig. 9.

Precambrian-Cambrian time, but slowed greatly during early Paleozoic time.

Northwestward thickening of the miogeocline by all the mechanisms discussed above is well documented in the Spring Mountains (Burchfiel and others, 1974). The zero isopach for upper Precambrian terrigenous rocks trends southward from the Spring Mountains, whereas the zero isopachs for Ordovician and Silurian Formations trend in a west to southwest direction. Recognition of the Avawatz Paleozoic section places important new constraints on the location of Silurian and Ordovician zero isopachs, and adds new evidence supporting the divergence of Precambrian and Paleozoic isopachs in the Mojave region (Fig. 9).

The configuration and subsidence history of the miogeocline is in part indicated by the orientation of zero isopachs for Silurian and Ordovician Formations. The zero isopach for the Lower and Middle Ordovician Pogonip Group crosses the Keystone thrust plate in the Spring Mountains between upper Red Rock Canyon and Mountain Springs Pass (Gans, 1974). To the west, the northernmost possible location of this isopach is constrained by Paleozoic sections in the Nopah Range (Burchfiel and others, in preparation), Argus Range (Streitz and Stinson, 1974) and Talc City Hills (Hall and MacKevett, 1962). The southernmost possible location is constrained by exposures in the Clark Mountains (Burchfiel and Davis, 1971) and Avawatz Mountains (Fig. 9). The zero isopach for the Middle Ordovician Eureka Quartzite lies within the Keystone thrust plate about 10 to 20km north of the zero isopach for the Pogonip Group (Burchfiel and others, 1974). Its location west of the Spring Mountains is constrained by the same sections as for the Pogonip Group. The zero isopach for Silurian rocks lies within the Keystone and Lee Canyon thrust plates and trends west-southwestward through the northern Nopah Range, where it lies at least 20km north of exposures of the Eureka Quartzite (Burchfiel and others, in preparation). Farther west it is constrained only to lie between sections in the Argus Range and Avawatz Mountains (Fig. 9). Structural correlation of the Lee Canyon thrust plate and the structurally lowest rocks of the Nopah range (Burchfiel and others, in preparation), indicate that the Silurian zero isopach extends for 40km in a S30W direction and is not offset by Mesozoic thrust faults. This provides

excellent control on the orientation of a segment of this isopach.

Depositional facies boundaries and isopachs of Paleozoic mio-
geoclinal sedimentary rocks trend in a southwesterly direction toward
a poorly defined boundary across which occur rocks of eugeoclinal
affinity in the western Mojave region (Fig. 9). These eugeoclinal
metasediments include the Garlock Formation in the El Paso Mountains
(Dibblee, 1967), its equivalent offset by the Garlock fault to Pilot
Knob (Smith and Ketner, 1970), rocks in the Goldstone Lake area
(Miller and others, 1979), Lane Mountain area (McCulloh, 1952, 1960),
and possibly in the Shadow Mountains west of Victorville (Miller, 1977;
Troxel and Gunderson, 1970). These rocks are generally composed of
chert, phyllite, greenstone, limestone, impure sandstone, and conglomer-
ate, and do not in any way resemble miogeoclinal sequences. Some of
these rocks are similar to rocks in west-central Nevada which were
deposited in a continental slope and rise environment and later thrust
eastward onto the outer continental margin during Paleozoic deformations
(Roberts and others, 1958).

Three hypotheses have been proposed to explain the apparent
juxtaposition of the western Mojave eugeoclinal rocks with the mio-
geocline. Poole (1974) suggested that these rocks represent the
southward continuation of the Antler and Sonoma orogenic belts from
west-central Nevada. According to this hypothesis, eugeoclinal rocks
were thrust southeastward across the continental margin during middle
and late Paleozoic time, and the present, irregular trace of the
boundary across which eugeoclinal rocks occur is attributed to oro-
clinal bending of Mesozoic and Cenozoic age (Poole and Sandberg, 1977).

Dickinson (1981) attributed the irregular trend of the boundary
to an irregular continental margin inherited from a Precambrian rift-
ing event. Promontories and recesses in the continental margin formed
as a result of an irregular pattern of rifts and transform faults.
He proposed that the western Mojave siliceous rocks are not appreciably
out of place with respect to their original depositional setting rela-
tive to the miogeocline.

Burchfiel and Davis (1981) rule out Dickinson's hypothesis be-
cause isopachs in the miogeocline trend at a high angle toward the

boundary, and show no sign of curving to conform with it. They attribute this pattern to tectonic juxtaposition of the two rock types, but also rule out the overthrust model of Poole (1974) and Poole and Sandberg (1977) because of the absence of a clastic wedge (overlap assemblage) shed southeastward onto the miogeocline from overthrust allochthonous rocks. Alternatively, Burchfiel and Davis (1981) propose that the western Mojave siliceous rocks have been tectonically juxtaposed against the continental margin by a left-lateral, strike-slip faulting event in Permo-Triassic time. The present sinuous trace of this strike-slip fault boundary is, in their opinion, the result of right-lateral oroclinal bending in Mesozoic or Cenozoic time.

The discovery of the Paleozoic carbonate section in the Avawatz Mountains contributes to resolution of these interpretations. The Avawatz Mountains lie at the west edge of the miogeocline, and thus represent an important control point for isopachs as they project toward the western Mojave siliceous rocks. Specifically, Ordovician and Silurian zero isopachs lie north of the Avawatz Mountains. The Ordovician zero isopach is farther north than previously suspected (Burchfiel and Davis, 1981). The result of adding this control point to the Ordovician and Silurian zero isopachs is to further accentuate the angular discordance between isopachs of the miogeocline and the boundary across which lie the western Mojave eugeoclinal rocks. This angular discordance now appears to be approximately 90° (Fig. 9). Isopachs show no evidence whatsoever of bending to the south and wrapping around the western Mojave eugeoclinal rocks. This adds substantial evidence in favor of a tectonic juxtaposition model, as advocated by Burchfiel and Davis (1981) and Poole (1974), and against an irregular continental margin model, as proposed by Dickinson (1981). It also adds evidence in favor of the hypothesized tectonic truncation of the Cordilleran miogeocline, as originally envisioned by Hamilton and Meyers (1966) and Hamilton (1969), and later expounded by Burchfiel and Davis (1972, 1975).

Stewart (1967) proposed that approximately 80km of right-lateral offset had occurred on the Death Valley Furnace Creek fault zone,

based on apparent offset of isopachs of upper Precambrian and Lower Cambrian terrigenous rocks. Restoration of such movement would produce a southward curve in the Silurian and Ordovician zero isopachs, and thus reduce the angular discordance between these isopachs and the boundary with the western Mojave siliceous rocks. Wright and Troxel (1967), however, argue that displacement on the southern Death Valley fault zone is limited to 8km or less by linear geologic features that cross the fault. Stewart and others (1968) reconcile apparently conflicting data by proposing that large-scale oroclinal bending has accommodated right-lateral offset in the southern Death Valley area. Wright and Troxel (1970) argue forcefully against the oroclinal bending hypothesis. They note that geologic features in the Black Mountains are not rotated with respect to regional orientations, as would be expected if significant oroflexural bending had occurred. In addition, the southern Death Valley fault zone places diorite against diorite on the east flank of the Avawatz Mountains (Plates 1, 2), constraining offset to probably less than 10km.

Data from the Avawatz Mountains does not allow elimination of one of the two proposed hypotheses for the nature of tectonic juxtaposition, but does place the following constaint. Silver and Anderson (1974) envision a left-lateral "megashear" extending from Sonora, Mexico, northwestward into southern California and across the Mojave region. Although major displacement is now ruled out within the Mojave region (Stewart and Poole, 1975; Miller, 1977), a splay off this fault could cross the central Mojave region to connect with the long, linear, northwest-trending part of the tectonic juxtaposition boundary. This hypothesis would include a left-lateral displacement of at least 30km for both the western Mojave siliceous rocks and the southwestern Mojave miogeoclinal rocks from a position adjacent to west-central Nevada. However, restoration of more than about 100km of left-lateral displacement would place the Sidewinder Mountain carbonate section, containing no Ordovician or Silurian rocks (Miller, 1977), north of the southwestward projection of the Ordovician zero isopach. Thus, total left-lateral offset on all hypothesized left-lateral faults across the central Mojave region is limited to perhaps 100km, and may actually be non-existent or right-lateral.

APPENDIX A

Appendix A contains the descriptions of a composite measured
section extending from the upper Johnnie Formation of late Pre-
cambrian age to the upper part of the upper Devonian Crystal Pass
Member of the Sultan Limestone. It is divided into individual measured
sections which represent single traverses through each part of the
section. The location of each measured section is shown on Firgures
10 and 11.

Fig. 10. Simplified geologic map of central pendant and surrounding
areas. Locations of measured sections 1, 4, 5, 6, and 7 are shown
by numbered lines. Stromatoporoid-bearing rocks are shown by circled
dots.

Fig. 11. Simplified geologic map of southern pendant and surrounding
areas. Locations of measured sections 2 and 3 are shown by numbered
lines. Stromatoporoid-bearing rocks are shown by circled dots.

Fig. 10.

Fig. 11.

UPPER JOHNNIE FORMATION AND STIRLING QUARTZITE,
MEASURED SECTION #1A

	Thickness
Covered.	1.0m
Dolomite marble, medium gray, with local tan-weathering, siliceous, irregular blobs. Faint light and dark gray laminations and etched grooves define bedding. Sedimentary breccia with clasts 5-15cm occurs locally.	7.8m
Quartzite, fine-grained, orange to maroon to brown on weathered surfaces, white to gray on fresh surfaces. Beds weather into plates 1-5cm thick with local rythmic beds 2-3cm thick and local coarse layers 2-5cm thick.	19.0m
Calcareous dolomite marble, fine-grained, tan- to gray-weathering.	2.5m
Quartzite, fine-grained, orange to maroon to brown on weathered surfaces, white to gray on fresh surfaces. Beds weather into plates 1-5cm thick.	5.9m
Quartzite, fine-grained, as below but locally containing silt.	12.7m
Covered.	4.2m
Siltstone, white to light orange, platy.	.5m
Dolomite marble, tan-weathering.	.7m
Quartzite, black to dark gray, finely laminated, fractures on bedding to form blocky outcrops and talus. Local 1cm-thick graded beds with sharp upper and lower contacts.	5.8m
Covered.	7.0m
Covered.	9.6m
Quartzite, orange-weathering, white on fresh surfaces, fine-grained and not very resistant. Thickly bedded with thin, shaley partings.	13.6m
Calc-silicate rock, green to brown, with pyrite(?) crystals up to .4cm, laminated.	1.0m
Dolomite marble and calc-silicate rock, greenish-gray to brown, thin-bedded with sparse brown siliceous stringers.	15.3m

Dolomite marble and calc-silicate rock. Tan dolomite is
gradationally interbedded with dark to light green calc-
silicate rock.   4.8m

BASE OF STIRLING QUARTZITE   (Minimum Thickness - 224.1m)
BASE OF LOWER MEMBER, STIRLING QUARTZITE   (Total Thickness - 63.7m)

Quartzite, maroon to tan, fine- to medium-grained, thin-
to medium-bedded, slightly to moderately resistant.   4.8m

Quartzite, maroon, orange and brown, thick-bedded.   3.3m

Quartzite, maroon, orange and brown, medium- to coarse-
grained, medium- to thick-bedded, resistant.   34.9m

Quartzite, white, medium- to fine-grained, thin-bedded.   20.7m

BASE OF MIDDLE MEMBER, STIRLING QUARTZITE   (Total Thickness - 51.6m)

Quartzite, argillaceous, very fine-grained, black and gray,
laminated and thinly bedded.   4.5m

Dolomite marble, tan-weathering.   .5m

Quartzite, argillaceous, very fine-grained, black and gray.   4.9m

Dolomite marble, tan-weathering.   1.4m

Quartzite.   .5m

Dolomite marble, tan.   .3m

Quartzite, argillaceous, black.   1.5m

Dolomite marble, tan.   .5m

Quartzite, black to dark orange, fine-grained, thinly bedded
to laminated, platy.   10.3m

Quartzite, black to dark orange, fine-grained, thinly bedded
to laminated, platy, locally with starved ripples (flaser beds)
1-2cm thick. Argillaceous beds .1-1cm thick. Parting on
argillaceous beds.   27.2m

BASE OF UPPER MEMBER, STIRLING QUARTZITE

Quartzite, medium brown to light gray, medium-grained beds
3-20cm, cross-beds 2-5cm and lamination occur locally on
otherwise massive appearing orthoquartzite.   14.5m

Quartzite, massive to thick-bedded.

STIRLING QUARTZITE, MEASURED SECTION #1B

	Thickness
Quartzite, fine-grained, black, thinly bedded and laminated.	2.5m
Covered.	4.2m
Quartzite, fine-grained, greenish brown to red, massive.	.8m
Dolomite and calc-silicate rock, interbedded, tan and green. Diffuse interbeds 2-10cm. Greenish calc-silicate layers weather to form recesses.	2.1m
Covered.	1.3m
BASE OF UPPER MEMBER OF STIRLING QUARTZITE (Minimum Thickness - 108.8m)	
Quartzite, fine-grained, pale green to pale red-brown, locally white, weathers along bedding to form blocks.	4.2m
Orthoquartzite, pearly white to reddish-brown, medium- to coarse-grained, massive.	.5m
Quartzite, fine-grained, greenish gray, faintly laminated and thin-bedded.	2.0m
Quartzite, pearly white and splotchy red-brown with sparse, planar bedding plane fractures.	.4m
Quartzite, fine-grained, dull greenish gray, faintly bedded with sparse 5-10cm beds of white quartzite.	3.8m
Orthoquartzite, medium- to coarse-grained, white to reddish brown or orange, weathers on bedding planes to form plates and blocks 5-40cm thick. Low-angle cross-beds occur locally. Massive, brown-weathering beds up to 1m thick occur locally in upper half of this unit.	41.2m
Orthoquartzite, as below, but with local purplish-brown and white quartzite beds .5-2cm thick in packets 5-10cm thick between massive to faintly bedded blocky quartzite.	19.0m
Pebbly quartzite, pebbles to 1cm diameter, white, slightly crumbly.	2.2m
Quartzite, locally pebbly, white and brown.	4.9m
Covered. Possible fault here.	1.2m
Quartzite, fine- to medium-grained, brown and white, thinly bedded.	29.4m

Fault.

WOOD CANYON FORMATION, MEASURED SECTION #2A

	Thickness
BASE OF WOOD CANYON FORMATION (Total Thickness - 139.2m)	
Dolomite, massive, slightly calcareous, chocolate brown on weathered surfaces, medium gray on fresh surfaces.	.4m
Quartzite, medium-grained, light gray, light yellow or red. Dark gray micaceous laminations form parting surfaces. Local 1-4cm thick cross-beds.	1.7m
Quartzite, medium-grained, orange- to light gray-weathering, with biotite-rich laminations.	.8m
Quartzite, fine-grained, dark gray to black, laminated.	.3m
Quartzite, interbedded coarse- and fine-grained, light orange-weathering, with local biotite-rich layers 2-10cm thick.	.6m
Quartzite, interbedded black, laminated, fine-grained quartzite and crumbly, biotitic, medium-grained quartzite. Interbeds 2-15cm.	.8m
Dolomite, chocolate brown, gray on fresh surfaces, massive.	.7m
Quartzite, medium- to coarse-grained, black and gray laminated, crumbly with parting on biotite-rich laminations.	2.7m
Covered.	2.5m
Quartzite, fine- to medium-grained, black and gray laminations, with biotite-rich parting layers.	1.7m
Covered.	1.7m
Quartzite, fine- to medium-grained with local coarse, lenticular beds several tens of centimeters long, 2-5cm thick. Fine- to medium-grained quartzite is thinly laminated, black, and micaceous. Local cross-beds 2-5cm thick. Fairly resistant overall.	3.3m
Quartzite, as below but medium- to coarse-grained with local pebbly layers, pebbles up to .5cm diameter.	3.3m

Quartzite, medium-grained, locally coarse-grained, medium to light gray, thinly laminated, with local coarse, lenticular, cross-bedded sand layers. Paring and weathering on biotite-rich laminations.   8.1m

Quartzite, fine-grained, black, micaceous, crumbly.   .9m

Quartzite, fine- to coarse-grained, light to dark gray or reddish orange, crumbly, with biotite-rich laminations and parting surfaces.   3.1m

Quartzite, fine grain, black.   .5m

Quartzite, fine- to coarse-grained, crumbly, with biotite-rich laminations and parting surfaces, with local cross-beds 2-5cm thick.   2.7m

Quartzite, interbedded 1) fine-grained, black, biotitic quartzite, 2) cross-bedded, medium-grained, light gray quartzite, with biotite-rich layers, and 3) coarse-grained, light gray to light yellow-orange, diffusely and sparsely laminated quartzite, locally with well-developed cross-beds. Beds 2-20cm thick.   11.7m

Quartzite, interbedded 1) fine-grained, thinly laminated, biotitic, dark gray-black and green-black quartzite, and 2) medium- to coarse-grained, white to light gray or medium orange quartzite with sparse biotite laminations and local cross-beds 3-7cm thick. Beds generally 20-30cm thick.   26.0m

Quartzite, fine-grained, crumbly, dark red to maroon.   5.0m

Quartzite, fine-grained, dark to medium gray, faintly and thinly laminated.   9.9m

Quartzite, medium-grained, medium gray to light gray, with sparse biotite laminations, sparse cross-beds.   3.0m

Quartzite, medium- to fine-grained, brown-weathering, with wavy laminations.   3.0m

Quartzite, medium- to fine-grained, with sparse to abundant biotite partings.   4.9m

Quartzite, fine-grained, laminated, micaceous, dark gray to black.   4.9m

Quartzite, fine-grained, micaceous, dark, interbedded with medium- to fine-grained, greenish-gray quartzite, massive to faintly laminated. Interbeds 5-20cm thick.   5.0m

Quartzite and calc-silicate(?) rock, fine-grained, green to red, parting on local green chloritic(?) laminations.   2.5m

Quartzite, fine-grained, red-gray and green-gray, beds 2-10cm.   4.0m

Dolomite marble, calcareous, tan-weathering, with sparse blue-
green and brown calc-silicate stringers.   1.5m

Quartzite, fine-grained, massive to bedded, light green-gray and
red-gray.   2.5m

Dolomite marble, light tan.   .5m

Quartzite, fine-grained, light gray, light greenish-gray, and
light red-gray, platy to thick bedded with sparse calc-silicate
stringers. Beds 2-20cm thick.   2.0m

Calc-silicate rock, green and tan, with local calcite crystals
and associated green, dark, acicular mineral (actinolite?) and
white, fibrous mineral (tremolite?).   .1m

Quartzite, greenish, fine-grained.   .9m

Calc-silicate rock, crumbly, green and light red.   2.0m

Quartzite, fine-grained, greenish to reddish gray, with sparse
calc-silicate stringers, gray-green in color with actinolite(?).   2.5m

Quartzite, fine-grained, red-orange to dusty white-orange, beds
5-15cm thick.   4.5m

Quartzite, fine-grained, platy, thinly laminated, gray, crumbly.   2.0m

Covered.   5.0m

TOP OF WOOD CANYON FORMATION

White, massive, orthoquartzite.

ZABRISKIE QUARTZITE, MEASURED SECTION #2B

                                                             Thickness

Quartzite, fine- and medium-grained, with beds 2-8cm thick,
locally crumbly, greenish-gray, medium to light brown, orange,
and dark red. Greenish beds may be calc-silicate layers.
Generally weathers into plates, not very resistant.   3.0m

BASE OF ZABRISKIE QUARTZITE   (Total Thickness - 50.9m)

Quartzite, medium-grained, orange- and orange-brown-, black-
weathering.   .5m

Orthoquartzite, coarse-grained, dark pink, faintly bedded.   8.5m

Orthoquartzite, coarse- to medium-grained, very faintly bedded, light pink to white, with very sparse, thin, micaceous laminations.   23.0m

Quartzite, medium-grained, dark gray-brown.   .3m

Orthoquartzite, coarse- to medium-grained, faintly bedded, pink to white, with sparse, thin micaceous laminations more apparent in upper part.   9.6m

Quartzite, medium-grained, tan-brown.   .5m

Orthoquartzite, medium- to fine-grained, faintly bedded pink to white, with local brown streaks and diffuse laminations and layers.   8.5m

TOP OF ZABRISKIE QUARTZITE

Quartzite, crumbly orange-red, micaceous, with sparse beds of pink orthoquartzite 5-20cm thick.   1.0m

CARRARA FORMATION, MEASURED SECTION #3

                                               Thickness

Orthoquartzite, pink, coarse-grained.   .8m

BASE OF CARRARA FORMATION   (Total Thickness - 97.1m)
BASE OF LOWER MEMBER, CARRARA FORMATION   (Total Thickness - 26.2m)

Argillaceous quartzite, black, gray, or brown, fine-grained, micaceous, parts on bedding, poorly resistant.   .8m

Orthoquartzite, pink, coarse-grained.   .6m

Argillaceous quartzite, black, gray, or brown, fine-grained, micaceous, cleaves on argillaceous layers to form plates 1-4cm thick. Locally laminated with laminations 1-2mm thick.   1.9m

Covered.   3.4m

Argillaceous quartzite, black, gray, or brown, fine-grained, faintly bedded, no laminations.   1.7m

Argillaceous quartzite, dark gray to brown, micaceous, laminated, platy to blocky weathering.   1.7m

Covered.   1.7m

Quartzite, orange, fine-grained, slightly calcareous(?), massive
to thick-bedded.   2.5m

Covered.   4.2m

Quartzite, orange, fine-grained, massive to thick-bedded.   .8m

Covered.   4.2m

Covered.   2.7m

BASE OF MIDDLE MEMBER, CARRARA FORMATION   ˙ (Total Thickness - 34.0m)

Calcite marble, medium to light gray, slightly silty, diffusely
laminated.   8.2m

Calcite marble, silty, diffusely layered, crumbly.   1.8m

Calcite marble, medium to light gray, diffusely laminated,
slightly silty, layers .5-4cm thick, platy to blocky weathering.   6.8m

Calcite marble, medium to light gray, massive to faintly bedded.   3.9m

Covered.   3.9m

Calcite marble, blue-gray to brown-gray, diffusely layered and
laminated.   9.4m

BASE OF UPPER MEMBER, CARRARA FORMATION   (Total Thickness - 36.9m)

Argillaceous quartzite, black to brown, fine-grained, micaceous,
parting and weathering on beds 3-15cm thick. Less resistant
than underlying carbonate.   6.8m

Covered.   1.7m

Argillaceous quartzite, spotted, platy, micaceous, locally with
green-tan calc-silicate layers up to 20cm thick.   .9m

Covered.   .9m

Calcite marble, dusty brown-gray to faintly greenish-gray,
silty, diffusely laminated and layered.   .9m

Argillaceous quartzite, spotted, platy, micaceous, locally with
green-tan calc-silicate layers.   3.4m

Argillaceous quartzite, slightly calcareous, platy, micaceous.   1.7m

Covered.   7.6m

Quartzite, fine-grained, with slight calc-silicate mineraliza-
tion indicated by faint green color.   1.3m

Covered.   2.0m

Quartzite, tan, very fine-grained, with brown, recess form-
ing calc-silicate layers 1cm thick.   1.3m

Covered.   .7m

Calcite marble, silty, greenish-gray to tan, with parting on
tan calc-silicate layers up to 1cm thick.   1.3m

Quartzite, very fine-grained, with calc-silicate laminations.   .7m

Calcite marble, silty, greenish-gray to tan.   .7m

Quartzite, fine-grained, gray to brown, laminated, spotted.   5.0m

TOP OF UPPER MEMBER, CARRARA FORMATION
TOP OF CARRARA FORMATION
BASE OF BONANZA KING FORMATION

BONANZA KING, MEASURED SECTION #4

                                                                      Thickness

Quartzite, very fine-grained, crumbly, green-gray, with thin
white and brown laminations. May actually be a siltstone.   .5m

Quartzite, medium-grained, massive, pink to white.   1.5m

Calc-silicate rock and calcite marble, thinly bedded in layers
2-5cm thick. Calc-silicate rock is gray-green, fissile. Mar-
ble is platy, laminated, and dull gray-brown in color.   1.0m

Quartzite, fine-grained, laminated, dark gray-white.   1.0m

Covered.   1.0m

BASE OF BONANZA KING FORMATION   (Total Thickness - 408.0m)
BASE OF PAPOOSE LAKE MEMBER   (Total Thickness - 172.1m)
BASE OF UNIT #1   (Total Thickenss - 23.7m)

Calcite marble and siliceous calcite marble, laminated, with wavy
or crinkley texture. Layers and laminations generally are up to
1cm thick with one siliceous layer 8cm thick. Boudinage apparent
in some 1cm siliceous layers.   5.0m

Calc-silicate rock, with calcite crystals to 3mm suspended in a
fissile silt or calc-silicate matrix. Sparse 2-5cm thick sili-
ceous calcite marble layers, and one calc-silicate layer, 5cm
thick, with green acicular crystals up to 1cm long.   .5m

Calcite marble, platy weathering, light gray, coarse to fine crystals.   3.5m

Covered.   2.0m

Calcite marble and calc-silicate rock. Calcite marble is platy weathering, dusty gray-brown, and of highly variable crystal size. Silty layers are crumbly, up to 30cm thick, and may contain sand.   5.0m

Calcite marble, fissile to platy in layers 1-5cm. Platy layers protrude from siltier layers. Individual crystals stand out on weathered surfaces of calcite layers - resistant dolomite(?) crystals.   1.9m

Calcite marble, diffusely bedded, fissile, with local calcite crystals to .5cm. This bed is base of cliff forming Bonanza King Formation.   1.0m

Calcite marble, gray and white laminations approximately .5cm thick. Slightly platy. Laminations have stringy or irregular shape. Gray layers may be partially dolomite.   1.9m

Calcite marble, coarse-grained, fissile, diffusely bedded and laminated, slightly silty, with sparse cross-beds(?) 20cm thick.   1.4m

Calcite marble, dirty light gray-brown, bedded, slightly crumbly, probably slightly silty.   2.4m

BASE OF UNIT #2   (Total Thickness - 48.4m)

Calcite marble, laminated gray and white ("zebra stripe") laminations typically .5cm thick, relatively pure (i.e. no silt), blocky weathering.   2.9m

Calcite marble, laminated (.5cm thick) gray and white, slightly mottled, relatively pure (i.e. no silt), blocky weathering.   4.0m

Calcite marble, laminated (.5cm thick) gray and white, well mottled, relatively pure, blocky weathering.   1.5m

Calcite marble, very mottled gray and white, layered marble. White calcite is coarse with crystals up to .4cm. Gray calcite is blue-gray on fresh surfaces, dull blue or brown-gray on weathered surfaces.   5.4m

Calcite marble, well layered, 2-5cm layers of gray and white marble ("zebra stripe"). Moderately mottled.   5.9m

Calcite marble, "zebra stripe" as in underlying bed, but layers thinner (.3-2cm) and more mottled.   2.0m

Calcite marble, as in underlying bed, but fewer gray layers.   3.0m

Calcite marble, white with irregular gray stringers and sparse, wavy, light brown silty stringers. 2.0m

Calcite marble, layered gray and white. 1.0m

Calcite marble, light and medium gray stringy and mottled laminations. Laminations weather differentially to form recesses and protrusions. 3.7m

Calcite marble, faintly laminated to massive. 2.7m

Calcite marble, light gray with dark gray-brown laminations and thin beds .5-2cm thick, slightly mottled. .9m

Calcite marble, as in underlying bed, but thinner beds and laminations (.5-1cm) and less mottled. 1.0m

Covered. 1.0m

Calcite marble, dark and light gray, mottled layers up to 3cm thick. 1.5m

Covered. .5m

Calcite marble, very crumbly, chalky, massive to faintly layered, pink-white to tan-white, possibly silty. .5m

Calcite marble, laminations, stringers, and thin beds, medium and dark gray. 1.9m

Calcite marble, faintly laminated, slightly mottled, gray. 4.3m

Calcite marble, slightly platy weathering, laminated gray to light gray. 1.9m

Calcite marble, weathers along bedding to form blocks, very light gray to very light tan. .9m

BASE OF UNIT #3   (Total Thickness - 13.9m)

Dolomite marble, weathers on bedding planes to form blocks. One 5-10cm thick white calcite bed near top. 1.9m

Dolomite marble, massive, tan-weathering, white on fresh surfaces, slope forming. 6.6m

Calcite marble, white on fresh and weathered surfaces, very coarsely crystalline (crystals to .5cm), massive. Weathers into irregular chunks. Slightly fissile near top. Grades upward over 30cm into next unit. 5.4m

BASE OF UNIT #4   (Total Thickness - 38.0m)

Calcite marble, diffusely laminated and slightly mottled, layers
dark gray and light gray to white.   2.2m

Covered.   .7m

Siliceous(?) calcite marble, very crumbly, silty, burrowed(?).   .2m

Calcite marble, laminations .3-1cm thick are slightly diffuse
and mottled, typically bifurcating and converging with a slightly
wavy or stringy texture. About 50% light and 50% dark gray
laminations.   3.7m

Calcite marble, crumbly recess forming bed, white to gray-
white to tan-white, with 2-5cm thick, protruding marble layers.   1.0m

Calcite marble, diffusely laminated and thinly bedded, with
layers up to 2cm thick, slightly to moderately mottled, medium
gray.   2.5m

Calcite marble, platy to crumbly, recess forming, silty(?),
fissile, with 1-5cm resistant gray marble "ribs". Dusty white
to pink to light brown.   1.0m

Calcite marble, medium gray to blue-gray, with light brown,
grainy, protruding, stringy or mottled layers and laminations
up to 2cm thick but may be amalgamated into 30cm thick layers.   14.9m

Covered.   .8m

Calcite marble, silty, crumbly, recess forming, dusty pink-
brown, weathers into plates and blocks with some very crumbly
silty material between plates.   .2m

Calcite marble, variably stringy laminations and thin beds up
to 2-3cm thick, slightly to moderately mottled. Grades upward
from medium to light gray.   2.4m

Calcite marble, sparsely laminated and layered, light gray to
white, coarsely crystalline, with sparse stringers and blebs.   1.4m

Calcite marble, light gray to white with medium gray layers
and stringers up to 5cm thick.   1.0m

Calcite marble, well-layered or bedded, with gray and light
tan to white layers typically 1-4cm thick. Gray layers are
medium to dark gray on weathered surfaces, light gray on
fresh surfaces. One 10cm thick gray layer with 10cm thick
white layers above and below.   4.8m

Interbedded calcite marble and calcareous dolomite marble. White
calcite marble with tan, calcareous dolomite stringers and layers
up to 8cm thick. This is a distinctive marker bed.   1.4m

BASE OF UNIT #5   (Total Thickness - 48.1m)

Dolomite marble, massive, tan on weathered surfaces, white
on fresh surfaces.   6.1m

Interbedded calcite and dolomitic calcite marble. Calcite
marble is dull white-weathering, coarsely crystalline. Dolo-
mitic calcite marble is light tan-weathering. Interbedded with
beds 10-40cm with sparse, faint stringers. All fresh surfaces
are white.   2.8m

Calcite marble, crumbly, silty(?), weathers to plates, chunks,
and dirt. Sparse, resistant, 2-5cm thick protruding layers.   .5m

Calcite marble, white to light brown with stringers and ir-
regular blebs of light tan calcite marble.   4.2m

Calcite marble. As in underlying beds, but with mottled, gray
stringers.   .9m

Calcite marble, white and light gray, with diffuse laminations
and stringers. Weathers on sparse bedding planes 5-40cm apart
and 1-2cm thick to form blocky talus. May be silty(?).   3.7m

Covered.   .9m

Calcite marble, slightly crumbly, light gray to light tan, with
faint stringers and laminations. Sparse 1cm thick planar re-
cesses help define bedding, some fracturing on these planes.   2.3m

Covered.   .9m

Calcite marble. As in underlying beds, but with fainter
stringers and laminations.   1.4m

Calcite marble. Bedded with 2-10cm medium to light gray and
tan-white beds. With sparse light tan stringers.   .5m

Calcite marble, light gray and white in diffuse and faint beds
5-30cm. With tan-weathering, mottled stringers and blebs up
to 2cm thick.   .9m

Calcite marble, light gray-brown, with faint stringer like
recesses but otherwise massive. Weathers on bedding planes
to form blocks.   2.3m

Calcite marble, crumbly, light tan to light gray, bedding faint.   .9m

Calc-silicate rock and calcite marble. Calc-silicate rock is
very crumbly with calcite crystals suspended in tan-weathering,
crumbly, silty, matrix. Calc-silicate beds .4-4cm thick, planar,
with little mottling. Calcite marble light gray to dusty
white.   .9m

Calcite marble, massive, white, with faint light tan stringers and blebs of silty, very crumbly calc-silicate rock, locally forming up to 30% of rock.   1.4m

Calcite marble. Diffusely to sharply laminated and layered light gray marble with tan-weathering, slightly mottled silty layers .5-2cm thick.   1.4m

Calcite marble, white-gray, crumbly, coarse-grained, with poorly defined bedding.   .5m

Calcite marble, massive, tan-weathering.   .9m

Covered.   .9m

Calcite marble, faintly laminated, light gray to light tan.   1.2m

Dolomite marble, light brown-weathering, white on fresh surfaces, massive.   1.1m

Calcite marble, light brown-weathering, white on fresh surfaces. Weathers on bedding planes to form blocks 5-50cm thick, massive within blocks.   3.6m

Calcite marble, faintly laminated and layered, medium blue-gray, may be slightly silty.   3.6m

Calcite marble, medium-light blue-gray and white, with diffuse, darker stringers and lamella.   3.6m

BASE OF BANDED MOUNTAIN MEMBER
BASE OF UNIT #6

Covered.   .8m

Calc-silicate rock and calcite marble. Calc-silicate rock is very crumbly, silty, with suspended calcite crystals in a silty matrix, in layers 1-5cm thick. Thinly interbedded with thin layers of calcite marble.   .2m

Calcite marble, bedded to locally laminated with diffuse silt.   .2m

Calcite marble, grainy, fissile, diffusely bedded.   .5m

Calcite marble, light to medium gray, faintly laminated and stringy.   .8m

Calc-silicate rock. Poorly cemented, poorly layered, light brown or gray rock, with easily disaggregated calcite crystals in a silty matrix.   1.2m

Calcite marble, gray and white, laminated to mottled.   .4m

Calc-silicate rock, as in underlying bed, weathers brown to red
to gray, with 1-5cm thick protruding layers of gray or reddish
brown calcite marble. One 5cm thick marble is structurally
disaggregated possibly as a result of incipient boudinage.　　　3.1m

Calcite marble, gray, with diffuse tan-weathering silty string-
ers and bands. Weathers on bedding to form plates and blocks.　　1.9m

Fault.　　　　　　　　　　　　　　　　　　　　　　　　Total 9.1m

BONANZA KING, MEASURED SECTION #6

                                                              Thickness

BASE OF UNIT #6   (Total Thickness -   8.0m)

Calc-silicate rock. Silty calcite marble, recess forming.   .5m

Calc-silicate rock. Variably silty calcite marble, laminated,
light gray.   1.5m

Calc-silicate rock, platy weathering, fissile, orange and green. 1.5m

Calc-silicate rock, laminated, light brownish gray- to green-
weathering, silty. Weathers on bedding to form blocks and
locally, plates.   1.0m

Calc-silicate rock, fissile and recess forming, with locally
developed, silty, protruding, calcite marble beds 1 to 3cm thick.   .2m

Calcareous dolomite marble, silty, thinly laminated, light
brown, or locally gray-green-weathering. Brown-weathering,
protruding, siliceous laminations are .1-.3cm thick. Weathers
on sparse bedding planes to form blocky outcrops.   3.3m

Calcareous dolomite marble, silty, platy weathering.   .1m

BASE OF UNIT #7

Dolomite marble, light brown-weathering, thinly laminated, light
to white-gray on fresh surfaces. Brown- or light green-weathering
laminations are .1 to .5cm thick.   .9m

Dolomite marble, light brown-weathering, massive to faintly
laminated.   3.0m

Dolomite marble, bedded to thickly laminated, light to medium
brown-weathering, with dark brown siliceous laminations .3 to
1cm thick.

BONANZA KING, MEASURED SECTION #5

                                                                    Thickness

BASE OF BANDED MOUNTAIN MEMBER   (Total Thickness - 235.9m)
BASE OF UNIT #6   (Total Thickness - 19.6m)

Section starts at top of last thick gray limestone below siltstone.

Covered.   .6m

Calc-silicate rock. Crumbly, fissile, grainy, silty, dusty
white-gray calc-silicate rock with matrix supported calcite
crystals. With 1cm thick protruding calcite layers.   .3m

Calcite marble, light to medium gray, platy to blocky weather-
ing.   .9m

Calcite marble, gray, blocky weathering with sparse bedding
plane fractures. Also with faint stringers and lamella of
slightly brown-weathering, recess forming silty layers.   .5m

Covered.   .8m

Calcite marble, very crumbly, grainy, dusty white-gray, probably
silty.   .2m

Calcite marble, gray, blocky, with faint stringers and lamella
of slightly brown-weathering silty(?) recess forming calcite
marble.   .5m

Covered.   2.2m

Calcite marble, very crumbly, grainy, dusty white-gray,
probably silty.   .2m

Calcite marble, light gray-brown with 1cm long tan-orange blebs.   .2m

Covered.   3.1m

Calcite marble, very crumbly, grainy, dusty white-gray, probably
silty.   .2m

Calcite marble, weathers light pinkish-brown to light gray to
dull tan-brown. Weathers into small slightly to moderately
irregular blocks. Brown-weathering siliceous laminations and
stringers, forming 3-30% of rock, form protrusions and recesses
and are variably mottled.   3.4m

Covered.   .2m

Calcite marble, light brown-weathering, with brown-weathering
laminations and layers up to 2cm thick. Local weathering on
bedding planes. A few 5cm thick, silty, white, very crumbly
layers.   2.3m

Covered.   .3m

Calcite marble, light gray-brown, coarse-grained, with poorly
defined bedding.   .6m

Calc-silicate rock, light greenish-white, crumbly.   .2m

Calcite marble, brown- to dusty brown-red-weathering, with
light green- or brown-weathering cherty nodules and bands.
Light green bands may be fine-grained quartzite, in beds up
to 20cm.   2.2m

BASE OF UNIT #7   (Total Thickness - 35.1m)

Covered.   6.0m

Dolomite marble, massive, tan-orange to tan-gray, internally
brecciated.   12.7m

Dolomite marble, tan-orange, with faint, deeper orange-weathering,
layers and laminations.   2.9m

Dolomite marble, massive, tan-orange to tan-gray, internally
brecciated.   3.9m

Dolomite marble, massive, tan-orange to tan-gray, internally
brecciated, with sparse stringers and layers of more orange
to dark brown siliceous layers.   1.9m

Dolomite marble, medium brown-orange, siliceous(?), hard,
slightly calcareous, internally shattered.   .4m

Dolomite marble, tan-orange to tan-gray, internally brecciated,
with faint, slightly deeper orange stringers and laminations.
Slightly variable shades of orange define vague beds more than
1m thick. Sharp contact with overlying, light tan dolomite
marble.   7.4m

BASE OF UNIT #8   (Total Thickness - 24.8m)

Dolomite marble, light tan-gray to white, splotchy, fractures
and faint laminations define bedding.   5.7m

Dolomite marble, dusty gray to white, weathers on bedding planes
to form blocky talus, sparse white-yellow blotchy siliceous
deposits.   6.8m

Covered.   5.2m

Dolomite marble, massive, light dusty tan-gray to white, coarse-grained.   3.5m

Covered.   3.5m

BASE OF UNIT #9   (Total Thickness - 72.3m)

Calcite marble, white, with sparse silty laminations up to .5cm thick. Also sparse, irregular, cherty(?), siliceous blebs.   1.7m

Calcite marble, banded and laminated with layers .4-3cm thick, white to light gray, almost no mottling.   .8m

Covered.   .3m

Calcite marble, white, with light tan-brown, diffuse laminations and bands .3-5cm thick.   1.7m

Covered.   .6m

Calcite marble to dolomitic calcite marble, white to light gray to light brown. Bedding plane fractures and differential weathering define bedding. Thin and diffuse silty(?) and cherty laminations up to 1cm thick.   3.4m

Calcite marble, light tan-brown to light blue-gray, faint laminations, sparse weathering on bedding.   .6m

Calcite marble, laminated and bedded white, blue-gray, gray-brown and dark brown, with protruding siliceous laminations. This bed grades laterally into fine-grain blue-gray limestone with white, lensoidal shaped blebs 2cm across and .2cm thick. Locally abundant, thin (.2-.5cm), protruding siliceous laminations. Rests on undulating top of tan calcite marble. Knots, blebs, and lensoidal shaped siliceous deposits and white calcite marble deposits are suggestive of fossil or bioturbation origin. This is a distinctive marker bed.   .6m

Calcite marble, light gray to gray-white, diffusely laminated.   1.7m

Calcite marble, white-gray, with discontinuous layers of medium gray.   .9m

Calcite marble, white-gray with moderately continuous medium gray layers. Generally irregularly fracturing with local fracturing and weathering on bedding planes.   1.9m

Covered.   8.5m

Calcite marble, thinly laminated, crumbly, blue-gray and white, with small lensoidal blebs of white calcite and bedding plane fractures.   .5m

Calcite marble, laminated, medium blue-gray and white, in
beds 5-20cm thick. Small birds-eye or lensoidal blebs
of white calcite.   1.4m

Calcite marble, moderately laminated with laminations .2-1cm
thick of light gray and white-gray to white. Some fracturing
and weathering on bedding but generally irregularly fracturing.   9.4m

Calcite marble, laminated, light gray, slightly platy.   2.8m

Calcite marble, diffusely laminated, light gray and white-
gray. Weathering may produce recesses along laminations.   1.9m

Covered.   2.8m

Calcite marble, brecciated, white to light gray.   1.2m

Covered.   .8m

Dolomite marble, slightly brecciated, massive, tan-weather-
ing, white on fresh surfaces.   .8m

Covered.   1.6m

Calcite marble, faint laminations and bands up to 2cm thick,
slightly silty, local brecciation.   4.0m

Covered.   1.6m

Dolomitic calcite marble, light brown, brecciated, with
irregular blobs of opaline siliceous material.   1.6m

Covered.   1.6m

Dolomite marble, light brown-weathering, white on fresh
surfaces, bedding defined by etched grooves.   .8m

Calcite marble and dolomitic calcite marble, light gray-
brown with irregular, siliceous, protruding, blebs and
stringers.   1.6m

Calcite marble, slightly fissile, white, bedding defined by
weathering, platy jointing, and faint laminations. Locally
well-bedded to laminated.   2.4m

Calcite marble, light gray, faint to medium laminations,
with some bedding plane weathering and fracturing, weathers
into blocks.   2.0m

Covered.   2.0m

Calcite marble and dolomite marble, thinly interbedded. Calcite marble is white, dolomite marble is light brown. Beds up to 10cm, may contain or appear as discontinuous stringers or lamella.   1.4m

Dolomite marble, light brown, massive appearing but with etch-weathering in discontinuous stringers and lamella.   .7m

Covered.   1.1m

Calcareous dolomite marble, coarse-grained, medium brown.   .4m

Covered.   1.1m

Calcite marble, diffusely laminated, light gray to light brown.   1.4m

BASE OF UNIT #10

Dolomite marble, light tan-brown, massive except for faint, aligned weathering pits that define bedding.

BONANZA KING, MEASURED SECTION #7

                                                                    Thickness

Dolomite marble, massive, light tan-weathering, white on fresh surfaces, coarsely crystalline. Generally irregularly fracturing with sparse fracturing on bedding planes.   17.0m

Dolomite marble, massive light tan-weathering, white on fresh surfaces, coarsely crystalline, with brown-weathering siliceous stringers and blebs.   3.4m

BASE OF UNIT #9   (Total Thickness - 89.6m)

Calcareous dolomite marble, slightly fissile, coarse-grained, white-tan.   .7m

Calcite marble, light gray to white.   .2m

Calcite marble to dolomitic calcite marble, slightly fissile, coarse-grained. Etched grooves on weathered surfaces define bedding.   .6m

Covered.   2.7m

Dolomite marble, massive, light tan-weathering, white on fresh surfaces. Bedding defined by etched grooves on weathered surfaces and by sparse bedding plane fractures.   1.4m

Dolomite marble, massive, light brown.   3.4m

Covered.   2.7m

Dolomitic calcite marble, light brown to light gray, crumbly.   2.0m

Calcite marble, white to white-gray, coarse-grained, with minor
light brown, laterally discontinuous dolomite marble beds up
to 5cm thick.   .5m

Calcite marble, interbedded light brown, medium gray, white,
and blotchy white and gray beds 5-20cm thick.   1.0m

Calcite marble, light gray to white, massive, irregularly
fracturing.   .7m

Calcite marble, light gray and white, with laminations, swirls,
speckles and blebs.   .7m

Calcite marble, light gray and white, faint, discontinuous, tan-
weathering bands and laminations define weathering.   2.0m

Calcite marble, tan-white- to white-weathering, locally gray,
with tan laminations.   1.4m

Calcite marble, diffuse tan- and brown-weathering bands 10cm
thick with rounded, brown-weathering, siliceous(?), blobs up
to 10cm diameter.   .3m

Calcite marble.   .3m

Dolomitic calcite marble to calcareous dolomite marble, light
tan-weathering. Diffuse to distinct, sparse, brown-weathering
laminations are white to light gray on fresh surfaces. Inter-
bedded with light gray to white-gray calcite marble, beds 1-2cm.
Also dark brown, continuous, protruding siliceous laminations
and stringers .3-1cm thick make up 1% of rock.   8.0m

Covered.   .8m

Dolomitic calcite marble, light tan-weathering.   .4m

Calcite marble, coarsely crystalline, white, massive, ir-
regularly fracturing.   1.3m

Dolomitic calcite marble, light tan-weathering, with distinct
to diffuse, brown-weathering, planar, continuous laminations,
less than 1cm thick.   1.7m

Dolomitic calcite marble, massive, light brown-weathering.   .6m

Calcite marble, white, coarse-grained, with irregular fractures,
and light brown-weathering laminations forming 5% of the rock.   2.3m

BASE OF MORE CLIFF FORMING MIDDLE SUBUNIT OF UNIT #9

Calcite marble, very well-laminated, laminations .5-1cm thick, white, light medium or dark gray, not mottled at all.   2.1m

Covered.   .4m

Calcite marble, massive, light gray-brown-weathering, light gray on fresh surfaces.   .8m

Calcite marble, coarse-grained, medium dark gray and white. Beds 5-20cm, sparse laminations .5-1cm.   .8m

Calcite marble, interbedded and diffusely and sparsely inter-laminated, light brown, light gray, and white-weathering, calcite marble. Beds 5-40cm thick. Laminations slightly wavy and mottled, are typically light brown-weathering and slightly protruding, surrounded by white or light gray.   3.6m

Calcite marble, thinly and finely laminated, medium dark and light gray, white, and light brown laminations, with small, lensoidal, white calcite blebs and sparse 5-10cm eliptical, brown-weathering blobs. This is a distinctive marker bed.   .4m

Calcite marble, fine to diffusely laminated, light and medium dark gray or white, coarse-grained. Some bedding plane weather-ing.   1.8m

Calcite marble, diffusely and extensively laminated, medium to light gray.   .9m

Calcite marble, very diffusely laminated, medium to light gray and light tan-gray.   1.3m

Calcite marble, moderate to finely laminated, light gray to white, sparse fractures on bedding planes.   1.3m

Calcite marble, gray, with faint laminations and etching along bedding.   .9m

Covered.   1.3m

Calcite marble, gray, brecciated.   1.3m

Calcite marble, white, sparsely laminated.   .4m

TOP OF CLIFF FORMING MIDDLE SUBUNIT OF UNIT #9

Calcite marble, laminated, light gray. Medium gray to brownish-gray, slightly discontinuous and wavy laminations form 20% of the rock.   3.5m

Calcite marble, grainy, gray, laminated, grainy layers (dolomi-tic?) .3-2cm thick, locally amalgamated into grainy bands. Some mottling.   .9m

Covered.   1.8m

Dolomite marble, massive, light brown-weathering.   .4m

Calcite marble, medium gray and white, laminated and bedded.   .9m

Calcite marble, laminated (.5-2cm), grainy, medium gray, brown,
and white. Well-layered. Grainy layers weather light brown,
protrude slightly. Layers are laterally persistant but locally
broken by boudinage. Boudin necks contain soft, fine-grained,
green material.   2.2m

Dolomite marble, gray, mottled, with sparse white laminations
and layers of grainy dolomite marble up to 2cm thick.   1.8m

Covered.   .9m

Dolomitic calcite marble, light brown-weathering, with one
discontinuous, 3cm-thick, protruding, dark brown and green layer.   .9m

Covered.   1.8m

Calcite marble, massive, light brown- to white-weathering,
coarse-grained.   .9m

Calcareous dolomite and dolomite marble, medium to light brown,
faintly layered and laminated. Light gray-brown on fresh sur-
faces.   1.4m

Covered.   .9m

Calcite marble and calcareous dolomite marble, interbedded in
beds 5-10cm thick. Calcite marble weathers white to tan with
dark brown blobs 3-4cm across. Dolomite weathers light tan.   .2m

Covered.   3.0m

Calcite marble, white, with faint, light brown and gray
laminations.   .9m

Calcite marble, calcareous dolomite marble and dolomite marble,
interbedded and interlaminated, laminations wavy and stringy,
beds up to 15cm thick. Calcite marble is gray- and white-
weathering, dolomite marble and calcareous dolomite marble are
brown-weathering.   1.8m

Covered.   .9m

Calcite marble, dolomite marble, and calcareous dolomite marble,
interbedded and interlaminated with laminations typically .5-
1cm thick and beds up to 1cm thick.   8.2m

Dolomite marble, medium brown, with dark brown, irregular sili-
ceous layers and stringers.   .4m

Calcite marble and dolomite marble, interbedded and inter-
laminated. Calcite marble is white, dolomite marble is light
brown.   2.1m

<u>BASE OF UNIT #10</u>   (Total Thickness - 66.8m)

Dolomite marble, massive, light brown-weathering, white on fresh
surfaces, with very sparse white calcite marble beds up to
20cm thick. Also sparse calc-silicate layers 2-4cm thick with
local boudinage.   5.8m

Dolomite marble, massive, tan, coarse-grained, white on fresh
surfaces.   13.9m

Dolomite marble, massive, tan, coarse-grained, white on fresh
surfaces, with sparse calc-silicate stringers and nodules.   19.8m

Dolomite marble, massive, tan, coarse-grained, white on fresh
surfaces, with sparse brown-weathering siliceous blobs, 2-5cm
diameter.   16.4m

Dolomite marble, massive, tan, coarse-grained.   10.8m

TOP OF BONANZA KING FORMATION
TOP OF BANDED MOUNTAIN MEMBER
TOP OF UNIT #10

BIBLIOGRAPHY

Barnes, H., and Palmer, A. R., 1961, Revision of Stratigraphic nomenclature of Cambrian rocks, Nevada Test Site and vicinity, Nevada: U. S. Geological Survey Professional Paper 424-C, p. C100-C103.

Bereskin, S. R., 1976, The Sultan Limestone of Northeastern San Bernardino County (California) and Clark County (Nevada): Geological Society of America Abstracts with Programs, v. 8, n. 3, p. 355-356.

Burchfiel, B. C., and Davis, G. A., 1971, Clark Mountain Thrust Complex in the Cordillera of Southeastern California: Geologic summary and field trip guide: University of California, Riverside, Campus Museum Contributions, v. 1, p. 1-28.

Burchfiel, B. C., and Davis, G. A., 1972, Structural Framework and Evolution of the Southern Part of the Cordilleran Orogen, Western United States: American Journal of Science, v. 272, p. 97-118.

Burchfiel, B. C., and Davis, G. A., 1975, Nature and Controls of Cordilleran orogenesis, western United States: Extension of an earlier synthesis: American Journal of Science, v. 275-A, p. 363-396.

Burchfiel, B. C., and Davis, G. A., 1977, Geology of the Sagamore Canyon-Slaughterhouse Spring Area, New York Mountains, California: Geological Society of America Bulletin, v. 88, p. 1623-1640.

Burchfiel, B. C., and Davis, G. A., 1981, Mojave desert and environs, in Ernst, W. G., editor, The Geotectonic Development of California; Prentice-Hall, New Jersey, p. 217-252.

Burchfiel, B. C., Fleck, R. J., Secor, D. T., Vincelette, R. R., and Davis, G. A., 1974, Geology of the Spring Mountains, Nevada: Geological Society of America Bulletin, v. 85, p. 1013-1022.

Burchfiel, B. C., Hamill, G. S. IV, and Wilhelms, D. E., in preparation, Structural Geology of the Montgomery Mountains and the North Half of the Nopah and Resting Spring Ranges, Nevada and California.

Cornwall, H. R., and Kleinhampl, F. J., 1961, Geology of the Bare Mountain Quadrangle, Nevada: U. S. Geological Survey Map GQ-157.

Davis, G. A., and Burchfiel, B. C., 1973, Garlock fault of southern California - an intercontinental transform structure: Geological Society of America Bulletin, v. 84, p. 1407-1422.

DeWitt, E., 1980, Geology and Geochronology of the Halloran Hills, Southeastern California, and implications concerning Mesozoic Tectonics of the southwestern Cordillera: Pennsylvania State University, University Park, Pennsylvania, Ph.D. thesis, 267 p.

Dibblee, T. W., Jr., 1967, Areal Geology of the western Mojave Desert, California: U. S. Geological Survey Professional Paper 522, 153 p.

Dickinson, W. R., 1981, Plate tectonics and the continental margin of California, in Ernst, W. G., editor, The Geotectonic Development of California, Rubey Volume I, p. 1-28.

Gans, W. T., 1974, Correlation and redefinition of the Goodsprings Dolomite, southern Nevada and eastern California: Geological Society of America Bulletin, v. 85, p. 189-200.

Hall, W. E., and MacKevett, E. M., Jr., 1962, Geology and Ore Deposits of the Darwin Quadrangle, Inyo County, California: U. S. Geological Survey Professional Paper 368, 87 p.

Hamilton, W., 1969, Mesozoic California and the underflow of the Pacific mantle: Geological Society of America Bulletin, v. 80, p. 2409-2430.

Hamilton, W., and Meyers, W. B., 1966, Cenozoic Tectonics of the Western United States: Reviews of Geophysics, v. 4, p. 509-549.

Hazzard, J. C., 1937, Paleozoic section in the Nopah and Resting Springs Mountains, Inyo County, California: California Division of Mines, Journal of Mines and Geology, v. 33, n. 4, p. 273-339.

Hazzard, J. C., 1954, Rocks and structure of the northern Providence Mountains, San Bernardino County, California, in Jahns, R. H., editor, Geology of southern California: California Division of Mines and Geology Bulletin 170, part 4, p. 27-35.

Hazzard, J. C., and Mason, J. F., 1936, Middle Cambrian formations of the Providence and Marble Mountains, California: Geological Society of America Bulletin, v. 47, p. 229-241.

Hewett, D. F., 1931, Geology and ore deposits of the Goodsprings quadrangle, Nevada: U. S. Geological Survey Professional Paper 162, 172 p.

Hewett, D. F., 1956, Geology and mineral resources of the Ivanpah quadrangle, California and Nevada: U. S. Geological Survey Professional Paper 275, 172 p.

Jahns, R. H., Troxel, B. W., Wright, L. A., 1971, Some structural implications of a Late Precambrian-Cambrian section in the Avawatz Mountains, California: Geological Society of America Abstracts with Programs, v. 3, n. 2, p. 140.

Kupfer, D. H., 1960, Thrustfaulting and chaos structure, Silurian
    Hills, San Bernardino County, California: Geological Society
    of America Bulletin, v. 71, p. 181-214.

Lanphere, M. A., 1964, Geochronologic studies in the eastern Mojave
    Desert, California: Journal of Geology, v. 72, n. 4, p. 381-
    399.

Lanphere, M. A., Wasserburg, G. J. F., and Albee, A. L., 1963, Redis-
    tribution of strontium and rubidium isotopes during metamorphism,
    World Beater Complex, Panamint Range, California: Isotopic and
    Cosmic Chemistry, North-Holland Publishing Company, p. 269-320.

McCulloh, T. H., 1952, Geology of the southern half of the Lane
    Mountain Quadrangle, California: University of California,
    Los Angeles, Ph.D. thesis, 182 p.

McCulloh, T. H., 1960, Geologic map of the Lane Mountain Quadrangle,
    California: U. S. Geological Survey open file map, scale
    1:48,000.

Miller, E. L., 1977, Geology of the Victorville region, California:
    Rice University, Houston, Texas, Ph.D. thesis, 226 p.

Miller, E. L., Burchfiel, B. C., and Carr, M. D., 1979, Enigmatic
    Metasedimentary Rocks of the Western Mojave Desert, California
    and Their Relationship to Cordilleran Miogeocline/Platform
    Rocks: Geological Society of America Abstracts with Programs,
    v. 11, n. 3, p. 92.

Nolan, T. B., 1929, Notes on the stratigraphy and structure of the
    northwest portion of the Spring Mountains, Nevada: American
    Journal of Science, v. 217, p. 461-472.

Novitsky-Evans, J. M., 1978, Geology of the Cowhole Mountains, Cali-
    fornia: Rice University, Houston, Texas, Ph.D. thesis, 102 p.

Palmer, A. R., and Halley, R. B., 1979, Physical Stratigraphy and
    Trilobite Biostratigraphy of the Carrara Formation (Lower and
    Middle Cambrian) in the southern Great Basin: U. S. Geological
    Survey Professional Paper 1047, 131 p.

Palmer, A. R., and Hazzard, J. C., 1956, Age and correlation of Corn-
    field Springs and Bonanza King Formations in southeastern Cali-
    fornia and southern Nevada: American Association of Petroleum
    Geologists Bulletin, v. 46, p. 2494-2499.

Poole, F. G., 1974, Flysch deposits of the Antler foreland basin,
    western United States, in Dickinson, W. R., editor, Tectonics
    and Sedimentation: Society of Economic Paleontologists and
    Mineralogists Special Publication n. 22, p. 58-82.

Poole, F. G., and Sandberg, C. A., 1977, Mississippian paleogeography and tectonics of the western United States: Pacific Section, Society of Economic Paleontologists and Mineralogists, Pacific Coast Symposium 2, p. 67-85.

Roberts, M. T., 1976, Stratigraphy and Depositional Environments of the Crystal Spring Formation, southern Death Valley Region, California: California Division of Mines and Geology, Special Report 106, p. 35-43.

Roberts, R. J., Hotz, P. G., Gilluly, J., and Ferguson, H. G., 1958, Paleozoic rocks of north-central Nevada: American Association of Petroleum Geologists Bulletin, v. 42, p. 2813-2857.

Schweickert, R. A., 1976, Shallow-level plutonic complexes in the eastern Sierra Nevada, California, and their tectonic implications: Geological Society of America Special Paper 176, 58 p.

Silver, L. T., 1960, Age determination on Precambrian diabase differentiates in the Sierra Ancha, Gila County, Arizona: Geological Society of America Bulletin, v. 71, n. 12, p. 1973-1974.

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

Silver, L. T., McKinney, C. R., and Wright, L. A., 1961, Some Precambrian ages in the Panamint Range, Death Valley, California: Geological Society of America Special Paper 68, p. 55.

Smith, G. I., and Ketner, K. B., 1970, Lateral displacement on the Garlock fault, southeastern California, suggested by offset sections of similar metasedimentary rocks: U. S. Geological Survey Professional Paper 700-D, p. D1-D9.

Stewart, J. H., 1967, Possible large right-lateral displacement along fault and shear zones in the Death Valley-Las Vegas area, California and Nevada: Geological Society of America Bulletin, v. 78, p. 131-142.

Stewart, J. H., 1970, Upper Precambrian and Lower Cambrian strata in the southern Great Basin, California and Nevada: U. S. Geological Survey Professional Paper 620, 206 p.

Stewart, J. H., and Poole, F. G., 1975, Extension of the Cordilleran Miogeoclinal belt to the San Andreas fault, southern California: Geological Society of America Bulletin, v. 86, p. 205-212.

Stewart, J. H., Albers, J. P., and Poole, F. G., 1968, Summary of regional evidence for right-lateral displacement in the western Great Basin: Geological Society of America Bulletin, v. 79, p. 1407-1414.

Streitz, R., and Stinson, M. C., 1974, Death Valley Sheet, Geologic
    Map of California, 1:250,000, California Division of Mines and
    Geology.

Troxel, B. W., 1967, Sedimentary Rocks of Late Precambrian and Cam-
    brian Age in the southern Salt Spring Hills, southeastern
    Death Valley, California: California Division of Mines and
    Geology Special Report 129, p. 33.

Troxel, B. W., and Butler, P. R., 1979, Tertiary and Quaternary
    fault history of the intersection of the Garlock and Death
    Valley fault zones, southern Death Valley, California: un-
    published report submitted to U. S. Geological Survey, Menlo
    Park, California, 29 p.

Troxel, B. W., and Gunderson, J. N., 1970, Geology of the Shadow
    Mountains and northern part of the Shadow Mountain SE Quadrangles,
    western San Bernardino County, California: California Division
    of Mines and Geology, Preliminary Report 12.

Tyler, D. L., 1975, Stratigraphy and structure of the Late-Precambrian-
    Early Cambrian clastic metasedimentary rocks of the Baldwin Lake
    Area, San Bernardino Mountains, California: M. S. thesis, Rice
    University, 50 p.

Wasserburg, G. J. F., Wetherill, G. W., and Wright, L. A., 1959, Ages
    in the Precambrian terrane of Death Valley, California: Journal
    of Geology, v. 67, p. 702.

Williams, E. G., Wright, L. A., and Troxel, B. W., 1976, The Noonday
    Dolomite and Equivalent Stratigraphic Units, Southern Death
    Valley Region, California: California Division of Mines and
    Geology, Special Report 106, p. 45-59.

Wright, L. A., 1968, Talc deposits of the southern Death Valley-Kingston
    Range region, California: California Division of Mines and
    Geology Special Report 95, 79 p.

Wright, L. A., and Troxel, B. W., 1966, Strata of Late Precambrian-
    Cambrian age, Death Valley Region, California-Nevada: American
    Association of Petroleum Geologists, v. 50, n. 5, p. 846-857.

Wright, L. A., and Troxel, B. W., 1967, Limitations on right-lateral
    strike-slip displacement, Death Valley and Furnace Creek fault
    zones, California: Geological Society of America Bulletin, v. 78,
    n. 8, p. 933-949.

Wright, L. A., and Troxel, B. W., 1970, Summary of regional evidence
    for right-lateral displacement in the western Great Basin: A
    Discussion: Geological Society of America Bulletin, v. 81, p.
    2167-2174.

Wright, L. A., Troxel, B. W., Williams, E. G., Roberts, M. T., Diehl,
    P. E., 1976, Precambrian Sedimentary Environments of the Death
    Valley Region, Eastern California: California Division of Mines
    and Geology, Special Report 106, p. 7-15.

Wrucke, C. T., and Shride, A. F., 1972, Correlation of Precambrian diabase in Arizona and southern California: Geological Society of America Abstracts with Programs, v. 4, n. 3, p. 265-266.

MESOZOIC MAGMATISM, DEFORMATION, AND GEOCHRONOLOGY

INTRODUCTION

Mesozoic calc-alkaline plutonic rocks, and to a lesser degree, volcanic rocks, form the aerially most extensive and abundant pre-Quaternary rock types in the Mojave region (Fig. 1). These rocks are generally thought to represent intrusive and extrusive magmas generated by plate convergence and subduction at an Andean type continental margin (e. g., Hamilton, 1969; Dickinson, 1970). The Mesozoic magmatic arc is superimposed at a high angle across pre-existing facies belts, indicating that major tectonic modification of the continental margin occurred in late Permian or early Triassic time and immediately prior to the inception of plate convergence and magmatic arc development (Fig. 2) (Hamilton and Meyers, 1966; Hamilton, 1969; Burchfiel and Davis, 1972, 1975). The consistently high $^{87}Sr/^{86}Sr$ initial ratios of Mesozoic plutonic rocks in the Mojave region indicate that the Mesozoic magmatic arc developed on or within Precambrian continental crust (Kistler and Peterman, 1973, 1978).

Published ages of Mesozoic igneous rocks in the Mojave region are primarily determined by the K-Ar method. Most of these ages probably represent cooling or uplift ages, and may be partially or completely reset by younger magmatic activity. For these reasons it is difficult to establish the detailed temporal and spacial distribution of arc magmatism or to determine the details of arc evolution. Existent data generally indicate that magmatic activity shifted irregularly with time across the entire Mojave region.

Magmatic activity occurred intermittently during Mesozoic time in the eastern Mojave region, and began during Middle and Late Triassic time along a line extending southeastward from the Inyo Range (Merriam, 1963) through the southern Panamint Range (Johnson, 1957), Soda Mountains (Grose, 1959) and Old Dad Mountains (Dunne, 1977). Deposition of the lower Jurassic Aztec sandstone, without interbedded volcanics, in the eastern and central part of the Mojave region (Dunne, 1977; Novitsky and Burchfiel, 1973; Miller and Carr, 1978) suggests that the earlier period of magmatic activity ended by this time. Magmatism resumed in

Fig. 1. Simplified geologic map outlining all Mesozoic plutonic rocks

in the Mojave and southern Great Basin region. Also shown are Cenozoic

faults (heavy lines) and Mesozoic thrust faults along the eastern side

of the Cordilleran orogen (heavy lines with teeth, dashed where infer-

red). Note that 1) the Avawatz Mountains ("A" on map) are located at

the east edge of the Mesozoic batholith belt, and 2) Mesozoic thrust

faults curve to the south where they approach the batholith belt.

SN = Sierra Nevada; SB = San Bernardino Mountains; G = Granite Mountains;

S = Soda Mountains; A = Avawatz Mountains; O = Owlshead Mountains;

HM = Hunter Mountain pluton; SH = Silurian Hills; HH = Halloran Hills;

W = Wheeler Pass thrust; LC = Lee Canyon thrust; K = Keystone thrust;

C = Contact thrust; MP = Mesquite Pass thrust, Clark Mountains; WP =

Winters Pass thrust, Clark Mountains.

Fig. 1.

Figure 2. Schematic tectonic map of California and
Nevada during Late Triassic to Middle Jurassic time
(modified from Schweickert, 1976, and Burchfiel and
Davis, 1972).

late Early Jurassic and Middle Jurassic time, as indicated by ages of
190 to 160 m.y. (Armstrong and Suppe, 1973; Dunne and others, 1978),
and continued until at least early Late Cretaceous time 80 to 90 m.y.
ago (Burchfiel and Davis, 1981).

Mesozoic folding, thrusting and deformation in the Cordilleran
orogen records a period of east-west crustal compression and shortening.
North of southernmost Nevada, the westward thickening Paleozoic and Pre-
cambrian miogeoclinal wedge formed a strongly anisotropic crustal medium,
dominated by subhorizontal compositional layering, that controlled the
style and geometry of crustal shortening. The miogeocline was thrust
eastward to southeastward onto the craton along thrusts with traces
that generally follow isopach lines within the miogeocline, forming the
Cordilleran fold-thrust belt. Thrusting was generally synchronous with
arc magmatism to the west. The south-southwest trending miogeocline
converges southward with the southeast trending magmatic arc, and the
two meet and cross in the Mojave region so that the magmatic arc extends
southeastward onto the area of the Paleozoic craton (Fig. 2). Where the
fold-thrust belt intersects the magmatic arc in southern Nevada and
southeastern California, the thrusts curve to the southeast and are
parallel to the magmatic arc and at a high-angle to the miogeoclinal wedge
(Fig. 1). The magmatic arc and the associated belt of thrusting and
deformation trend southeast across Precambrian crystalline rocks overlain
by Paleozoic marginal miogeoclinal and cratonal rocks, and thrusting and
deformation involve Precambrian crystalline rocks.

North of southernmost Nevada, the nature and style of crustal
shortening and deformation was controlled by pre-existing shallow crustal
anisotropy in the miogeoclinal wedge, whereas to the south, crustal an-
isotropy resulted from contemporaneous crustal heating and arc magmatism.
Thus, contrasting styles of Mesozoic deformation result from contrasting
controls of lithosphere ductility (Burchfiel and Davis, 1975).

It is a working hypothesis that, in the absence of pre-existing
crustal anisotropy, increased heat flow can trigger yielding and crustal
shortening in a compressive stress field at an Andean type continental
margin (Burchfiel and Davis, 1976). Crustal heating will not trigger
shortening below a certain stress magnitude, and shortening will occur at

very high stresses regardless of thermal effects. Thus, thermal controls of crustal deformation are effective only within a certain range of stress magnitudes. The extent of this range or stress "window" is not presently known.

Evidence for the thermal control or triggering of deformation should be present in the eastern Mojave region where there is no known pre-existing crustal anisotropy. If the stress window over which thermal controls are effective is large, the locus of deformation should migrate in space and time with the locus of magmatism. Our present knowledge of the locus and timing of magmatic and structural events is inadequate to determine if temporal and spacial association is consistent or only sporadic. Constraints on the timing of magmatism are especially poor.

Study of the nature and geochronology of Mesozoic rocks and structures in the Avawatz Mountains was undertaken to further resolve space-time patterns of magmatism, deformation, and uplift in the Mojave region, and to clarify the nature of the relationship between magmatism and deformation.

Methods of study

Besides careful field mapping, samples of plutonic rocks were collected for thin section petrographic analysis, modal mineral analysis of stained slabs, and K-Ar and Rb-Sr geochronologic study.

Slabs of rock were stained for plagioclase and K-feldspar, and a sheet of transparent mylar with a grid of small black dots was placed on each slab. With the aid of a microscope, at least 600 points of quartz and feldspar were counted and the rock was classified according to the scheme proposed by Streckeisen (1973).

Potassium-argon dating was done at the U. S. Geological Survey in Menlo Park. All mineral separations and argon extractions were done by the author. Mass spectrometer analysis was done by the author and by personnel at the U. S. Geological Survey. The grain size of all dated mineral separates fall within the standard seive size range of -45 to +100, and all mineral separates were greater than 99% pure. The Avawatz quartz monzodiorite was the primary rock type dated. It contains no pyroxene, and thus excess argon is not a serious problem due to

pyroxene contamination. Epidote, magnetite, and biotite are probably
the principle contaminants in hornblende mineral separates. Epidote
and magnetite contain insignificant potassium or argon, and thus are
not regarded as serious contaminants. Biotite contamination from
rocks that yield discordant ages will have the effect of slightly re-
ducing the radiometric age, and could have slightly influenced some
samples. Fresh samples were difficult to locate in the Avawatz Mountains.
Thin sections of many of the dated rocks reveal minute cracks through-
out some weathered hornblende grains. The effect of this weathering is
not known, but it possibly causes some argon release and thus reduced
ages. Biotite separates did not have any significant contaminants or
weathering effects.

Rubidium-strontium whole-rock isotope analyses were done by the
author at M.I.T.. Rubidium and strontium concentration were determined
by standard isotope dilution techniques, and all analyses were done on
the NIMA-B mass spectrometer under the supervision of Professor Stan
Hart.

ROCK UNITS

Granite of Avawatz Peak
Leucocratic, pink to white granite crops out along the crest of
the Avawatz Mountains. In the southern half of the range, this rock
can clearly be divided into an older and a younger granite. Granites
in the northern half of the range are mapped as the younger granite on
Plate 1A because they look most like the younger granite, but this
correlation should be considered tentative. This correlation was not
made on Plate 2, and a third, undifferentiated unit is used for north-
ern granite exposures (Fig. 3).

Older granite of Avawatz Peak (Mg_1)
The older granite of Avawatz Peak crops out almost exclusively as
a single elongate body within a fault block between the east and west
branches of the Arrastre Spring fault (Plates 1A, 2). A single small
outcrop, about 100 meters wide, occurs about 300 meters northwest of

Fig. 3. Simplified geologic map of the Avawatz Mountains. Location of all K-Ar samples shown by numbered dots. Sample 132 is located on lower Sheep Creek, upper Sheep Creek is located south of sample 132. Md = Avawatz quartz monzodiorite; Mgm = granodiorite of Mormon Canyon; Mg = granite of Avawatz Peak, undifferentiated; Mmc = quartz monzonite of Cave Spring wash; Tav = Avawatz Formation.

the intersection of the two branches of the Arrastre Spring fault (Plate 1A). This rock is locally intruded by brown fine-grained dikes, and by the younger granite of Avawatz Peak and the quartz monzodiorite complex. It is white to light beige or light greenish white, and contains very few mafic minerals. Modal mineral analysis of two stained slabs indicate that the rock falls well within the granite field (Fig. 4, Table 1).

In thin section, the rock is medium-grained and contains small phenocrysts of microcline. The phenocrysts are subhedral, whereas quartz and andesine (An_{30}-An_{34}) are anhedral. Biotite and magnetite are minor constituents of the rock. Sericite(?) alteration of plagioclase is variable, and the rock is locally crushed and fractured. Secondary calcite may form the matrix and cement of crushed granite.

Age: The older granite is clearly older than the Avawatz quartz monzodiorite, which is at least 176 m.y. old. In the nearby Soda Mountains, shale and limestone of probable Early Triassic age (Moenkopi Formation) overly a calc-silicate sequence containing sparse conglomerates (Grose, 1959). Some conglomerate beds contain perhaps one to two percent metaplutonic clasts (Jon Spencer, unpublished data). This indicates that a small nearby or possibly large distant body of exposed plutonic rock was a source of detritus during late Permian or early Triassic time. It is possible that the older granite of Avawatz Peak, if it was originally a small intrusive body, was the source of this plutonic detritus and thus could be as old as Late Permian. However, the older granite is very similar to the younger granite of Avawatz Peak which intrudes metavolcanic rocks of probable Middle Triassic age or younger. If the two granites are genetically related, as suggested by lithologic similarity, then the older granite is probably Middle Triassic or younger.

Younger granite of Avawatz Peak (Mg_2)

The younger granite of Avawatz Peak crops out over a larger area than the older granite. The younger granite is well-exposed in the southern part of the range where it intrudes the older granite. It is also well-exposed to the north around Avawatz Peak, and northwest of

Fig. 4. Ternary diagrams for modal composition of plutonic rocks. Total quartz + plagioclase + K-feldspar are normalized to 100%. Estimated 2σ error for all points is less than $\pm 4\%$ (Van der Plas and Tobi, 1965). See Table 1 for data. Classification according to Streckeisen (1973) is as follows: G = granite; Gd = granodiorite; T = tonalite; QS = quartz syenite; QM = quartz monzonite; QMD = quartz monzodiorite; QD = quartz diorite; S = syenite; M = monzonite; Md = monzodiorite; D = diorite.

Fig. 4.

Sample #	Type	Plag.	K-spar	Quartz	Total Light	% Plag.	% K-spar	% Quartz	Other	% Other
19	Md	318	90	94	502	63	18	19	350	41
73	Md	297	81	63	441	67	18	14	220	33
93	Md	287	164	92	543	53	30	17	189	26
132A	Md	247	85	57	389	63	22	15	211	35
132B	Md	342	151	121	614	56	25	20	218	26
133	Md	464	102	116	682	68	15	17	208	23
696	Md	324	145	142	611	53	24	23	264	30
1060	Md	200	---	2	202	99	--	1	99	33
1064	Md	327	224	45	596	55	38	8	403	40
1069	Md	325	97	122	544	60	18	22	337	38
1075	Md	310	160	130	600	52	27	22	231	28
1076	Md	265	160	104	529	50	30	20	278	34
1078	Md	212	188	205	605	35	31	34	37	6
1078A	Md	209	236	160	605	35	39	26	56	8
1080C	Md	277	180	135	592	47	30	23	146	20
464	Mg_1	144	283	196	623	23	45	31	8	1
476	Mg_1	106	276	221	603	18	46	37	3	--
288	Mg_2	160	286	156	602	27	48	26	14	2
474	Mg_2	124	289	189	602	21	48	31	18	3
540	Mg_2	118	295	197	610	19	48	32	5	1
600	Mg_2	121	265	217	603	20	44	36	5	1
1271	Mg_2	87	253	261	601	14	42	43	9	1
1537	Mg_2	118	266	218	602	20	44	36	10	2
1540	Mg_2	127	301	172	600	21	50	29	13	2
1541	Mg_2	127	323	151	601	21	54	25	6	1
1542	Mg_2	122	267	226	615	20	43	37	25	4
961	Mmc	270	159	58	487	55	33	12	114	19
970	Mmc	328	242	49	619	53	39	8	114	16
101	Mgm	154	114	137	405	38	28	34	36	8
1409	Mgm	327	126	158	611	54	21	26	40	6

Table 1. Modal mineral analyses of plutonic rocks in the Avawatz Mountains. Percent plagioclase, K-spar, quartz, given normalized to P + K + Q = 100%. Percent other = color index.

Avawatz Peak on the west side of the Arrastre Spring fault where it intrudes Mesozoic metavolcanic rocks. Like the older granite, it crops out only at the highest structural levels along the crest of the range where it forms roof pendants in the Avawatz quartz monzodiorite. Total area of exposure is about 12 square kilometers.

The younger granite is remarkably similar to the older granite in appearance. Both are leucocratic, medium-grained, and slightly porphyritic. Modal mineral analysis of stained slabs indicates that the rock types are practically identical with both falling well within the granite field (Fig. 4; Table 1). Like the older granite, the younger granite is composed of anhedral andesine and quartz with small subhedral phenocrysts of microcline and minor biotite and opaques. In the field, the two rock types are distinguishable because the older granite is white, beige-white, or light greenish-white, whereas the younger granite is consistently light to dark pink or reddish-pink.

Age: The younger granite of Avawatz Peak is intruded by quartz monzodiorite yielding a K-Ar age of 176 m.y.. This is a minimum age limit for the granite. The younger granite intrudes a thick sequence of Mesozoic metavolcanic rocks west of the Arrastre Spring fault. These metavolcanic rocks are probably not older than metavolcanic rocks in the Soda Mountains that disconformably overlie Lower(?) Triassic metasedimentary rocks. Thus the younger granite is probably younger than Early Triassic.

Avawatz Quartz Monzodiorite complex (Md)

The Avawatz quartz monzodiorite complex is a petrographically diverse body of dark, equigranular plutonic rock extending from one end of the range to the other and cropping out on both sides of the Mormon Spring and Leach Lake faults (Fig. 3). It forms much of the pre-Cenozoic basement and is aerially the most extensive rock type in the range. The quartz monzodiorite complex is clearly intrusive into the older granite of Avawatz Peak and into pre-Mesozoic metasedimentary rocks.

Modal mineral analysis of 15 rock slabs stained for plagioclase and K-feldspar from widely spaced locations within the range indicate that the average composition of the rock is quartz monzodiorite (Fig. 4;

Table 1) (Streckeisen, 1973). Designation of this rock as diorite by past workers (Troxel and Butler, 1979) should be abandoned. Individual samples are variable in composition, ranging from diorite to quartz monzodiorite, granodiorite, granite and quartz monzonite. Most samples form a cluster centered within the quartz monzodiorite field (Fig. 4). Two samples that fall well within the granite field (1078, 1078A) are from irregular, relatively leucocratic dikes which occur locally within the quartz monzodiorite east of Mormon Spring fault and are interpreted as late stage magmatic differentiates. Modal analyses also indicate that the color index (% of dark minerals) is about 25 to 40%, which is unusually dark for calc-alkaline plutonic rocks.

Field examination of the quartz monzodiorite complex indicates that it is not a single homogenous plutonic body. At the highest structural levels along the crest and southwest side of the range, it is generally more fine-grained and biotite-rich, whereas at structually deeper levels in canyon bottoms along the northeast flank of the range, hornblende is more abundant. The petrographic diversity of the quartz monzodiorite complex within these deeply incised canyons is striking. Here, the rock is composed of an amalgamated mass of irregular dikes and bodies typically 5 to 50 meters across. Individual intrusive units vary from biotite-rich (locally up to 80%?!) with hornblende moderate to absent, to hornblende-rich with or without biotite. Hornblende crystals vary from equideminsional to acicular and from fine- to coarse-grained. Color index is also quite variable, and pegmatitic and fine-grained mafic dikes are locally abundant. The abundance of K-spar is variable, and one stained slab contained no K-spar.

In thin section, the rock is generally composed of medium- to fine-grained, non-porphyritic, anhedral biotite, hornblende, oligoclase $(An_{42}-An_{46})$, orthoclase, and quartz. Accessory minerals include opaques (magnetite?), zircon, apatite, sphene(?), and secondary epidote. Epidote is moderately abundant in some samples and may be clearly visible in hand specimen. Hornblende, biotite, and epidote are locally subhedral. Alteration of plagioclase to sericite(?) and associated development of cryptocrystalline mineraloids is highly variable within the

quartz monzodiorite. Completely unaltered rock is rare.

Rb-Sr analysis: Six whole-rock samples (3-5kg each) collected from a
0.5 mile transect along lower Sheep Creek were analyzed for rubidium
and strontium content and strontium isotopes. It was hoped that
strontium isotope analysis would allow accurate determination of the
crystallization age of the rock. Sheep Creek was selected for sample
collection because of the petrographic diversity of the quartz monzo-
diorite here, and because fresh, relatively unaltered rock crops out
at the base of the canyon walls. A petrographically diverse site was
selected because a range of rubidium-strontium ratios is needed for an
accurate isochron.

Four of the six analyzed samples define an isochron of 195 $\pm$ 35
m.y. with an ^{87}Sr/^{86}Sr initial ratio of .70834 $\pm$ .00026 (Fig. 5; Table
2). Although this age does not accurately constrain the age of the quartz
monzonite, it indicates that at least some, and perhaps most, of the
rocks in lower Sheep Creek were homogenized with respect to strontium
isotopes at the time of crystallization. Two samples, however, have
substantially different strontium isotope composition. Model isochrons
of 195 m.y. through points representing these samples yield ^{87}Sr/^{86}Sr
initial ratios of .7062 and .7127 (Fig. 5). The sample with the low
initial ratio (Av-697) is the most altered of all samples collected
(they all contained small amounts of epidote and cryptocrystalline
mineraloids(?)), and the sample with the highest initial ratio (AV-703)
is finer-grained than the others. In general, however, these differences
are minor.

It is remarkable that these samples all look relatively similar in
the field and in thin section, yet have such different strontium isotope
compositions. Alteration probably did not cause such large variations,
and the different strontium isotope compositions are interpreted as
representing original differences at the time of crystallization.

Two processes, not mutually exclusive, may have produced the
isotopic variations in the Avawatz quartz monzodiorite. Variable
contamination of the protolith magma by assimilation of Precambrian
crystalline rocks with high ^{87}Sr/^{86}Sr ratios may have increased the
^{87}Sr/^{86}Sr ratios of all samples except possibly Av-697. It is not

Fig. 5. Isochron diagram for six whole rock samples of Avawatz

quartz monzodiorite. Analyses of four samples define an isochron

of 195 $\pm$ 35 m. y. (1σ error calculated from York (1969)). Model

isochrons of same age through other two points yield $^{87}Sr/^{86}Sr$

initial ratios of .7127 and .7062. $\lambda^{87}Rb = 1.42 \times 10^{-11}yrs$. See

Table 2 for data.

Fig. 5.

	Sample	Rb (ppm)	Sr (ppm)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr^* \pm 2\sigma$
Md	AV 697	29.144	468.70	.18007	.706605 ± .000035
	AV 698	103.900	684.05	.43987	.709433 ± .000050
	AV 700	93.241	568.65	.47485	.709929 ± .000050
	AV 701	113.035	552.77	.59220	.710024 ± .000043
	AV 702	105.423	564.23	.54110	.709642 ± .000035
	AV 703	72.458	496.21	.42288	.713870 ± .000036
Mgm	AV 1406	91.918	683.12	.38967	.710770 ± .000043
	AV 1407	97.570	592.74	.47671	.710832 ± .000043
	Av 1408	82.632	642.31	.37256	.710949 ± .000043

* Sr isotopic composition normalized to $^{86}Sr/^{88}Sr = 0.1194$, and Eimer and Amend $SrCO_3 = .70800$

Table 2. Rubidium and strontium isotope analyses of whole rock samples of Avawatz quartz monzodiorite (Md) and granodiorite of Mormon Canyon (Mgm). Exact location of samples is shown by numbered dots on Plate 1A.

possible to demonstrate a contamination or magma mixing model, however, because there is no apparent relationship between strontium content and isotope composition. Alternatively, strontium isotope variation may reflect the isotopic variability of the source area, and the absence of isotope homogenization during magma transport and crystallization. Magmas derived from an isotopically heterogeneous source may have had similar major element chemistry and undergone similar pathways of crustal fractionation and compositional evolution, perhaps in the same magma conduit, yielding an intrusive body with similar major element composition and petrology yet diverse strontium isotope character. Because of the lithologic and isotopic diversity and complexity of the quartz monzodiorite, it is termed the Avawatz quartz monzodiorite complex.

Age and significance: The oldest K-Ar date obtained from mineral separates of the Avawatz quartz monzodiorite was 176 m.y. on hornblende. Considering the discordant 139 m.y. K-Ar age of biotite from the same rock and the widespread resetting of K-Ar ages in the quartz monzodiorite, this should be considered only a minimum age (Fig. 6; Table 3).

Stratigraphic data from the Soda Mountains located less than 15km south of the Avawatz Mountains constrains the maximum age limit of the quartz monzodiorite. Volcanic flows and breccias in the Soda Mountains disconformably overly limestone and shale containing gastropods of probable Early Triassic age (Grose, 1959). It is doubtful that the limestone and shale were deposited during or after emplacement of the Avawatz quartz monzodiorite because the large size of this intrusive body probably indicates that it was overlain by a substantial volcanic edifice. The absence of significant volcanic detritus in Early(?) Triassic rocks of the Soda Mountains probably indicates that major magmatic activity did not begin in this area until Middle(?) Triassic time or later.

Based on this evidence, the Avawatz quartz monzodiorite is considered to be 200 ± 20 m.y. old. This range falls entirely within the 1 sigma range of the rubidium-strontium isochron.

Other dark, "dioritic" rocks crop out in the Halloran Hills east of the Avawatz Mountains (DeWitt, 1980) and in the Black Mountains of southern Death Valley (Otton, 1976). Quartz diorite in the Halloran

K-Ar ages

Fig. 6. Histogram of K-Ar ages from intrusive rocks of the Avawatz Mountains. Horizontal lines and x's represent dis-cordant hornblende and biotite ages from the same rock. Vertical axis has no significance.

Sample #	Mineral	Formation	Long. (116°)	Lat. (35°)	K_2O (%)	$^{40}Ar_{rad}$ (x 10^{-10} moles/gr)	% $^{40}Ar_{rad}$	Age (m.y.)*
132	B	Md	21' 26"	34' 01"	8.785	15.6211	87.6517	119.5 ±.7
132	B	Md	21' 26"	34' 01"	8.785	15.6384	88.2803	119.6 ±.7
132	H	Md	21' 26"	34' 01"	1.464	2.41351	88.1228	111.0 ±.7
132	H	Md	21' 26"	34' 01"	1.464	2.48675	83.5511	114.3 ±.7
133	B	Md	26' 34"	35' 54"	9.12	11.6063	91.3331	86.3 ±.5
133	H	Md	26' 34"	35' 54"	1.4575	1.96038	53.8217	91.1 ±.6
134B	B	Gneiss clast, Tav₂(?)	24' 12"	36' 12"	9.145	15.3288	91.6181	116.4 ±.7
145	M	Md-pegmatite	30' 53"	36' 10"	10.76	9.83358	91.5298	62.4 ±.4
167	B	p€gn	13' 48"	25' 49"	9.455	13.3903	87.4089	95.8 ±.6
172	B	Md	30' 44"	36' 15"	9.24	6.13402	76.6948	45.5 ±.3
172	B	Md	30' 44"	36' 15"	9.24	6.17652	80.8356	45.8 ±.3
172	H	Md	30' 44"	36' 15"	1.5725	1.77696	80.2345	76.8 ±.5
338	B	Md	16' 16"	29' 06"	8.71	10.5821	88.1774	82.5 ±.5
338	H	Md	16' 16"	29' 06"	1.201	2.33190	90.2969	129.0 ±.8
338	H	Md	16' 16"	29' 06"	1.201	2.32931	90.2279	128.9 ±.8
538	H	Md	19' 13"	30' 05"	.802	1.70125	81.7593	141.6 ±.9
738	H	Hornblendite, Tav₂	18' 24"	27' 51"	.697	2.86098	89.4535	264.7 ±1.6
902	M	Breccia, Tav₄	18' 20"	23' 48"	9.095	20.0217	82.7093	146.8 ±.9
1070	H	Md	29' 22"	14' 41"	.728	1.15175	81.534	106.7 ±.7
1269	H	Md	18' 14"	27' 58"	.712	1.90184	82.2603	175.6 ±1.1
961	B	Mmc	26' 7"	35' 28"	8.995	18.1695	91.9429	135.1 ±.8
1070	H	Md	29' 22"	14' 41"	.728	1.15175	81.534	106.7 ±.7
1070	B	Md	29' 22"	14' 41"	9.36	9.14203	77.0318	66.6 ±.4
1269	H	Md	18' 14"	27' 58"	.712	1.90184	82.2603	175.6 ±1.1
1269	B	Md	18' 14"	27' 58"	8.635	17.9391	91.9542	138.8 ±.8

* Age calculated using new decay and abundance constants:

$^{40}K/K = 1.167 \times 10^{-4}$ mol/mol

$\lambda_\beta = 4.962 \times 10^{-10}$ yr^{-1}

$\lambda_\epsilon + \lambda'_\epsilon = 0.581 \times 10^{-10}$ yr^{-1}

Table 3. Potassium-argon analyses of mineral separates from rocks from the Avawatz Mountains. H = hornblende; B = biotite; M = muscovite. Samples 134B and 738 are from clasts within the Tertiary Avawatz Formation. All others are from the pre-Cenozoic basement.

Hills intrudes rocks that were metamorphosed prior to 140 m.y., indi-
cating a post 140 m.y. age for emplacement (DeWitt, 1980). This suggests
that the Avawatz quartz monzodiorite was emplaced several tens of millions
of years before the quartz diorite in the Halloran Hills, and that they
are not correlative.

Granodiorite of Mormon Canyon (Mgm)

A small stock of white granodiorite intrudes the Avawatz quartz
monzodiorite in Mormon Canyon. Several smaller bodies are exposed in
the canyon north of Mormon Canyon, but have not been completely mapped.
The total exposed area of this rock is about three square kilometers.
Intrusive contacts with the quartz monzodiorite are highly irregular and
appear to dip outward. Dikes eminate outward from the stock into the
quartz monzodiorite host, indicating that the granodiorite is clearly
the younger intrusive rock.

Modal mineral analysis of two stained slabs indicate that rock type
is variable within the stock. Sample 101 (Fig. 4; Table 1) is from the
margin of the stock and is of granitic composition, whereas sample
1409 is from the interior of the stock and is granodiorite in composi-
tion. Although not demonstrated by the two analyzed samples, field
observations indicate that marginal rocks contain less dark minerals
(mainly biotite) than interior rocks. The granodioritic composition
of the interior rocks is probably the dominant composition.

In thin section, the rock is composed of anhedral K-feldspar,
andesine (An_{25-30}), quartz, and subhedral biotite and accessory opaques
(magnetite?). Plagioclase is slightly altered to cryptocrystalline
mineraloids. The rock is medium-grained equigranular to slightly
porphyritic.

Three samples were analyzed for rubidium and strontium content and
isotopes. Model isochrons representing the approximate possible range
of ages of this rock indicate that the $^{87}Sr/^{86}Sr$ initial ratio is .7111
$\pm$ .0002 (Fig. 7; Table 2).

Quartz monzonite of Cave Spring Wash (Mmc)

This rock type crops out along the eastern end of the Leach Lake
fault in the northwestern Avawatz Mountains (Fig. 3). It is exposed over

Fig. 7. Isochron diagram of three samples of granodiorite of Mormon

Canyon. Model isochrons of 80 and 200 m. y. define $^{87}Sr/^{86}Sr$ ratio

of .7111 ± .0002. Additional data was not produced due to lack of

time and disappointing $^{87}Rb/^{86}Sr$ range.

Fig. 7.

an area of about six to eight square kilometers and occurs on both sides
of the Leach Lake fault. Its contact with other Mesozoic plutonic rocks
is faulted or not presently mapped. It clearly intrudes the Pennsylvanian
to Permian Bird Spring Formation, and has yielded a K-Ar biotite age of
135 m.y..

The rock is composed of oligoclase $(An_{42}-An_{48})$, orthoclase, quartz,
biotite, and hornblende with accessory opaques. Mineral grains are sub-
hedral to anhedral, and the texture is medium-grained and equigranular.
Oligoclase exhibits slight normal zoning. Modal mineral analysis of
two stained slabs indicates that the rock is a quartz monzonite (Fig. 4;
Table 1).

The quartz monzonite of Cave Springs wash contains less than 10%
quartz and when plotted on a KQP ternary diagram falls well below the
typical range of "Sierran type" calc-alkaline plutonic rocks (Fig. 8).
It does, however, fall within the field of Mesozoic alkalic plutonic
rocks occurring sparsely in southern and eastern California (Miller,
1978; Dunne and others, 1978). The chemical and mineralogical composi-
tion of these alkalic plutons indicates that these granitic rocks were
derived from an alkalic mantle source distinctly different from the source
of the volumetrically more abundant calc-alkaline plutonic rocks (Miller,
1978).

Based on gross petrologic similarity to alkalic-plutonic rocks in
southern and eastern California which are consistently of early Jurassic
or Triassic age, the quartz monzonite of Cave Springs wash is tentatively
assigned an early Jurassic to Triassic age. A 136 m.y. K-Ar age on
biotite is an absolute minimum age for this rock (Table 3).

Metavolcanic rocks (Mv)

Mesozoic metavolcanic rocks crop out west of the northern Arrastre
Spring fault and were mapped only in reconnaissance. These rocks are
generally green to brown with sparse to abundant feldspar microlites
and phenocrysts. Mafic phenocrysts are altered beyond recognition. Fine-
grained chlorite and epidote are visible in some hand specimens. Flows,
breccias, and volcaniclastic rocks are all present. Metasedimentary rocks
are locally interbedded with the volcanic rocks. These are composed of
maroon, fissile shale, light pale green to dark gray laminated and indurated

Fig. 8. Ternary diagram of essential modal mineralogy of plu-
tonic rocks in southern California. Solid line = field of alkalic
intrusives; dashed line = Sierran type calc-alkaline intrusives;
dotted line = calc-alkaline intrusives of eastern Sierra Nevada
and southwestern Great Basin. Data from Dunne and others (1978)
and references therein. Circles = granite of Avawatz Peak, undif-
ferentiated; x's = Avawatz quartz monzodiorite; solid squares =
quartz monzonite of Cave Spring wash.

argillite, and one 30cm thick bed of gray limestone containing shale
chips.

Intrusive rocks, undifferentiated (Mi)

Two small bodies of plutonic rock intrude late Paleozoic carbonate
rocks at the extreme northwestern end of the map area covered by Plate
1A. These are composed of medium-grained, slightly porphyritic, biotite
bearing granitic rock. The rock looks similar to the Avawatz Peak granites
except that it is grayer in color.

MESOZOIC DEFORMATION

Evidence of Mesozoic deformation occurs in three areas in the
Avawatz Mountains: 1) Pre-Mesozoic metasedimentary rocks, especially
in the south and central pendants, record deformation before and/or during
emplacement of the granites of Avawatz Peak and the Avawatz quartz
monzodiorite. Emplacement of these plutonic rocks left the metasedi-
mentary rocks completely recrystallized, indurated, and surrounded by a
rigid plutonic envelope, rendering them highly resistant to later defor-
mation. 2) East of the central pendant, Paleozoic and Precambrian meta-
sedimentary rocks that were originally separated by thousands of meters
of sedimentary rocks are now separated by only a few hundred meters of
Mesozoic plutonic rock. Early Mesozoic thrust faulting may have juxta-
posed these rocks prior to emplacement of the quartz monzodiorite. 3) In
one area along the Mule Spring fault, the quartz monzodiorite is penetra-
tively deformed into a schistose gneissic rock.

Deformed pre-Mesozoic metasedimentary rocks

Paleozoic and upper Precambrian metasedimentary rocks are invariably
deformed by at least one and probably several events. None of these
deformations was studied in detail or in a statistically rigorous manner
because of small outcrop areas, brecciation, and metamorphism. A num-
ber of different structural styles are apparent. Interbedded quartzite
and silty quartzite in the lower Cambrian Wood Canyon Formation have been
boudinaged. Argillite in the lower and upper members of the Cambrian
Carrara Formation show extreme lateral thickness variations and are highly

attenuated in the southern pendant. Two large, overturned, west-vergent folds are exposed within rocks at the north end of the south pendant and in the upper Precambrian Stirling Quartzite in the central pendant (Plate 1A). At least a half dozen north vergent overturned folds occur in the south pendant. These are typically 5 to 50 meters in amplitude and are tight to isoclinal. The largest of these folds contains the Carrara-Bonanza King contact in the northern part of the southern pendant (Plate 1A). Well-layered and laminated calcite and dolomitic calcite marble of the Papoose Lake Member of the Bonanza King Formation is locally foliated and in places bedding is completely transposed. Brecciation of carbonate rocks, especially dolomite, is moderate to extreme in the Banded Mountain Member and overlying car-bonate rocks. Precambrian and Cambrian quartzites are also variably brecciated. Some of this brecciation may be the result of Tertiary fault-ing. Carbonate rock in numerous small pendants in the Avawatz Mountains are invariably folded and brecciated, but no consistent fold pattern or structural style is apparent. Disharmonic folds are often associated with brecciation.

Styles of deformation can be divided into several categories, each possibly representing a discrete episode of deformation: 1) penetrative deformation with the development of transposition foliation, 2) north-vergent folding, 3) west-vergent folding, 4) boudinage and attenuation, and 5) brecciation and disharmonic folding. The sequence of these events is not discernable on the basis of structural overprinting, and can only be inferred in some cases on the basis of regional correlation or structural style. The transposition foliation in the marbles of the Papoose Lake Member is similar in structural style to penetratively de-formed Pennsylvanian and Permian carbonate rocks in the Bird Spring For-mation in the nearby Soda Mountains. This deformation pre-dates deposi-tion of probable Lower Triassic calc-silicate rocks and conglomerates in the Soda Mountains, and thus is Late Permian or Early Triassic in age. Penetrative deformation in the Avawatz Mountains could be correlative with this deformation in the Soda Mountains, or it could be younger and unrelated. North- and west-vergent folding and boudinage and attenuation clearly predate final emplacement of the Avawatz quartz monzodiorite, and probably predate emplacement of the granites of Avawatz Peak. Disharmonic

folding could be related to any of the other deformational events, or it could post-date them. Brecciation probably occurred during the later stages of metamorphism and deformation and reflects the increasingly brittle character of the rock following recrystallization due to earlier metamorphic and plutonic events.

Mesozoic thrust or high-angle faulting

The distribution of several small pendants located east of the central pendant and below a Tertiary low-angle fault has significant structural implications. Three irregular bodies of brecciated and folded, gray and white calcite marble (PCu) form small pendants in the quartz monzodiorite, and are located east of the southern end of the central pendant (Plate 1A). Single elongate, white calcite crystals, up to 1cm diameter, occur sparsely in the marble, and are probably crinoid columnals. These are common in the Pennsylvanian to Permian Bird Spring Formation, and the marble pendants are correlated with this formation on the basis of fossil and lithologic evidence. Precambrian quartzite (p€q) forms a pendant in the quartz monzodiorite approximately one kilometer north of the marble pendants, and a pendant of Precambrian metasediments (p€u) occurs several hundred meters to the east. When these rocks were in their original stratigraphic sequence, they were separated by 5 to 10 thousand meters of sedimentary rock. The present small distance between these rock types can be interpreted two ways: 1) the rocks were juxtaposed during intrusion of the quartz monzodiorite by primarily vertical movements, or 2) these rocks were juxtaposed during early Mesozoic thrust faulting prior to quartz monzodiorite intrusion. In the second interpretation, early Mesozoic thrusting placed Precambrian metasedimentary rocks over late Paleozoic marble, and the thrust contact was subsequently obliterated by intrusion of the quartz monzodiorite.

Foliated Avawatz quartz monzodiorite

Penetratively deformed Avawatz quartz monzodiorite is well-exposed in the large fault block between the Leach Lake and Mule Spring faults. Three canyons are deeply incised into this fault block (Plate 2) and the quartz monzodiorite is well-exposed in the canyon walls. Foliation is faint in the eastern canyon, faint to moderately developed in the

middle canyon, and very well-developed in the eastern canyon. Field examination of areas between canyons confirms the gradual westward increase in foliation intensity from east to west.

The foliation is primarily defined by biotite orientation. In thin section, parallel subhedral biotite grains are surrounded by anhedral quartz, plagioclase, orthoclase, epidote, hornblende and opaques with no preferred orientation. Foliation is so well-developed in western exposures that the rock cleaves readily on foliation planes and has a schistose character. Numerous aplitic dikes are intruded parallel or subparallel to the foliation plane. Locally, these aplite dikes, typically 1 to 20cm thick, contain biotite or garnet. Rare biotite grains within aplite dikes are oriented parallel to the foliation plane, indicating that their emplacement predated or was synchronous with foliation development. At least one 10 to 20cm thick aplite dike is folded isoclinally with the foliation plane parallel to the fold axial plane. Thin, 1cm thick dikes intruded at high angles to the foliation plane may or may not be pervasively folded. A few aplite dikes cut at high angles across the foliation plane and clearly post-date foliation development. Local pegmatite dikes up to two meters thick also cut across the plane of foliation and clearly post-date it.

Over a hundred foliation attitudes were measured in the field on the western part of the fault block (Figs. 9, 10). When plotted on a stereonet, these attitudes form a consistent pattern. The highest density of points define a foliation plane striking N88W and dipping 38° to the south. North dipping foliation is also locally present and defines a separate locus of points on the stereonet. These north dipping foliations occur on the north dipping limbs of folds in the foliated quartz monzodiorite. As defined by the stereonet, and consistent with field observation, these folds have horizontal, east-west trending axes.

The timing of deformation is poorly constrained. The quartz monzodiorite is 200 ± 20 million years old, placing an approximate older age limit for the deformation. A K-Ar date of 62.4 m.y. was obtained on a muscovite separate from a pegmatite dike that cross-cuts the foliation at a high-angle and is not itself foliated (Table 1). This indicates that the deformation is Mesozoic and not Cenozoic in age.

Fig. 9. Simplified geologic map of northwestern Avawatz Mountains.
Md = foliated Avawatz quartz monzodiorite. Attitude of foliation
shown by line with solid triangle. Tav_2 = Avawatz Formation(?),
second member.

Fig. 9.

Fig. 10. Lower hemisphere projection of normal to foliation planes. Contours show number of points per 1% area. One hundred and twenty points counted.

POTASSIUM-ARGON GEOCHRONOLOGY

Sixteen mineral separates from nine locations within the Mesozoic plutonic complex in the Avawatz Mountains were dated by the potassium-argon method (Table 1). Twelve mineral separates represent six biotite-hornblende pairs from six widely spaced locations within the Avawatz quartz monzodiorite (Fig. 3, samples 132, 133, 172, 338, 1070, 1269). One hornblende separate is from another location within the Avawatz quartz monzodiorite (sample 538), one biotite separate is from the Precambrian gneiss (sample 167), one muscovite separate is from a pegmatite that intrudes foliated Avawatz quartz monzodiorite (sample 145), and one biotite is from the quartz monzonite of Cave Springs wash (sample 961). These ages are plotted on a histogram in figure 6.

Potassium-argon dating was undertaken in order to determine the cooling history of the crystalline rocks, to constrain the timing of pluton emplacement, and to constrain the timing of deformation. The oldest age obtained, 176 m.y., is considered a minimum age for the Avawatz quartz monzodiorite, and the 62 m.y. age obtained from muscovite from a post-tectonic pegmatite is considered a minimum age for the deformation of the Avawatz quartz monzodiorite complex.

The cooling history of the crystalline basement has been very complex. Biotite and hornblende mineral pairs from single samples yielded ages that are discordant (different from each other) by as much as 47 m.y. (Fig. 6, sample 338), and one pair is reversely discordant (sample 132). Two patterns are apparent from the distribution of ages. The two oldest ages (samples 538, 1269) are from the structurally highest part of the range where the quartz monzodiorite is in intrusive contact with older roof pendants of granite and pre-Mesozoic metasedimentary rocks. The youngest ages are from the western end of a fault block of foliated quartz monzodiorite (samples 145, 172), and from east of the Mormon Spring fault (sample 1070). The other ten ages fall in a 58 m.y. time span extending from 140 to 82 m.y.. These are almost all from samples collected at lower elevations on the east and north flanks of the range, and have undergone significant vertical uplift and un-roofing in late Cenozoic time.

Clearly all the biotite and hornblende K-Ar ages, except possibly

one, are younger than the age of crystallization. This may be the result of two end member processes: 1) K-Ar ages represent the time that each mineral passed through its blocking temperature for argon during a long period of gradual cooling and uplift following crystallization, or 2) all or most minerals passed through their K-Ar blocking temperatures shortly after crystallization, but were later reheated and underwent partial or complete outgassing as a result of renewed thermal activity.

The Avawatz Mountains were certainly in a tectonically active area during Mesozoic time. The west dipping Keystone and Winters Pass thrust faults are exposed 60km east of the Avawatz Mountains in the Clark Mountains (Fig. 1). The Keystone thrust cuts granitic rocks yielding K-Ar biotite ages of 136 and 138 m.y., and appears to predate plutonic rocks dated at 92 and 94 m.y. (K-Ar, hornblende) (Sutter, 1968; Burchfiel and Davis, 1971). Movement on the Winters Pass thrust fault is poorly dated, although it probably ended before 92 m.y. (Burchfiel and Davis, 1971). In the southwestern Silurian Hills located 25km east of the Avawatz Mountains, the south dipping Riggs thrust fault is intruded by syn-tectonic (Abbott, 1971) or post-tectonic (DeWitt, 1980) granitic rocks yielding K-Ar ages of 86 and 88 m.y. (Sutter, 1968). The Avawatz Mountains are probably in the upper plate of all of these thrusts, and may have undergone significant vertical uplift and cooling as a result of movement on them.

Potassium-argon ages younger than 80 m.y. are obtained from rocks located at the west end of a large fault block along the Garlock fault zone. A biotite-hornblende pair from this locality yields discordant ages of 45 and 77 m.y. (sample 172, Table 3). Tectonic uplift of fault blocks along the Garlock fault may have begun in early Tertiary time (Spencer, this report), although the most severe tectonic activity probably did not begin until Miocene time. The discordance of K-Ar ages from sample 172 indicate that the rocks were not simply uplifted over a short period of time through the argon blocking temperatures for biotite and hornblende. A two-stage uplift history, with the rocks first passing through the hornblende blocking temperature, followed by a second, distinct episode of uplift during which the rocks passed through the biotite blocking temperature, is not likely because there

is no evidence for the older tectonic event, and the younger event post-dates inferred tectonism by 5 to 15 m.y.. More likely, these young K-Ar ages represent partial resetting as a result of late Tertiary magmatic activity. Several Tertiary dikes are present in the foliated quartz monzodiorite, and Tertiary volcanic rocks occur nearby. A sample of one of these dikes yields K-Ar whole rock ages of 11.4 and 12.4 m.y..

Thermal activity and reheating of the Avawatz quartz monzodiorite may have led to partial or complete argon loss in some or all of the dated minerals. However, over the entire exposed area of quartz monzodiorite, younger plutonic rocks are exposed in only one area. The granodiorite of Mormon Canyon intrudes the quartz monzodiorite in the central Avawatz Mountains, but its age is unknown. Potassium-argon dates from nearby areas do not show evidence of Tertiary thermal activity, indicating that the granodiorite of Mormon Canyon is probably Mesozoic in age. Late Mesozoic plutonic rocks may, however, occur at greater depths but are not presently exposed.

Except for sample 172 at the far northwest corner of the study area and sample 1070 east of the Mule Spring fault, all biotite and hornblende ages are older than 80 m.y.. Except for two hornblende ages from samples collected along the crest of the range, all dated minerals yield ages younger than 140 m.y.. A plausible model for the origin of these dates is that a heating event occurred at about 90 m.y. ago. At this time, most hornblende ages were partially reset, while one sample (133) was completely reset. Biotite ages were partially or completely reset, and those that were completely reset did not cool through the argon blocking temperature until perhaps 5 to 10 m.y. after the heating event (samples 133, 338). This heating event could be the result of emplacement of the granodiorite of Mormon Canyon, or of magma emplacement at deeper levels not yet exposed. If emplacement of the granodiorite of Mormon Canyon is the cause of resetting, this intrusive body must be far more extensive at depth than is indicated by surface exposures. Hornblende from sample 133 yields an age of 91 m.y. even though it is located at least 10km from exposures of the granodiorite.

Evidence for magmatic or thermal activity at about 90 m.y. is also

present in the Halloran Hills, Silurian Hills, and Clark Mountains. Hornblende and biotite from two plutonic bodies in the Halloran Hills, located about 25 to 30km southeast of the Avawatz Mountains, yield flat $^{40}Ar/^{39}Ar$ release spectra indicating rapid cooling at 90 to 94 m.y. ago (DeWitt, 1980). Biotite from plutonic rocks in the Silurian Hills yield K-Ar ages of 86 and 88 m.y., and hornblende from the Teutonia quartz monzonite located about 50 to 60km SSE of the Avawatz Mountains yield K-Ar ages of 86 and 88 m.y. (Sutter, 1968). These ages could be significantly younger than the crystallization age of the dated minerals and host rock, or they could approximately date crystallization. The flat $^{40}Ar/^{39}Ar$ release spectra of the samples from the Halloran Hills indicate rapid cooling, suggesting that these minerals date a crystallization event rather than a tectonic uplift event.

In conclusion, it is not possible to eliminate one of two possible causes for resetting of K-Ar ages in the Avawatz Mountains. Late Tertiary magamtic activity probably caused resetting in sample 172 and possible muscovite sample 145 and biotite sample 1070, but probably can not be invoked to account for resetting of the other samples. The wide range of late Mesozoic K-Ar ages and the highly variable nature of age discordance indicates that this area did not undergo simple broad uplift, erosion, and cooling. More likely, age resetting resulted from heating events related to magmatism about 90 m.y. ago and possibly older heating events as well. These heating events may have produced complex, irregular geothermal gradients which changed rapidly in both space and time. Thermal changes resulting from tectonic activity may have been superimposed on geothermal gradients resulting from magmatic activity, leading to an even more complex thermal history.

CONCLUSION

Major Mesozoic deformational, magmatic, and thermal events in the Avawatz are: 1) foliation development in some Paleozoic carbonate rocks prior to emplacement of the Avawatz quartz monzodiorite, 2) folding and attenuation of pre-Mesozoic metasediments prior to emplacement of the Avawatz quartz monzodiorite, 3) emplacement of the Avawatz quartz monzo-

diorite at about 200 ± 20 m.y. ago, 4) deformation of the Avawatz
quartz monzodiorite after 200 ± 20 m.y. and before at least 62 m.y.,
and 5) final cooling of most of the Avawatz quartz monzodiorite between
140 and 80 m.y. ago.

The transposition foliation in the marbles of the Papoose Lake
Member of the Cambrian Bonanza King Formation may represent the oldest
deformation. The similarity in structural style of these rocks to
penetratively deformed carbonate rocks in the Soda Mountains suggests that
they are correlative. Deformation in the Soda Mountains occurred during
late Permian to Early Triassic time (Burchfiel and others, 1980), and sig-
nificantly pre-dates regional Middle Triassic to Early Jurassic deformation.

Folding of pre-Mesozoic rocks is probably younger and related
regionally to a belt of Middle Triassic to Early Jurassic deformation
extending from the northern Panamint and Inyo ranges north of the
Avawatz Mountains to the Cowhole and Clark Mountains to the south and
east. In the northern area, east directed thrusting in the northern
Panamint and Inyo ranges occurred prior to emplacement of the 183 m.y.
old Hunter Mountain pluton (Burchfiel and others, 1970; Dunne and others,
1978). The largest of these thrusts, the Last Chance thrust, probably
extends northward into south-central Nevada (Burchfiel and others, 1970).
South of the Avawatz Mountains, a thrust fault involving Paleozoic rocks
is overlain by the upper Triassic(?)-lower Jurassic Aztec sandstone
(Novitsky and Burchfiel, 1973). East of the Avawatz Mountains, plutons
dated at 190 and 200 m.y. cut east vergent structures in the Clark Moun-
tains (Burchfiel and Davis, 1971). Regional stratigraphic and struc-
tural relationships suggest that all of these deformations are post-
early Triassic. If these structures are all remnants of a once continuous
belt of deformation (Burchfiel and Davis, 1981), then pre-Mesozoic rocks
in the Avawatz Mountains were probably involved in this deformation.
Thus, folding of pre-Mesozoic rocks in the Avawatz Mountains is tenta-
tively correlated with middle Triassic to early Jurassic deformation in
a belt extending from the northern Panamint and Inyo Mountains, through
the Avawatz Mountains to the Clark and Cowhole Mountains. One problem with
this correlation is that structures in the Avawatz Mountains are north
or west vergent while the other structures of this belt of deformation

are east vergent. This could be attributed to locally complex patterns of stress and strain within the belt.

Attenuation and boudinage of pre-Mesozoic rocks may have occurred in response to stresses generated by emplacement of either the granites of Avawatz Peak or the Avawatz quartz monzodiorite.

The Avawatz quartz monzodiorite is at least 176 m.y. old (K-Ar, hornblende). Biotite from the same rock yields a K-Ar age of 139 m.y.. The discordance between these two dates indicates that the rock has had a complex thermal history, and the 176 m.y. age should be considered only a minimum age. Stratigraphic relationships in the nearby Soda Mountains indicate that the quartz monzodiorite is not older than about Middle Triassic (about 220 m.y.). The age of the Avawatz quartz monzonite is thus considered to be 200 $\pm$ 20 m.y.. If this plutonic body is greater than about 200 m.y. old, it is possible that it could have been part of the thermal "trigger" for pre-200 m.y. deformation and thrusting in the nearby Clark Mountains.

Penetrative deformation of the Avawatz quartz monzodiorite in the northwestern Avawatz Mountains occurred between 200 $\pm$ 20 m.y. and 67 m.y.. Unfortunately, the timing can not be better constrained. Preliminary paleomagnetic data from late Miocene dikes in the foliated quartz monzodiorite indicate that it has undergone 20° to 40° of counterclockwise rotation, and south-side-down rotation about an east-west trending horizontal axis (Lisa Kanter, written communication, 1981). Restoration of this movement places the foliation in a subhorizontal orientation. The origin of this foliation is not known.

Final cooling of the Avawatz quartz monzodiorite occurred primarily during a 60 million year interval between 80 and 140 million years. The long and complex cooling history of these rocks is attributed to a complex interplay of reheating due to magmatism and cooling due to tectonic uplift. A heating event at about 90 m.y. is suggested on the basis of K-Ar age distributions and regionally extensive magmatism at this time. K-Ar ages younger than 80 m.y. are attributed to late Cenozoic magmatism and heating.

The broad distribution of K-Ar ages clearly illustrates the difficulty of determing crystallization ages of plutonic rocks in batholith terranes. Uranium-lead zircon dating and rubidium-strontium whole

rock dating of dozens or even hundreds of individual plutons in the Mojave region will probably have to be completed before a clear space-time pattern of the locus of magmatism is determined. Further field mapping is also needed to clarify the nature, style, and location of areas of Mesozoic deformation. We have only just begun to understand the nature of the relationship between magmatism and deformation in this important part of the Cordilleran orogen.

BIBLIOGRAPHY

Abbott, E. W., 1971, Structural Geology of the southern Silurian Hills, San Bernardino County, California: Unpublished M. S. Thesis, Rice University, 48 p.

Armstrong, R. L., and Suppe, J., 1973, Potassium argon geochronology of the Mesozoic igneous rocks in Nevada, Utah and southern California: Geological Society of America Bulletin, v. 84, p. 1375-1392.

Burchfiel, B. C., and Davis, G. A., 1971, Clark Mountain thrust complex in the Cordillera of southeastern California: Geologic summary and field trip guide: University of California, Riverside, Campus Museum Contributions, v. 1, p. 1-28.

Burchfiel, B. C., and Davis, G. A., 1972, Structural Framework and Evolution of the Southern Part of the Cordilleran Orogen, Western United States: American Journal of Science, v. 272, p. 97-118.

Burchfiel, B. C., and Davis, G. A., 1975, Nature and Controls of Cordilleran orogenesis, western United States: Extensions of an earlier synthesis: American Journal of Science, v. 275-A, p. 363-396.

Burchfiel, B. C., and Davis, G. A., 1976, Compression and crustal shortening in an Andean-type orogenesis: Nature, v. 260, p. 693-694.

Burchfiel, B. C., and Davis, G. A., 1981, Mojave desert and environs, in Ernst, W. G., editor, The Geotectonic Development of California; Prentice-Hall, New Jersey, p. 217-252.

Burchfiel, B. C., Cameron, C. S., Guth, P. L., Spencer, J. E., Carr, M. D., Miller, E. L., and McCulloh, T. H., 1980, A Triassic overlap assemblage in northern Mojave/Death Valley region: an interpretation: Geological Society of America Abstracts with Programs, v. 12, n. 7, p. 395.

Burchfiel, B. C., Pelton, P. C., and Sutter, J., 1970, An early Mesozoic deformation belt in south-central Nevada-southeastern California: Geological Society of America Bulletin, v. 81, p. 211-216.

DeWitt, E., 1980, Geology and Geochronology of the Halloran Hills, southeastern California, and implications concerning Mesozoic Tectonics of the southwestern Cordillera: Pennsylvania State University, University Park, Pennsylvania, Ph.D. thesis, 267 p.

Dickinson, W. R., 1970, Relations of andesitic volcanic chains, granitic batholith belts, and derivative graywacke-arkose facies to the tectonic framework of arc trench systems: Review of Geophysics and Space Physics, v. 8, p. 813-860.

Dunne, G. C., 1977, Geology and structural evolution of Old Dad Mountains, Mojave Desert, California: Geological Society of America Bulletin, v. 88, p. 737-748.

Dunne, G. C., Gulliver, R. M., and Sylvester, A. G., 1978, Mesozoic evolution of rocks of the White, Inyo, Argus and Slate Ranges, eastern California, in Howell, D. G., and McDougall, K. A., editors, Mesozoic paleogeography of the Western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2, p. 189-207.

Grose, L. T., 1959, Structure and petrology of the northeast part of the Soda Mountains, San Bernardino County, California: Geological Society of America Bulletin, v. 70, p. 1509-1548.

Hamilton, W., 1969, Mesozoic California and the underflow of the Pacific mantle: Geological Society of America Bulletin, v. 80, p. 2409-2430.

Hamilton, W., and Meyers, W. B., 1966, Cenozoic Tectonics of the Western United States: Reviews of Geophysics, v. 4, p. 509-549.

Johnson, B. K., 1957, Geology of a part of the Manly Peak Quadrangle, southern Panamint Range, California: University of California Publications in Geological Sciences, v. 30, p. 353-423.

Kistler, R. W., and Peterman, Z. E., 1973, Variations in Sr, Rb, K, Na, and initial Sr 87/Sr 86 in Mesozoic granitic rocks and intruded wall rocks in central California: Geological Society of America Bulletin, v. 84, p. 3489-3512.

Kistler, R. W., and Peterman, Z. E., 1978, Reconstruction of Crustal Blocks of California on the Basis of Initial Strontium Isotopic Compositions of Mesozoic Granitic Rocks: U. S. Geological Survey Professional Paper 1071, 17 p.

Merriam, C. W., 1963, Geology of the Cerro Gordo Mining District, Inyo County, California: U. S. Geological Survey Professional Paper 408, 83 p.

Miller, C. F., 1978, An early Mesozoic alkalic magmatic belt in western North America, in Howell, D. G., and McDougall, K. A., editors, Mesozoic paleogeography of the Western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2, p. 163-173.

Miller, E. L., and Carr, M. D., 1978, Recognition of possible Aztec-equivalent sandstones and associated Mesozoic metasedimentary deposits within the Mesozoic magmatic arc in the southwestern Mojave Desert, California, in Howell, D. G., and McDougall, K. A., editors, Mesozoic paleogeography of the Western United States: Society of Economic Paleontologists and Mineralogists, Pacific Section, Pacific Coast Paleogeography Symposium 2, p. 283-289.

Novitsky, J. M., and Burchfiel, B. C., 1973, Pre-Aztec (Upper Triassic(?)-Lower Jurassic) thrusting, Cowhole Mountains, southeastern California (abstract): Geological Society of America Abstracts with Programs, v. 5, n. 7, p. 755.

Otton, J. K., 1976, Geologic Features of the Central Black Mountains, Death Valley, California: California Division of Mines and Geology Special Report 106, p. 27-33.

Streckeisen, A. L., 1973, Plutonic rocks-classification and nomenclature recommended by IUGS subcommission on Systematics of Igneous Rocks: Geotimes, v. 18, n. 10, p. 26-30.

Sutter, J. F., 1968, Geochronology of major thrusts, southeastern Great Basin, California: Unpublished M. S. Thesis, Rice University, 32 p.

Troxel, B. W., and Butler, P. R., 1979, Tertiary and Quaternary fault history of the intersection of the Garlock and Death Valley fault zones, southern Death Valley, California: unpublished report submitted to U. S. Geological Survey, Menlo Park, California, 29 p.

Van der Plas, L., and Tobi, A. C., 1965, A chart for judging the reliability of point counting results: American Journal of Science, v. 263, p. 87-90.

York, D. E., 1969, Least squares fitting of a straight line with correlated errors: Earth and Planetary Science Letters, v. 5, p. 320-324.

CENOZOIC STRATIGRAPHY AND STRUCTURE

INTRODUCTION

The left-lateral Garlock fault extends for 260km from the San
Andreas fault to its apparent termination in the Avawatz Mountains
where it meets the right-lateral Death Valley fault zone (Fig. 1).
The Mule Spring fault forms the northern branch of the eastern
Garlock fault. The linear, predominantly strike-slip western seg-
ment of the Mule Spring fault changes to the curving, predominantly
reverse eastern segment at its intersection point with the Death
Valley fault zone (Figs. 2, 3) (Troxel and others, 1979). The
southern and central branches of the Death Valley fault zone are
overridden and truncated by the reverse fault part of the Mule Spring
fault (Troxel and Butler, 1979; Brady and Troxel, 1981). Quaternary
reverse movement on the eastern segment of the Mule Spring fault is
responsible for the present morphology of the Avawatz Mountains, that
is dominated by the steep, arcuate mountain face on the north and
east flanks of the range. A simple way to view the fault movements
is to imagine that the triangular shaped fault block between the Death
Valley fault zone and the western strike-slip segment of the Mule
Spring fault is being pulled northwestward away from the Avawatz
Mountains, while the other two blocks are converging to take its
place, producing the eastern reverse segment of the Mule Spring fault.
These three faults and fault blocks can be adequately represented by
a triple junction diagram normally used for plate tectonic interactions.

The Garlock fault is a major, east-west trending structure separa-
ting the Basin and Range tectonic province, undergoing active east-
west crustal extension, from the relatively topographically and tec-
tonically subdued Mojave block. Correlation of dike swarms on the south
side of the Garlock fault with the southern part of the Independence
dike swarm north of the fault was first proposed by Smith (1962), who
concluded that approximately 64km of offset had occurred since Meso-
zoic emplacement of the dikes (Fig. 4). Smith and Ketner (1970) and
Davis and Burchfiel (1973) identified several other offset features,

Fig. 1. Fault map of Mojave-southwestern Great Basin regions showing location of Avawatz Mountains and major Cenozoic faults. MSF = Mule Spring fault.

Fig. 2. Fault map of Avawatz Mountains. Only major faults are shown. Note that the western segment of the Mule Spring fault is a linear strike-slip fault, whereas the eastern segment is a curving reverse fault. The transition occurs at the intersection point with the southern branch of the Death Valley fault zone.

Fig. 3. Simplified geologic map of the Avawatz Mountains.
Cross-hatch = Arrastre Spring Group (Tas)
Horizontal Lines = first member, Avawatz Formation (Tav_1)
Vertical Lines = second member, Avawatz Formation (Tav_2)
Dots = third member, Avawatz Formation (Tav_3)
Dashed Lines = fourth member, Avawatz Formation (Tav_4)

Fig. 4. Schematic geologic map showing features offset along

Garlock fault (modified from Davis and Burchfiel, 1973).

A-A' = Rand schist; B-B' = Paleozoic metasediments (Garlock

Formation) at Pilot Knob Valley and El Paso Mountains (Smith

and Ketner, 1970); C-C' = Jurassic Independence dike swarm

(Smith, 1962); D-D' = Layton Well thrust and offset equivalent

in Granite Mountains; E-E' = Panamint Valley fault and Arrastre

Spring fault; F-F' = Cambrian metasediments.

Fig. 4.

all indicating major displacement. The Layton Well thrust, located
about 70km west of the Avawatz Mountains, is correlated with a thrust
fault on the south side of the Garlock fault located only 15 to 20km
west of the Avawatz Mountains (Davis and Burchfiel, 1973). If the
Garlock fault does indeed terminate at its intersection with the
Death Valley fault zone, then a 15 to 20km wide terrane south of the
fault must be equivalent to a 65 to 75km wide terrane north of the
fault. The Owlshead Mountains, which form most of this northern
terrane, must be internally extended by more than 300% to accommodate
such offset.

Lack of evidence for extreme extension in the Owlshead Mountains,
and evidence for extension east of the Owlshead Mountains (Noble, 1941),
led Davis and Burchfiel (1973) to propose that the Garlock fault once
continued as much as 30km farther east to a point of zero offset be-
tween the Kingston Range and the Silurian Hills. This hypothesized
eastern continuation of the Garlock fault has presumably been cut and
offset about 8km to the south by movement on the Southern Death Valley
fault zone, and is now inactive and entirely buried by Quaternary
alluvium. According to this hypothesis, offset of several tens of
kilometers occurred on the Garlock fault in the northern Avawatz
Mountains. Although recent geologic mapping places serious constraints
on the timing and nature of a through-going fault in the northern
Avawatz Mountains (Troxel and Butler, 1979; Brady and Troxel, 1981),
it is not possible to entirely rule out such a fault.

The Arrastre Spring fault is an older, high-angle fault extend-
ing from the southern end of the range northwestward to its intersection
with the Garlock fault (Fig. 2). It curves westward at its northern
end, and thus appears to be a branch of the Garlock fault (Noble and
Wright, Fig. 8, 1954). The core of the Avawatz Mountains, composed
almost entirely of pre-Cenozoic plutonic and metamorphic rock, forms
an elongate, uplifted block between the Arrastre Spring and Mule Spring
faults (Fig. 3). Structurally higher level Mesozoic metavolcanic rocks
form the exposed basement west of the northern Arrastre Spring fault,
indicating probable west-side-down vertical movement (Plates 1A, 2).

The Avawatz Formation is a thick sequence of siltstone, sandstone,
conglomerate, and sedimentary breccia deposited on both sides of the

southern half of the Arrastre Spring fault (Fig. 3). The formation is now tilted about 35° to 45° southward off the south flank of the range. It was first recognized and named by Henshaw (1939), who divided it into four members. He studied only the uppermost member in detail, as it contained late Tertiary vertebrate fossils. Ash beds from near the fossil-bearing beds yielded K-Ar ages of 10.7 and 11.0 m.y. (Evernden and others, 1964). Jahns and Engel (1949) recognized numerous sedimentary breccias within the Avawatz Formation.

MAP UNITS

Avawatz Formation

The Avawatz Formation is a large body of Tertiary clastic rocks underlying the south flank of the Avawatz Mountains. Henshaw (1939) named the formation and divided it into four members: A lower member of predominantly boulder and cobble conglomerate, a second member of siltstone, sandstone, and conglomerate, a third member of coarse sedimentary breccias, and a fourth member of sandstone, siltstone and tuff. Only the lower two members recognized by Henshaw were mapped in detail for this study, although all four members were mapped in reconnaissance. A previously unrecognized sequence of sandstone and siltstone occurs at the base of Henshaw's lower member on the west side of the Arrastre Spring fault, and is considered a separate member in this report. In addition, the third and fourth members defined by Henshaw have been mapped as one member because of problems in separating them.

Avawatz Formation, first member (Tav_1)

The first member of the Avawatz Formation crops out only on the west side of the Arrastre Spring fault (Fig. 3). It is approximately 615m (2,000 feet) thick and rests unconformably on Mesozoic metavolcanic basement. It is composed of siltstone, sandstone and conglomerate. Tan to greenish-gray siltstone is most prevalent near the base and to the west. Five to 30 centimeter thick sandstone beds containing symmetric to slightly asymmetric ripples are locally interbedded with siltstone. Sandstone and conglomerate occur throughout the first member,

but are most abundant immediately adjacent to the Arrastre Spring
fault and in the upper part of the member. Sub-rounded to sub-angular
porphyritic pink granite and quartzite clasts up to 20cm diameter,
and isolated, 1-2cm angular K-spar crystals occur as matrix supported
clasts in a very poorly sorted, brown to gray sand and silt matrix.
These may have been derived from weathering and erosion of quartzite
and pink granite that crop out extensively on the east side of the
Arrastre Spring fault. Conglomerate is more abundant in the upper
part of the member, although it rarely forms more than 10% of an out-
crop. Quartzite, pink granite, intermediate plutonic, and mafic plu-
tonic (Md?) clasts are generally 5 to 20cm diameter. Coarse sand and
gravel beds 5-40cm thick display sharp reverse grading at their base,
broad normal grading at their top, and are broadly lenticular in form
(greater than 20m wide). Scours with conglomerate fill occur locally.

Avawatz Formation, second member (Tav_2)

The second member of the Avawatz Formation is a boulder conglomer-
ate that crops out on both sides of both branches of the southern Ar-
rastre Spring fault. It rests conformably on the first member on the
west side of the fault, but rests unconformably on pre-Cenozoic crystal-
line basement elsewhere. Where it overlies the first member, its
basal contact is clearly defined over a 1/2 mile length where a sharp
upward transition occurs from a sequence containing less than 10% con-
glomerate beds to a sequence containing greater than 80% conglomerate
beds (Plate 1). The gradation occurs over an interval of about 0 to
5 meters. The second member is about 2,000m (6,700 feet) thick on
the west side of the Arrastre Spring fault (Plate 1B, cross section
A). It is much thinner on the east side of the west branch of the
Arrastre Spring fault (approximately 1,000m), and thins even more to-
ward the east branch (less than 800m). A relatively isolated exposure
of the second member, approximately 1 mi.^{2} in area, rests partly depo-
sitionally and partly in low-angle fault contact on Mesozoic and older
rocks in the central Avawatz Mountains (Plate 1A). Fault blocks of
the second member occur adjacent to this exposure within the east branch
of the Arrastre Spring fault. A small outcrop of the second member,
perhaps 20 meters across, rests depositionally on top of the carbonate

klippe east of the Arrastre Spring fault (Plate 1A), and another small patch rests depositionally on the upper Papoose Lake Member of the Bonanza King Formation in the central pendant (Plate 1A).

Conglomerate of the second member varies from bedded to massive. Bedding is most well developed in the upper part of the member, where clast supported to matrix supported conglomerate beds are lenticular, planar or contain channels. Boulder clast supported channel fill conglomerate occurs locally, with clasts typically 10-50cm across. Generally, the conglomerate is crudely stratified with sub-rounded to sub-angular cobbles and boulders set in a matrix of silt, sand and gravel. Stratification is defined primarily by differences in clast size, with coarse boulder beds interbedded with cobble beds and sparse gravel or sand beds. Cobble beds may have reversely graded, sandy bases and normally graded sandy tops. Most of the second member is re-markable for its crude to non-existent stratification and extremely poor sorting.

Conglomerate clasts are of all sizes and many lithologies. Most clasts are about 4 to 80cm across, but range up to 3 meters locally. Field descriptions from many areas indicate the following typical clast composition: 20-40% medium-grained mafic to intermediate plutonic; 15-20% coarse-grained, leucocratic granite; 5% quartzite; 5% limestone and dolomite; 5% fine-grained to aphanitic igneous rocks; 10% epidote-rich miscellaneous metamorphic rocks of unknown protolith; 5% gneiss; 1-3% dark hornblendite. Gneiss and hornblendite clasts are very distinctive and occur throughout the second member. Pink granite clasts are variable in abundance, and locally make up more than 50% of the clasts.

Dioritic to granodioritic, dark plutonic rock (Md) is the dominant rock type in the Avawatz basement, yet it is very poorly represented in the clast population. Hornblendite clasts have no known source terrane, and a source for the distinctive gneiss clasts is possible at the extreme southeastern end of the pre-Cenozoic Avawatz basement. Only the abundant pink granite clasts may be locally derived from ex-tensive exposures of granite (Mg_1, Mg_2) in the immediately adjacent basement complex.

Avawatz Formation, third member (Tav$_3$)

The third member is composed of sandstone, siltstone, and conglomerate. Its base is difficult to define, as it is gradational over 50 to 500 meters with the underlying second member. The third member, and the gradational basal transition zone, both thicken greatly toward the west. The basal contact is defined as the upward transition from less than 50% to greater than 50% sandstone beds. This contact is shown as a dashed line at two areas on Plate 1A, but is not drawn elsewhere because of poor exposure and/or difficulty in determining an unambiguous contact. Some areas within this transition zone are designated Tav$_{23}$. Above the transition zone, the average grain size steadily decreases upward into the thick siltstones and sandstones in the middle of the third member. The upper part of the third member is incompletely mapped, but is composed of sandstone, siltstone, and sparse conglomerate.

Sandstone and conglomerate in the lower part of the third member contain clast lithologies similar to those in the second member. The clasts are sub-rounded, and are almost always matrix supported. Conglomerate beds grade laterally into sandstone beds, and isolated cobbles occur within sandstone. Cobble conglomerate and gravely sandstone beds locally form repeating coarsening upward sequences, with each sequence 3 to 10 meters thick. Although a systematic paleocurrent study was not made, cross-stratification and cobble imbrication indicate transport directions to the south and west.

The middle and upper parts of the third member are composed of interbedded tan and gray siltstone and sandstone with sparse conglomerate. Gypsum layers up to 2cm thick, white volcanic ash beds up to 1m thick, and one 20cm-thick layer of conchoidally fractured milky white silica occur within the siltstone. Also present are unconformities with angular discordance varying from a few degrees to more than 60 degrees over a distance of several hundred meters. One of these is shown on Plate 1A as paired dotted and solid lines. Whereas folding and faulting are relatively minor in the first and second members, the middle and upper parts of the third member contain folds, faults, and angular unconformities. The upper part of the third member was mapped in reconnaissance only.

Avawatz Formation, fourth member (Tav$_4$)

The fourth member was mapped in reconnaissance only. Its base is placed at the base of the lowest carbonate breccia sheet above sandstone, siltstone, and conglomerate of the third member. This contact is offset by the inferred southward continuation of the west branch of the Arrastre Spring fault, but otherwise is a continuous, mappable contact broken only by recent alluvium in stream washes (Plate 2). Correlation of the basal contact across the inferred projection of the Arrastre Spring fault is not certain, but best explains the distribution of carbonate breccias. In western exposures, at least three breccia sheets are separated by sandstone, siltstone and conglomerate (Plate 1A). To the east, across the inferred trace of the Arrastre Spring fault, only one breccia sheet is exposed. This breccia is overlain by sandstone and siltstone containing the fossils and ash beds described by Henshaw (1939) and dated at 10.7 and 11.0 m.y. by Evernden and others (1964). Correlation of the sandstone and siltstone with similar rocks between the first and second breccia sheets on the west side of the fault is tentatively proposed, although ash beds have not been recognized in the western area.

Gypsiferous siltstone, sandstone, and cobble conglomerate interbedded with the breccias are very similar to sediments of the third member. Conglomerate cobbles are composed of sub-rounded felsic plutonic, metavolcanic, carbonate, and sandstone clasts. Clast imbrication indicates north directed transport. Cross-stratified and channelized beds are present locally.

Sedimentary breccia (br)

Sheet like masses of monolithologic or heterolithologic breccia are numerous in the Avawatz Formation. Most are about 8 to 30 meters thick, and extend laterally for 1/2 to 3 kilometers (Plate 1A). Abrupt lateral terminations are typical, with little or no evidence of thinning.

Most breccias are composed of pink, coarse-grained granite, a typical lithology of nearby Mesozoic basement exposures (Mg). A single heterolithologic breccia sheet occurs in the upper part of the first member. It is composed of mixed tan, coarse-grained dolomite marble

and fine-grained brown and black silty quartzite, overlain by massive orange to white quartzite. Breccias occur in the lower part of the second member and are abundant in the upper part of the second member and lower part of the third member (Plate 1A). Most of these are composed of coarse- to medium-grained pink granite, although one unusually thick and laterally continuous breccia is composed of white granite. Other locally occurring breccia types include fine-grained metavolcanic(?) rocks and banded gneiss. Breccias are sparse to absent in the middle and upper part of the third member, but form much of the fourth member. Unlike other breccias, these are composed almost exclusively of gray and white calcite marble and tan to brown calc-silicate rocks. Fusilinids occur in some clasts, confirming that the Pennsylvanian to Permian Bird Spring Formation was the primary source of this breccia. Magnetite iron ore forms lenses of monolithologic breccia within the carbonate breccia (Lamey, 1945). Large muscovite books occur within the iron ore deposit, and one yielded a K-Ar age of 147 m.y. (Table 1). The iron ore is presumably a skarn type deposit (Lamey, 1945) that formed at least 147 m.y. ago and was incorporated into a sedimentary breccia sheet in late Tertiary time.

All of the breccias are composed of closely packed angular to subangular clasts of diverse size with finer-grained material, including totally pulverized rock, filling interclast areas. The maximum clast size is highly variable. The largest clasts occur within the single white granite breccia sheet, which shows crude internal size grading with boulders up to 13m across near the top. Clasts in most breccias range up to about 1 meter across, but may have a maximum size of less than .5m, with most clasts less than 30cm across. The compositional layering in one banded gneiss breccia lines up from clast to clast, revealing a shattered mass of rock that has not been severely disordered. In contrast, one breccia mass along the main wash below the south pendant is composed of pink granite and fine-grained diorite(?) mixed in a complex, swirled pattern resembling marble fudge ice cream.

Depositional environment of the Avawatz Formation

The Avawatz Formation is characterized by rapid lateral and vertical facies changes. The thickest and coarsest boulder conglomerates

Sample #	Mineral	Formation	Long. (116)	Lat. (35)	K_2O	$^{40}Ar_{rad}$ (x 10^{-10} moles/gr)	$\%$ $^{40}Ar_{rad}$	Age (m.y.)*
140A	WR	Dike	30' 45"	36' 5"	3.29	.54514	27.4933	11.4 ± .1
140A	WR	Dike	30' 45"	36' 5"	3.29	.58750	29.1108	12.4 ± .1
759C	B	Tav_3	18' 47"	26' 46"	6.125	5.14372	14.2414	57.4 ± .3
759F	B	Tav_3	18' 47"	26' 46"	6.255	4.43526	8.32544	48.6 ±1.2
902	M	Tav_4	18' 20"	23' 48"	9.095	20.0217	82.7093	146.7 ± .9

* Age calculated using new decay and abundance constants:

$$^{40}K/K = 1.167 \times 10^{-4} mol/mol$$

$$\lambda_\beta = 4.962 \times 10^{-10} yr^{-1}$$

$$\lambda_\varepsilon + \lambda'_\varepsilon = 0.581 \times 10^{-10} yr^{-1}$$

Table 1. Potassium-argon ages of rocks from the Avawatz Mountains. Sample 140A is a Tertiary dike intruding foliated quartz monzodiorite. Sample 759 is an ash bed in the lower part of the third member of the Avawatz Formation, 759C is the coarse (-100 +150) biotite fraction, 759F is the fine (-150 +200) fraction. Sample 902 is a large muscovite crystal from a brecciated skarn deposit in the fourth member of the Avawatz Formation. WR = whole rock; B = biotite; M = muscovite.

form the second member immediately adjacent to the west branch of the Arrastre Spring fault. The conglomerates thin to the east and west and become finer-grained down section, up section, and to the west. Lacustrine siltstone at the base of the first member and the middle of the third mark the beginning and end, respectively, of coarse clastic deposition adjacent to the Arrastre Spring fault. The breccia sheets represent single sedimentary horizons. Conglomerates of the second member thicken westward below the lowest breccia sheet on the east side of the west branch of the Arrastre Spring fault (Fig. 5). This westward thickening wedge of conglomerate, resting depositionally on Mesozoic basement, records eastward transgression of the second member onto an uplifted and eroded basement block within the Arrastre Spring fault zone (Fig. 6).

The second member also rests on basement on the east side of the east branch of the Arrastre Spring fault, indicating that it transgressed completely across the Arrastre Spring fault zone and apparently completely buried the already topographically complex southern Avawatz basement complex. Although these clastic sediments are related to movement on the Arrastre Spring fault, they eventually completely buried it and large adjacent areas, indicating a source terrane that was perhaps one to several kilometers from the fault zone.

All of the ancient depositional environments inferred for sediments of the Avawatz Formation can be observed in modern depositional environments of the southern Death Valley-Mojave Desert region. Extremely poor sorting and stratification, and the coarse and angular character of the clasts in much of the second member, especially the middle part, indicate a talus slope to proximal alluvial fan environment. Underlying and overlying finer grained conglomerate and sandstone is interpreted as representing more distal alluvial fan environments.

Evidence of scour and fill sedimentation, with scours locally up to 2 to 3 meters deep, indicate local channels within a mid-fan or distal-alluvial fan environment. Cross-stratified sandstone indicates migrating ripples and small bars in a mid-fan to distal-alluvial fan or flood plane environment. Planar conglomerate beds, interbedded with sandstone and siltstone beds, are interpreted as flood related deposits in which slurries of coarse material were carried to the

Fig. 5. Simplified geologic map of southern end of west branch of Arrastre Spring fault. Monolithologic breccia beds within the second and third members of the Avawatz Formation are shown in black. A, A', A" is a distinctive white granite breccia. B, B', C, C' are pink granite breccias. These three breccia sheets are offset about .5km by the west branch of the Arrastre Spring fault. All sediments of the Avawatz Formation dip 35 to 50 degrees to the south. The base of the second member of the Avawatz Formation, where it rests unconformably on Mesozoic plutonic basement, is shown by paired dotted and solid lines. Note the westward thickening wedge of conglomerate above this unconformity and below a major breccia sheet. This indicates eastward onlapping of the Avawatz Formation onto an uplifted fault block within the Arrastre Spring fault.

Fig. 5.

Fig. 6. Schematic cross-section of Avawatz Formation.
Triangles represent sedimentary breccia sheets. Note lateral
changes in thickness of members. Both branches of the Arrastre
Spring fault are shown. Mv = Mesozoic volcanic rocks;
Mi = Mesozoic intrusive rocks; Tav = Tertiary Avawatz Formation.

distal part of an alluvial fan or to a fan-fringing lacustrine playa, where they were deposited over finer grained material. Mudcracks and gypsum deposits indicate that lacustrine siltstones were deposited in a shallow water, playa type environment.

Repeated coarsening upward packages of sandstone and conglomerate may represent prograding depositional lobes forming in front of advancing, levee bounded channels. Channel switching or destruction marks the end of an individual cycle and the beginning of a new one. This model is adopted from previously proposed models for deep-sea fan sedimentation (Mutti and Ricci Lucchi, 1978). It should be noted, however, that these coarsening upward packages are sparse, and the great majority of sediments show no orderly vertical changes in grain size, as would be expected for an alluvial fan environment with numerous, complexly braided channels.

The sedimentary breccias are interpreted as debris avalanches similar to the Pleistocene Blackhawk landslide in the southwestern Mojave Desert region (Shreve, 1968). The Blackhawk landslide is 3km wide, 8km long, and 10 to 30m thick. Clasts within it are closely packed in a three dimensional jigsaw puzzle manner, as in the Avawatz breccias. Color bands may extend from clast to clast without disruption, indicating a shattered but not highly disordered mass, as also observed in the Avawatz breccias. Both the Blackhawk landslide and the Avawatz breccia sheets have sharp lateral terminations, and both were deposited adjacent to an active fault scarp. Similar form, structure, and tectonic environment suggest these were deposited in a similar manner.

Field observations seemed to indicate that the abundant pink granite breccias were derived from the granites of Avawatz Peak (Mg_1, Mg_2) that are well exposed along the Arrastre Spring fault. However, modal mineral analysis of the granite breccias and the granites of Avawatz Peak indicates that they are distinctly different. The ratio of plagioclase to K-feldspar is generally significantly lower in the granites of Avawatz Peak than it is in the breccias (Fig. 7; Table 2). Only two out of the nine breccia clasts analyzed plot in the field defined by the granites of Avawatz Peak. This data indicates that the breccias were not derived from pre-Cenozoic rocks in the Avawatz

Fig. 7. Essential modal mineral composition of pink granite clasts (triangles) from monolithologic breccias of the second and lower part of the third members of the Avawatz Formation, and of pink granite (dots) in the pre-Cenozoic basement of the Avawatz Mountains. Note that seven out of nine clasts are dissimilar to the granite in the Avawatz basement, indicating that the breccias were derived from a different source of unknown location. Solid triangle is from large breccia block within the Arrastre Spring fault. G = granite; GD = granodiorite; QM = quartz monzonite.

Sample #	Type	Plag.	K-spar	Quartz	Total Light	% Plag.	% K-spar	% Quartz	Other	% Other
91	Mgmbr	160	269	189	617	26	44	30	19	3
766	Mgmbr	265	249	86	600	44	42	14	33	5
781	Mgmbr	138	271	207	616	22	44	34	20	3
798	Mgmbr	176	250	179	605	29	41	30	14	2
800	Mgmbr	76	384	199	659	12	58	30	28	4
807	Mgmbr	221	291	88	600	37	49	15	52	8
901	Mgmbr	131	273	205	609	22	45	34	13	2
1153	Mgmbr	170	213	223	606	28	35	37	4	1
1222	Mgmbr	170	252	190	612	28	41	31	8	1

Table 2. Modal mineralogy data for pink granite clasts from mono-lithologic breccias in the second and lower part of the third members of the Avawatz Formation. Percent plagioclase, K-feldspar, and quartz is normalized to P+K+Q = 100%. Percent other is color index.

Mountains. The source of the breccias, as with almost all the other clasts in the second and third members of the Avawatz Formation, is not known.

The lacustrine siltstone of the middle of the third member is significant in that it represents a transition to different sources of clastic sediments. Modern alluvial fans are relatively unidirectional in sediment transport direction, whereas lake-playas are generally fringed by distal alluvial fans supplying sediment from many directions. Consistent with this observation is the fact that conglomerates in the upper part of the third member and the fourth member have a volcanic rich provenance, unlike conglomerates lower in the section. In addition, the limestone breccia sheets of the fourth member are compositionally unlike any of the breccias lower in the section. A possible source of the limestone, known to be derived from the Bird Spring Formation, is in the upper Paleozoic carbonate rocks well exposed to the south in the Soda Mountains (Grose, 1959). The Red Pass Range, lying immediately south of the Avawatz Formation, is probably the source of the metavolcanic clasts in the upper part of the third member and the fourth member. Thus, lacustrine siltstones of the middle part of the third member marks the transition from northern and eastern sediment sources with a dominance of plutonic clasts and breccias, to southern sediment sources with a dominance of volcanic clasts and carbonate breccias.

Quaternary alluvial fan gravels, resting with profound angular unconformity on much of the Avawatz Formation, are significantly different sedimentologically from the underlying sediments. Quarternary fan gravels are well-stratified, well-imbricated, better sorted, and generally finer grained. Clast supported gravels may not even have any inter-clast matrix material. The lack of finer matrix material in these gravels is inferred to be the result of an arid climate. The higher percentage of matrix material in conglomerates of the Avawatz Formation, and other sedimentological characteristics described above, are inferred to be the result of a more humid climate and more rapid weathering during Tertiary time. A similar conclusion was reached by Henshaw (1939) based on paleontologic data.

Age: Henshaw (1939) assigned a lower Pliocene age to fossils found in
the fourth member of the Avawatz Formation, but later radiometric dating
of associated ash beds yielded ages of 10.7 and 11.0 million years
(Evernden and others, 1964). The fourth member can thus be confidently
assigned an upper Miocene age.

Several white, powdery ash beds occur within the lower part of the
third member. They are located 800m (2,500 feet) above the base of the
third member, and 3,400m (11,000 feet) above the base of the first
member of the Avawatz Formation (Cross section A, Plate 1B). They
are about 30cm to 1m thick, fine-grained, and not welded or appreciably
altered. Approximately twenty pounds of ash from this locality yielded
about one gram of approximately 98% pure biotite. The other two per-
cent consisted of silicic encrustations on biotite, and mafic minerals
not removed in the separation process. The biotite was split into a
100 to 150 mesh fraction and a 150 to 200 mesh fraction, weighing about
1/2 gram each. Potassium-argon dating of these separates gave ages
of 57.4 and 48.6 million years, respectively (Table 1, see Plate 1A and
2 for exact location). These ages are significantly older than anti-
cipated by perhaps 20 to 30 million years.

If these K-Ar ages are correct, then the middle and upper parts
of the third member span a time of at least 40 million years. This
is not inconceivable considering the recognition of angular unconformi-
ties within the third member, but if so, the unconformities represent
major depositional hiatuses. The possibility of older detrital bio-
tite in the ash bed, or xenocrystic biotite, can not be eliminated.
In light of these facts, the old K-Ar ages should be considered a
tentative age assignment only. The ash beds rest depositionally above
a continuous sequence of sediments extending down to the base of the
first member. Based on the ash bed dates, the first and second mem-
ber, and the lower part of the third member, are tentatively assigned
a Paleocene to Early Eocene age.

A hundred pounds of ash has been collected for fission track
dating to confirm the K-Ar age, and will be analyzed as soon as
possible.

Arrastre Spring Group (Tas)

The Arrastre Spring Group (new name) encompasses a number of iso-
lated bodies of predominantly conglomerate that crop out along the
northern half of the Arrastre Spring fault (Fig. 3). Most of these
rocks are composed of matrix supported conglomerate, with less abun-
dant sandstone, tuffaceous sandstone, siltstone, limestone, and vol-
canic flows and breccias. Clast compositions are variable both within
and between isolated bodies. Tertiary volcanic clasts are by far
the most abundant clast type, with pink granite clasts dominant lo-
cally in exposures at the head of Sheep Creek. Marble, quartzite,
and fine-grained metamorphic and dioritic (Md?) plutonic rocks are
also locally abundant. Conglomerate beds may form lenses 5 to 7
meters wide and 30cm to 1 meter thick. Unlike conglomerates of the
Avawatz Formation, these conglomerates have an abundance of silt-
and sand-sized matrix material, and dark brown, silty sandstone may
be volumetrically the most significant rock type overall. Fine-
grained siltstone and limestone are present in the northwesternmost
exposures of the Arrastre Spring Group.

Volcanic flows occur sparsely, with the most laterally continu-
ous of these located within conglomerate and sandstone at the head of
Sheep Creek (cross section E, Plate 1A). This particular flow is
approximately 5 meters thick and is composed of 60% to 80% brown
aphanitic groundmass with microlites and small phenocrysts of feldspar,
and green crumbly spots representing altered mafic phenocrysts. White,
silica-filled vesicles are shaped like rice or peanuts. Because this
flow occurs across several traces of the Arrastre Spring fault, dating
it will place a firm constraint on the timing of movement, and on the
age of at least part of the Arrastre Spring Group.

Arrastre Spring Group, volcanic rocks (Tasv)

This designation is applied to a body of volcanic rocks resting
on clastic sediments of the Arrastre Spring Group and on older base-
ment. It is located about 1/2 mile northwest of Arrastre Spring. The
rock is gray, non-vesicular, and has plagioclase microlites and pheno-
crysts up to 1/2cm and brown clots that may represent altered mafic
phenocrysts. Small fault blocks of this volcanic rock are also exposed

along the northwestern Arrastre Spring fault.

Age: The age of the Arrastre Spring Group is only poorly constrained.
It is younger than much of the movement on the Arrastre Spring fault,
which it partly overlies. Thus it is also younger than the lower two
members of the Avawatz Formation, which are offset significantly by
the fault. Some of the blocks of Arrastre Spring Group, however, are
cut and steeply tilted by the Arrastre Spring fault. Compositional and
structural variations between isolated blocks of Arrastre Spring Group
may indicate that they are of different ages. Tertiary volcanic acti-
vity in the Mojave region is generally unknown before about 22 m.y.
ago (Armstrong and Higgins, 1973), suggesting that the Arrastre Spring
Group is middle Miocene or younger in age.

Conglomerate of late Tertiary or early Quaternary age (QTcg)

This designation is used for conglomerates resembling Quaternary
alluvium, but are tilted or deformed to the point where a Quaternary
age assignment was questionable. Strata of this designation may be
moderately extensive in the low hills on the southwest flank of the
range.

Older alluvium (Qo)

This designation is used for well stratified fan gravel and con-
glomerate not in active stream channels, but which can be inferred to
have been deposited in Quaternary time. They are especially well
developed at the south end of the range where they cap broad, flat
ridges of easily eroded siltstone and sandstone of the third member of
the Avawatz Formation. On these ridges, stratified gravels up to 30m
thick rest unconformably on older Tertiary sediments. Bedding atti-
tudes in the gravels are parallel to flanking stream channel surfaces.
The parallelism of stratification in these gravels to modern stream
channels and washes, plus the well stratified character, indicate that
these sediments can be assigned a Quaternary age with a moderate degree
of certainty.

Alluvium (Qal)

This designation is used for unconsolidated or poorly consolidated
gravel and conglomerate in active or recently active stream channels,
washes, and talus fans. It was also applied to all the fan surfaces
on the east side of the range even though some of these surfaces are
probably as old as some of the older alluvium.

FAULTS

Arrastre Spring Fault

The Arrastre Spring fault extends across the entire length of the
Avawatz Mountains in a NW-SE direction (Figs. 2, 3). In the southern
half of the range it branches into two faults, an eastern and a western
branch. Previous geologic mapping indicated that it also branched at
its northern end (Jennings and others, 1962), however the eastern branch
of this northern set could not be located in spite of careful mapping,
and is considered to be non-existent. The other branch (actually the
only branch) curves westward toward the Garlock fault, and appears to
merge with it.

In general, the Arrastre Spring fault places Mesozoic plutonic and
Paleozoic metasedimentary rocks on the east against Mesozoic metavolcanic
and Tertiary sedimentary rocks on the west. Most of these rock types
also form fault bounded slivers along the northern Arrastre Spring fault,
especially in upper Sheep Creek canyon.

Attitude of fault

Control on the dip of the fault is moderate to poor, but its over-
all linear trace suggests it is sub-vertical. The western branch of
the fault, near the bifurcation point, clearly dips 65 degrees to the
west. This dip is also indicated by the interaction of the fault trace
with topography in the nearby area. Further south within the Avawatz
Formation, the west branch is clearly vertical within one well exposed
cliff face. In upper Sheep Creek canyon, the Arrastre Spring fault
branches into several sub-parallel faults, forming elongate fault blocks.
The southwestern most of these faults dips 26 degrees to the northeast
on the cliff face at the northwest edge of Sheep Creek canyon (Plate 1B,

cross section F). This fault probably retains its shallow dip only until it intersects other fault branches at depth. One kilometer southeast of this cliff face, the southwesternmost of two fault traces dips 57 degrees northeastward under a tilted block of the Arrastre Spring Group (Plate 1A). This approximate dip is present farther southeast under a block of Paleozoic carbonate. In conclusion, although dips are highly variable along the Arrastre Spring fault, especially when the fault is represented by more than one sub-parallel trace, the overall linear character of the fault indicates that it is sub-vertical.

The average strike of the Arrastre Spring fault is about N55W. The average dip of the Avawatz Formation is about 45 degrees to S10W. Restoration of the Avawatz Formation to a horizontal attitude by rotation of the Avawatz Mountains about a horizontal axis trending N80W results in a 40 degree rotation of the Arrastre Spring fault plane. Such rotation would give the Arrastre Spring fault a normal fault configuration at the time of faulting, with an average dip of 50 degrees to the southwest. This attitude of the fault should be considered only an estimate, but a normal fault configuration for the Arrastre Spring fault seems highly probable.

Magnitude of offset

Total offset on the Arrastre Spring fault is constrained by several features. Two large pendants within the Mesozoic plutonic complex each contain a west-dipping sequence of upper Precambrian to Devonian age metasedimentary rocks. Remarkably good alignment of these sequences across the east branch of the Arrastre Spring fault is prohibitive of more than about 2km of right-lateral offset, and suggest that total offset, both vertical and horizontal, is considerably less (Fig. 8; Plates 1A, 2). On the west branch, a distinctive trio of monolithologic breccia sheets appear to be offset about 1km in a right-lateral sense (Fig. 5). Since these rocks dip about 45 degrees to the south, this offset could be the result of an equal amount of southwest-side-down vertical movement, or of some combination of right-lateral and vertical movement. In any case, offset on this fault branch after

Fig. 8. Simplified geologic map of southern Arrastre Spring fault.

Horizontal lines represent all middle Cambrian through middle Devonian

carbonate rocks. Vertical lines represent all exposures of Cambrian-

Precambrian Wood Canyon Formation. The metasedimentary rocks shown

occur as west dipping sequences that line up across the east branch

of the Arrastre Spring fault and limit offset to less than about 2km

in a right-lateral sense.

Fig. 8.

deposition of the second member of the Avawatz Formation is small.
Coarse-grained, leucocratic pink granite (Mg_2 on Plate 1A) crops out
on both sides of both branches of the Arrastre Spring fault. Restora-
tion of five to nine kilometers of right-lateral movement on the
Arrastre Spring fault produces the most complete juxtaposition of
granite on opposite sides of the fault. However, granite is exposed
in several areas east of the northern Arrastre Spring fault, indicating
that no horizontal offset is permissible.

The structural level of exposure, or depth of erosion, is deeper
on the east side of the Arrastre Spring fault than on the west side.
On the west side of the west branch and main trace a thick sequence
of Tertiary clastic sediments (Avawatz Formation) rests depositionally
on Mesozoic metavolcanic rocks that are intruded by Mesozoic granite
(Mg). On the east side, a much thinner sequence of Tertiary clastic
sediments rest depositionally on pink granite and the structurally
deeper Mesozoic quartz monzodiorite complex, indicating that Mesozoic
metavolcanic rocks and the tops of shallow level plutons were eroded
before deposition of the Avawatz Formation. Erosion of these rocks
was the result of vertical movement on the Arrastre Spring fault, and
topographic relief resulting from this movement affected deposition
of the Avawatz Formation.

In conclusion, total right-lateral offset on the Arrastre Spring
fault is probably not more than about 10km, and may be considerably
less. Total right-lateral offset on the east branch of the Arrastre
Spring fault is less than 2km. Contrasting levels of early Tertiary
erosion and of thickness of the Avawatz Formation suggest that 2,000
to 3,000 meters of southwest-side-down vertical movement occurred on
the main trace and west branch of the Arrastre Spring fault. Verti-
cal movement on the east branch may increase southward toward the end
of the range.

Timing of offset

Deposition of the lower two members of the Avawatz Formation was
strongly influenced by movement on the Arrastre Spring fault. Lateral
thickness and facies changes indicate that clastic deposition occurred

adjacent to an active fault scarp. Most of the movement on the Arrastre Spring fault pre-dates deposition of the third member of the Avawatz Formation. Ash beds in the lower part of the third member are dated at 48.6 and 57.4 million years, indicating an Eocene to Paleocene age for most movement on the Arrastre Spring fault.

One body of Arrastre Spring Group sediments containing a volcanic flow is deposited across two traces of the Arrastre Spring fault. Anticipated fission track dating of mineral separates from this flow should provide a firm upper limit for the termination of movement on the Arrastre Spring fault. A Miocene age is suspected for this volcanic flow on the basis of regional relationships (Armstrong and Higgins, 1973).

Low-angle faults

Two major low-angle faults occur in the southern and central Avawatz Mountains. The southern of these two faults dips about 7 degrees west and places a klippe of Paleozoic carbonate rocks on top of Mesozoic quartz monzodiorite (Plate 1A, 1B, cross section D). Pink granite dikes in the carbonate klippe are also cut by the fault, and do not occur below the fault. This pink granite is characteristic of basement exposures to the west.

The northern low-angle.fault places Avawatz Formation, breccia, and Precambrian Johnnie Formation over quartz monzodiorite and various metasedimentary rocks. This fault was lost in breccias at its southern end, and its location to the north is inferred to extend under Quaternary alluvium for about 2km to a fault with a moderate to low-angle dip. Cross section B (Plate 1B) shows this fault as sub-horizontal on an east-west transect, but it actually dips southward here 5 to 20 degrees. Farther south where Tertiary rocks are cut, interaction with topography indicates that the fault dips north to northwestward, giving the fault surface an overall shovel like or concave upward form.

The nature of these two faults is not clearly understood. They have roughly similar geometry and are fairly close together, suggesting that they are related. In general, it appears that higher structural levels are represented in the hanging wall, and lower structural levels in the footwall, so that intermediate structural levels are missing and

the faults appear to attenuate an approximate vertical structural succession. Omission of intrusive contacts is apparent across both faults, and the depositional contact at the base of the Tertiary section is in part omitted. In addition, a west dipping fault lying entirely within the Stirling Quartzite near the north end of the northern low-angle fault cuts out most of the upper member of the Stirling and thus attenuates the section. This fault intersects the northern low-angle fault at a low angle just before both trend under Quaternary alluvium. These features all suggest that these faults are low-angle normal faults.

Westward direction of transport is suggested by two features. Three west vergent fold sets in the carbonate klippe near the low-angle fault surface have amplitudes of 5 to 20 meters and an interlimb angle of about 90 degrees. These folds may have resulted from westward transport and shear stresses in the hanging wall, or may be older features. Tertiary sedimentary rocks above the northern low-angle fault are tilted southeastward instead of south or southwestward as is typical elsewhere, suggesting that eastward tilting occurred in response to westward movement and rotation above a low-angle normal fault.

Total offset on these faults is poorly constrained. Quartz monzodiorite is present in both the hanging wall and footwall, restricting offset to perhaps less than 10 or 20km. Difficulties in mapping the northern continuation of the northern fault suggests that total displacement is small, perhaps less than 2km. The inferred northern extension of this fault under Quaternary alluvium places Cambrian Wood Canyon Formation against Wood Canyon Formation, further suggesting small offset. Where the northern extension of the fault re-emerges from under alluvium, a moderately west-dipping fault places lower Cambrian metasedimentary rocks in the hanging wall over quartz monzodiorite and Precambrian metasedimentary rocks in the footwall. This does not constrain the offset well, but it demonstrates the clearly irregular fault plane geometry. These features all support a small total offset, perhaps less than 2km.

The timing of offset is constrained to post-date deposition of the second member of the Avawatz Formation, but a younger age limit is lacking.

Garlock Fault Zone

The Leach Lake fault and Mule Spring fault are the two faults that form the Garlock fault in the northwestern Avawatz Mountains (Fig. 2). Fault scarps in Quaternary alluvium have been mapped along the Leach Lake fault (Clark, 1973), and along the Mule Spring fault (Roland Brady, personal communication, 1981). Displacement on the Leach Lake fault dies out in the northern Avawatz Mountains, but the Mule Spring fault curves around the northeast flank of the range and probably continues beyond the Avawatz Mountains (Fig. 2).

Leach Lake Fault

The Leach Lake fault was mapped in detail along about 3km of its length in the area east of lower Cave Springs Wash (Plates 1A, 2). In this area, the same plutonic rock (Mmc) occurs on both sides of the fault, limiting left-lateral offset to about 2 to 3km. The fault curves abruptly in the northern Avawatz Mountains from a roughly east-west strike to about S55E. The more southeasterly striking part of the fault is low-angle, dipping about 20 degrees to the southwest (Plate 1A). The cliff face traversed by this southeast-striking fault contains pervasive fault gouge several hundred meters thick with crude southwest dipping layering. At this locality, quartz monzonite of Cave Spring Wash (Mmc) is in low-angle fault contact above the Avawatz quartz monzodiorite. The fault is interpreted to be a reverse fault based on regional evidence for left-slip on the Garlock fault. Another low-angle reverse(?) fault with the opposite juxtaposition (quartz monzodiorite in the hanging wall) was mapped for a short distance north of the southeast trending part of the Leach Lake fault. Although mapping was not completed in this area, both of these faults appear to die out in the diorite complex to the southwest. An east-west trending linear body of carbonate rock within the intrusive complex lies between the Leach Lake and Mule Spring faults and is not cut or offset by any fault joining the two (Plates 1A, 2).

Mule Spring Fault

The Mule Spring fault was mapped only over a distance of about 3 to 5km in the northwestern corner of the area of Plate 2. Mapping

in this area by Roland Brady of U. C. Davis is in progress, and more easterly exposures of the fault have been mapped by Troxel and Butler (1979) and reported by Troxel and others (1979). Where mapped by the author, the fault is remarkably linear and interaction with topography suggests that it is vertical or steeply south dipping. Mesozoic quartz monzodiorite (Md) on the south side of the fault is juxtaposed with conglomerate of the southern Death Valley fault zone belt on the north side (Troxel and Butler, 1979).

Throughout its length, all rocks on both sides of the Mule Spring fault are dissimilar, with one possible exception. Conglomerate of the southern Death Valley fault zone belt contains clasts of gneiss, hornblendite, and pink granite, and is poorly sorted or massive. The matrix is a brown to red silt-sand-gravel mixture with crude to non-existent stratification. All these features are characteristic of the second member of the Avawatz Formation, and a tentative strati-graphic correlation with this unit is proposed. The conglomerate is somewhat finer grained that the Avawatz Formation conglomerates, and may represent a more distal facies. Ash beds are interbedded with the southern belt conglomerate in the Garlock fault zone (Roland Brady, personal communication, 1981). Fission track dating of these ash beds will test the validity of proposed stratigraphic correlation. Impli-cations of this correlation for offset on the Mule Spring fault will be discussed later.

Mormon Spring Fault

A small segment of the Mormon Spring fault was mapped by Troxel and Butler (1979) who considered it to be the southeastward continua-tion of the Mule Spring fault (Fig. 2). Where mapped by Troxel and Butler (1979), the Mormon Spring fault places Mesozoic quartz monzo-diorite over Quaternary fanglomerate along a west-dipping reverse fault. The quartz monzodiorite is crushed and pulverized over a dis-tance of 30 to 70 meters, whereas the underlying fanglomerate is disrupted for about 2 meters adjacent to the fault. The fault dips 39 degrees west at an exceptional exposure located north of Mormon Spring. Mormon Spring emanates from fault gouge along the fault zone at the base of the mountain. South of Mormon Spring, the fault places

quartz monzodiorite over quartz monzodiorite on a west dipping reverse fault (Plates 1A, 2; Fig. 2). This fault could not be located through much of the quartz monzodiorite because of the similarity of rock type above and below the fault.

The Mormon Spring fault is distinguishable from the Mule Spring fault by two features: 1) the Mormon Spring fault strikes north-south whereas the Mule Spring fault strikes northwest-southeast, and 2) the Mormon Spring fault places quartz monzodiorite over quartz monzodiorite, whereas the Mule Spring fault places quartz monzodiorite over an assemblage of completely different Precambrian to Tertiary rocks.

Two structural configurations are possible for the Mormon Spring fault: 1) it is the southern continuation of the Mule Spring fault, or 2) it is a minor splay fault off the Mule Spring fault, and the main trace of the Mule Spring fault is east of the quartz monzodiorite in the footwall of the Mormon Spring fault. In this second case, the southeastward continuation of the Mule Spring fault is now inactive and buried under Quaternary alluvial fan sediments.

Magnitude of offset

Quartz monzodiorite in the footwall of the Mormon Spring fault resembles quartz monzodiorite in most of the Avawatz Mountains, but is most similar to the rocks that crop out near Sheep Creek about 13km to the northwest. This suggests that total offset on the Mormon Spring fault is about 13km in a right-lateral sense, although this should be regarded as highly speculative. The distribution of dioritic rock in the Avawatz Mountains is such that the Mormon Spring fault could have a maximum of about 18 to 20km of right-lateral offset to perhaps 2 to 3km of left-lateral offset. Correlation of the Mormon Spring fault with the Mule Spring fault, and consideration of the maximum range of magnitudes of offset on the Mormon Spring fault, has major structural implications for the Death Valley and Garlock faults.

Death Valley-Garlock-Mule Spring Faults: An Overview

The Death Valley fault zone, southern Death Valley area, has been the subject of much speculation regarding its total strike-slip displacement. Geologic contacts in Precambrian rocks show a northeastward

truncation of successively older tilted units of the Pahrump Group by the unconformity at the base of the Noonday Dolomite. These contacts can be traced discontinuously in a linear west-northwest trending belt that crosses the southern Death Valley fault zone and permits no more than 8km of right-lateral displacement (Wright and Troxel, 1967). However, changes in facies and thickness of upper Precambrian and lower Cambrian formations suggested to Stewart (1967) that 80km of right-lateral strike-slip displacement occurred on the combined Death Valley-Furnace Creek fault zones. He proposed that this displacement has been accommodated in the southern Death Valley area by large scale, right-lateral, oroflexural folding, similar in structural style to the northern end of the Las Vegas Valley shear zone, and is thus compatible with the 8km maximum offset on the southern Death Valley fault zone. Unlike the Las Vegas Valley shear zone where rotation can be proven only in the Paleozoic sedimentary cover rocks, oroflexural folding in the southern Death Valley region would involve 90° rotation of a large terrane of Precambrian crystalline basement and overlying cover rocks. However, it is questionable whether such large rotations can be accommodated by bending and folding of crystalline rock, and except for apparent offset of facies and isopach trends, evidence for rotation of any magnitude is lacking (Wright and Troxel, 1970). It is also possible that changes in isopach and facies trends are primary features resulting from complexities in sedimentation and subsidence rates within the Cordilleran miogeosyncline, and do not result from later structural modification. To further complicate the problem, Troxel (personal communication, 1979) has mapped an alluvial fan sequence in the southern Death Valley fault zone that has been offset 20km from its source.

Controversy also exists concerning the extent of the Death Valley fault zone beyond the southern Death Valley area. Hamilton and Meyers (1966) proposed that it extends 250km southeastward into the mountain ranges north of Blythe, California, near the Colorado River. Although the Death Valley fault zone has been active in Quaternary time, there is no evidence for Quaternary movement on its proposed southeastern extension. Furthermore, geologic relations in the Old Woman Mountains and Kilbeck Hills preclude major right-lateral displacement in this area (Davis, 1977).

Evidence for large offset on the Garlock fault is compelling, and requires several tens of kilometers of left-lateral displacement in the northern Avawatz Mountains (Smith, 1962; Smith and Ketner, 1970; Davis and Burchfiel, 1973). Although Quaternary movement on the Garlock fault dies out in the Avawatz Mountains, Davis and Burchfiel (1973) proposed that an older Garlock fault continued eastward for 30km to a point of zero offset near the head of Kingston Wash. This allows gradual reduction of offset on the Garlock fault, with offset accommodated to the north by basin and range extension. The Mule Spring fault is the only branch of the Garlock fault on which major offset is possible. Since it curves toward the southeast, major left-slip on the Garlock fault would require several tens of kilometers of convergence and thrusting on the eastern segment of the Mule Spring fault. There is no evidence for convergence and thrusting of this magnitude, suggesting that the eastern segment of the Mule Spring fault, in its present configuration, is Quaternary in age. It may have had a linear, east-west trend in pre-Quaternary time, but if so, offset by, and interaction with, the Death Valley fault zone has resulted in major structural modification.

Nature of Mormon Spring fault

The Mule - Spring fault overrides the southern and central branches of the Death Valley fault zone (Fig. 2), and thus total offset on the southeastern end of the Mule Spring fault is the sum of movement on two branches of the Death Valley fault zone plus total movement on the Mule Spring fault west of its intersection with the Death Valley fault zone. Correlation of the Mormon Spring fault with the Mule Spring fault is not compatible with large offset on either the Mule Spring-Garlock or Death Valley faults. In view of compelling evidence for large magnitude offset on the Garlock-Mule Spring fault system, direct correlation of the Mule Spring and Mormon Spring faults is rejected. The Mormon Spring fault is herein regarded as a minor splay of the Mule Spring fault. Total Quaternary offset on the Mormon Spring fault may equal the magnitude of Quaternary offset on the Mule Spring fault, but pre-Quaternary offset on the Mule Spring fault may be greater by an order of magnitude. There is no evidence for pre-Quaternary offset on

the Mormon Spring fault.

The southeastern continuation of the Mule Spring fault is not exposed because it is not active and is covered by Quaternary fan sediments. A possibility is that the Mule Spring fault, Mormon Spring fault, and the eastern, buried extension of the Garlock fault all meet at a point now buried under alluvium but within a few hundred feet of the toe of the Avawatz Mountains, and located perhaps several hundred meters north of Mormon Spring (Fig. 2). An east-trending fault continuing east from this intersection is not consistent with an east trending magnetic lineament located several kilometers to the north (Plescia, 1981), but it is consistent with geologic data from the Avawatz Mountains. Another possibility is that the Mule Spring-Death Valley fault system extends southeastward for several kilometers and its displacement dies out somewhere north of Silver Lake (Davis, 1977). Alternatively, the fault system may curve around the footwall of the Mormon Spring fault and eventually join up with the Soda Mountain fault zone. These latter two hypotheses are not incompatible with Garlock extension hypotheses proposed by Davis and Burchfiel (1973). A geophysical study of the area east of Mormon Spring might be helpful in resolving this problem.

Folds

The first and second members of the Avawatz Formation are consistently tilted to the south-southwest, whereas the third member has been extensively folded. The largest of these folds is a northwest-plunging syncline whose limbs dip about 30 to 50 degrees. This fold involves siltstone and sandstone of the third member over at least two square kilometers in the southwestern part of the map area. The presence of unconformities within the syncline suggest that folding occurred during deposition. It is not certain, however, whether one episode of folding produced the unconformities and a second, later episode produced the synclinal structure, or if they are part of one continuous episode of deformation. Numerous other small folds and faults occur within the third member. At the extreme southeastern part of the map area (Plate 1A), Quaternary gravels are folded and tilted as much as 57 degrees.

Much of the folding in the Avawatz Formation may be due to
Quaternary uplift and south to southwestward tilting of the range.
The northwest trending syncline and other folds in the third and
fourth members may represent part of a northwest trending hinge
line separating steeply tilted fault blocks in the Avawatz Moun-
tains from the relatively inactive northeastern Mojave block.
The presence of these folds suggest NE-SW convergence and shorten-
ing across the hinge line.

CENOZOIC GEOLOGIC HISTORY

Cenozoic faulting and sedimentation in the Avawatz Mountains
can be divided into three distinct episodes: 1) early Tertiary
movement on the Arrastre Spring fault and deposition of the first,
second, and lower part of the third members of the Avawatz Formation,
2) late Tertiary, large scale left-lateral movement on a branch of
the Garlock fault extending eastward through the northern Avawatz
Mountains to a point of zero offset somewhere in Kingston Wash some
30km to the east, and 3) Quaternary development of the Death Valley-
Garlock-Mule Spring fault system in which all Quaternary offset on
the Death Valley and Garlock faults dies out in the Mule Spring re-
verse fault (Fig. 9). This last episode is responsible for the pre-
sent morphology of the Avawatz Mountains.

Early Tertiary Episode
Although early Tertiary potassium-argon ages on the ash bed with-
in the Avawatz Formation are considered tentative and possibly sus-
pect, their validity will be assumed for the purpose of this discussion.
These data indicate that coarse clastic sediment deposition adjacent
to the then active Arrastre Spring fault occurred during Paleocene and
Eocene time. The sedimentary rocks include the conglomerate and
breccia of the first, second and lower part of the third members of
the Avawatz Formation.

The Arrastre Spring fault is interpreted as a normal fault with
southwest-side-down vertical movement of two to three kilometers.

EARLY TERTIARY

MESOZOIC
DIKES

GOLER FM.

WITNET FM.

ARRASTRE
SPRING
FAULT

?——PROTO-GARLOCK——FAULT——?

AVAWATZ FM.

LATE TERTIARY

REGION OF 100% EXTENSION

GARLOCK   FAULT

QUATERNARY

PANAMINT
VALLEY
FAULT

SOUTHERN
DEATH
VALLEY
FAULT
ZONE

MULE SPRING FAULT

GARLOCK   FAULT

MILES

0   50

Fig. 9.   Three stage history of Cenozoic faulting in
northern Mojave-southwestern Great Basin.   Dashed line
represents inferred fault.   Dotted line represents
inactive fault.

Right-lateral offset of zero to eight kilometers is suggested by
the distribution of pink granite (Mg_2). The fault plane is now
roughly vertical, but restoration of the Avawatz Formation to a
horizontal attitude rotates the fault plane to a moderate, south-
west dip. This supports a normal fault origin, as opposed to a
reverse fault origin, for the Arrastre Spring fault.

Pink granite exposed in the basement rocks of the Avawatz
Mountains is not the source for most of the pink granite clasts in
the Avawatz Formation. Banded gneiss clasts and hornblendite
clasts are especially distinctive, but like most other clast types,
they are poorly or not at all represented in the local basement.
Banded gneiss clasts do resemble banded and ptygmatically folded
gneisses at the extreme southern end of basement rock exposures in
the Avawatz Mountains (p€g on Plates 1A, 2). Precambrian banded
gneisses are well exposed in the Halloran Hills 10 to 20km to the
east (DeWitt, 1980), but are unknown to the north, west or southwest
of the Avawatz Mountains. The source of the gneiss clasts, and pos-
sibly many other enigmatic clast types, is thus proposed to lie under
the alluvium between the southern Avawatz Mountains and the Halloran
Hills.

Correlation of the Arrastre Spring fault with the Soda Mountain
fault mapped by Grose (1959) was proposed by Davis (1977) based on the
similarity of rock types on opposite sides of the faults. Mesozoic
metavolcanic rocks are well exposed on the west side of both faults,
whereas structurally deeper plutonic and metamorphic rocks are widely
exposed to the east. Detailed mapping in the Avawatz Mountains sup-
ports this correlation, and supports Davis' (1977) conclusion that
a considerable, if not predominant, sense of displacement on both
faults was dip-slip.

Davis and Burchfiel (1973) correlated the Arrastre Spring fault
with the Panamint Valley fault on the west flank of the Panamint and
Owlshead Mountains. Westward curving of the northern Arrastre Spring
fault, and eastward curving of the southern Panamint Valley fault
(Muehlberger, 1954) might be interpreted as resulting from left-lateral
bending adjacent to the Garlock fault. However, elongate pendants

and inclusions in the Avawatz basement show a consistent northwest-ward "grain" or alignment from the southern end of the range to the Leach Lake branch of the Garlock fault, and thus show no indication of having been bent by drag adjacent to the Garlock fault. This evidence indicates that the curving trace of the Arrastre Spring fault is an original configuration, and correlation with the Panamint Valley fault would require a sigmoidal trace for a throughgoing fault. A sigmoidal trace is possible but seems unlikely, and I conclude that the Arrastre Spring fault is a Garlock fault related structure, and is not the offset equivalent of the Panamint Valley fault.

Lower Tertiary sedimentary rocks are rare in the southern Great Basin-Mojave region (Fig. 10). They do occur in the El Paso Mountains, where the Goler Formation is composed of up to 2,000m (6,500 feet) of non-marine clastic sediments (Dibblee, 1952; 1967). Dibblee (1952) divided the formation into two members: a lower conglomerate and breccia member up to 460m (1,500 feet) thick, and an upper arkosic sandstone and red mudstone member with local conglomerate lenses up to 1,800m (6,000 feet) thick. Vertebrate remains in the Goler Formation are considered to be of Paleocene age (McKenna, 1955; 1960).

The non-marine Witnet Formation in the Tehachapi Mountains is composed of a lower cobble conglomerate and pebbly arkosic sandstone overlain by arkosic sandstone and minor siltstone. It is as thick as 1,200m (4,000 feet), rests unconformably on Mesozoic granitic basement, and is unconformably overlain by Miocene volcanic and non-marine sedimentary rocks (Buwalda, 1954; Dibblee, 1967; Dibblee and Louke, 1970). Although the Witnet Formation contains no fossils or radiometrically datable rocks, Dibblee (1967) correlated it with the lower Tertiary Goler Formation to the east. Dibblee (1967) concluded that both the Witnet and Goler Formations were deposited by alluvial processes in a northeast trending lowland between the highlands in the present Mojave and Sierra Nevada regions.

The Witnet, Goler, and Avawatz Formations are the only known Eocene or Paleocene sedimentary rocks in the southwestern Great Basin or northern Mojave region (Fig. 10). They all are coarse clastic non-marine sedimentary rocks probably deposited in tectonically active fault bounded basins. Interestingly, they all fall on an east-west axis

Fig. 10. Distribution of early Tertiary rocks in the Mojave-southern Sierra Nevada, southern Great Basin. All known rocks of this age occuring northeast of the San Andreas fault are shown. Dotted line outlines position of Goler Formation relative to Mojave block before late Tertiary movement on the Garlock fault.

remarkably coincident with the Garlock fault. This co-linear arrangement is even more striking on a palinspastic base map on which left-lateral offset on the Garlock fault has been restored (Fig. 9).

The concept of an early Tertiary, proto-Garlock fault was first proposed by Nilsen and Clark (1975). This concept is supported by: 1) alignment of early Tertiary sedimentary basins along the Garlock fault, and 2) evidence that the Arrastre Spring fault is a Garlock related structure that was active in early Tertiary time. The nature of this proto-Garlock fault is uncertain, but the sense of movement on the Arrastre Spring fault suggests that movement on the proto-Garlock was right-lateral, opposite from the later, major left-lateral displacement.

The origin of such a fault is highly speculative, but may be the result of crustal contrasts between the Mojave and Sierra Nevada-Great Basin regions inherited from Mesozoic time or earlier. The Mojave block is a wide zone of primarily Mesozoic granitic basement with strontium isotope 87/86 initial ratios greater than .706 (Kistler and Peterman, 1978). Boundaries of strontium isotope .706 and .704 initial ratios trend roughly north-south within the central and southern Sierra Nevada basement, and on a reconstructed base map, project toward the central Mojave block (Kistler and Peterman, 1973). These lines swing abruptly westward in the southern Sierra Nevada and Tehachapi Mountains, and extend west of the Mojave block. If the .706 line indicates the western limit of Precambrian basement, then the Mojave block was underlain by a wide zone of Precambrian crust before pervasive plutonism. This wide zone of Precambrian crust, assembled in part during Permo-Triassic time, is roughly the pre-Mesozoic precursor to the Mojave block.

Mesozoic magmatism, confined to a relatively narrow northern belt in the Sierra Nevada and westernmost Great Basin, was distributed over a broader area to the south extending entirely across the Mojave region. The Sierra Nevada appears to have a relatively deep, narrow "root" or crustal base where seismic velocities sharply increase downward (Prodhel, 1979; Eaton, 1966). In contrast, the Mojave region has a shallower, flatter "root" or crust-mantle boundary (Prodhel, 1979; Hadley and Kanamori, 1977). These contrasts in crustal thickness are

presumably the result of differences in the distribution and concentration of Mesozoic magmatism, which may be in turn the result of pre-existing differences in crustal thickness, composition, and structure.

During latest Cretaceous to early Tertiary time, the locus of magmatism extended far inland, and the Mojave-Sierra Nevada-western Great Basin region was devoid of magmatism (Snyder and others, 1976; Coney and Reynolds, 1977). Inferred shallow subduction of the oceanic plate during this time was presumably responsible for underplating the western Mojave region with the Pelona-Orocopia-Rand schist (Burchfiel and Davis, 1981). This schist terrane may be tectonically correlative with the Franciscan assemblage, although its metamorphic facies is greenschist and not blueschist as in much of the Franciscan rocks.

Development of a proto-Garlock fault during early Tertiary time must be viewed in this tectonic context. Contrasts in crustal structure, composition and thickness between the Mojave region and the Sierra Nevada-southwestern Great Basin developed during Mesozoic time. The proto-Garlock fault apparently developed in response to contrasting stresses experienced by these areas during latest Cretaceous to early Tertiary shallow subduction along the Pacific margin of North America.

Late Tertiary Episode

Major left-lateral movement on the Garlock fault related to basin and range crustal extension during late Cenozoic time is well documented (Davis and Burchfiel, 1973; Smith, 1962). The location of a throughgoing Garlock fault in the northern Avawatz Mountains is, however, somewhat enigmatic. The Leach Lake trace of the Garlock fault can be positively ruled out because identical plutonic rocks occur on both sides. The other active trace of the Garlock fault, the Mule Spring fault, is a likely possibility. The only similar rock types on opposite sides of it are distinctive conglomerates containing gneiss, pink granite, and hornblendite clasts. Correlation of conglomerate north of the fault (southern belt facies of Troxel and Butler, 1979) with the second member of the Avawatz Formation, proposed herein, places little constraint on offset of the Mule Spring fault. This is because

reconstruction of the Garlock fault, following the model of Davis
and Burchfiel (1973), places the northern conglomerate northwest of
the Silurian Hills, and somewhat closer to its inferred source ter-
rane between the Halloran Hills and the Avawatz Mountains. However,
Roland Brady (personal communication, 1981) has identified similar
conglomerate in the Garlock fault zone south of the Mule Spring fault
indicating much less offset. The only remaining alternative location
for a throughgoing Garlock fault is the fault designated the southern
branch of the southern Death Valley fault zone by Troxel and Butler
(1979). Although considered by them to be a branch of the southern
Death Valley fault zone, this fault has an east-west trend. If new
evidence places serious offset constraints on this fault, or demonstrates
that it is not a branch of the Garlock fault, then a candidate for a
throughgoing Garlock fault as proposed by Davis and Burchfiel will be
lacking, and a complete re-evaluation of all evidence for major offset
on the eastern Garlock fault will be required.

Quaternary Episode

Right-lateral movement on the Death Valley fault has apparently
offset the Garlock fault, and caused reverse movement on the eastern
half of the Mule Spring fault. This reverse fault now accommodates
motion on both strike-slip faults, and Quaternary offset is demon-
strable on its entire trace.

The Leach Lake fault also has Quaternary offset and terminates
in the northern Avawatz Mountains where it changes abruptly from an
east-west trending high-angle fault to a shallowly southwest dipping
reverse fault. South of the Leach Lake fault, pre-Cenozoic contacts
trend southeast-northwest, whereas north of the fault, pre-Cenozoic
contacts within a fault block bounded by the Leach Lake and Mule Spring
faults trend east-west (Plate 1A). This contrast in pre-Cenozoic
structural "grain" may be due to counterclockwise rotation of the Mule
Spring-Leach Lake fault block from an originally southeast-northwest
orientation to an east-west orientation. This sense of rotation is
consistent with the left-lateral sense of movement on the Garlock
fault, and is supported by preliminary paleomagnetic studies of 11 m.y.

old dike rock in the Mule Spring-Leach Lake fault block (Lisa Kanter, written communication, 1981).

The Mormon Spring fault is considered a splay fault off of the Mule Spring fault, and probably has accommodated much of the Quaternary reverse movement on the Mule Spring fault. One locality on the Mormon Spring fault just north of Mormon Spring represents the southeastern-most exposure of demonstrable Quaternary offset on the Garlock-Death Valley-Mule Spring fault system. Farther south, the range front of the Avawatz Mountains probably was produced by Quaternary reverse faulting, but Quaternary fault scarps are not exposed.

South to southwestward tilting of the Avawatz Mountains is probably the result of reverse faulting on the Mule Spring fault. The Avawatz Formation is tilted as much as 50 degrees, and sediments of the third and fourth members are folded and faulted. Quaternary fan gravels are also faulted and folded, indicating that some of this deformation is Quaternary in age. This folding and faulting may be occurring above a line of flexure and convergence between the stable Mojave block and the Avawatz Mountains, now undergoing active uplift in response to reverse movement on the Mule Spring fault.

Bibliography

Armstrong, R. L., and Higgins, R. E., 1973, K-Ar Dating of the
    Beginning of Tertiary Volcanism in the Mojave Desert, Cali-
    fornia: Geological Society of America Bulletin, v. 84, p.
    1095-1100.

Brady, R. H., and Troxel, B. W., 1981, Eastern Termination of the
    Garlock Fault in the Avawatz Mountains, San Bernardino County
    California: Geological Society of America Abstracts with
    Programs, v. 13, n. 2, p. 46-47.

Burchfiel, B. C., and Davis, G. A., 1981, Mojave desert and en-
    virons, in Ernst, W. G., editor, The Geotectonic Develop-
    ment of California: Prentice-Hall, New Jersey, p. 217-252.

Buwalda, J. P., 1954, Geology of the Tehachapi Mountains, Cali-
    fornia, in Jahns, R. H., editor, Geology of Southern Cali-
    fornia: California Division of Mines Bulletin 170, p. 131-
    142.

Clark, M. M., 1973, Map showing Recently Active Breaks Along The
    Garlock and Associated Faults, California: U. S. Geological
    Survey Map I-741.

Coney, P. J., and Reynolds, S. J., 1977, Cordilleran Benioff zones:
    Nature, v. 270, p. 403-406.

Davis, G. A., 1977, Limitations on displacement and southwestward
    extent of the Death Valley fault zone, California: California
    Division of Mines and Geology Special Publication 129, p. 27-
    33.

Davis, G. A., and Burchfiel, B. C., 1973, Garlock fault of southern
    California - an intercontinental transform structure: Geolo-
    gical Society of America Bulletin, v. 84, p. 1407-1422.

DeWitt, E., 1980, Geology and Geochronology of the Halloran Hills,
    southeastern California, and implications concerning Mesozoic
    Tectonics of the southwestern Cordillera: Pennsylvania State
    University, University Park, Pennsylvania, Ph.D. thesis, 267 p.

Dibblee, T. W., Jr., 1952, Geology of the Saltdale quadrangle,
    California: California Division of Mines and Geology Bulletin
    160, p. 7-43.

Dibblee, T. W., Jr., 1967, Areal Geology of the western Mojave
    Desert, California: U. S. Geological Survey Professional
    Paper 522, 153 p.

Dibblee, T. W., Jr., and Louke, G. P., 1970, Geologic map of the Tehachapi quadrangle, Kern County, California: U. S. Geological Survey Miscellaneous Geologic Investigations Map I-607, scale 1:62,500.

Eaton, J. P., 1966, Crustal structure in northern and central California from seismic evidence: California Division of Mines and Geology, Bulletin 190, p. 419-426.

Evernden, J. F., Savage, D. E., Curtis, G. H., and James, G. T., 1964, Potassium-Argon dates and the Cenozoic mammalian chronology of North America: American Journal of Science, v. 262, p. 145-198.

Grose, L. T., 1959, Structure and petrology of the northeast part of the Soda Mountains, San Bernardino County, California: Geological Society of America Bulletin, v. 70, p. 1509-1548.

Hadley, D. M., and Kanamori, H., 1977, Seismic structure of the transverse Ranges, California: Geological Society of America Bulletin, v. 88, p. 1469-1478.

Hamilton, W., and Meyers, W. B., 1966, Cenozoic tectonics of the western United States: Reviews of Geophysics, v. 4, p. 509-549.

Henshaw, P. C., 1939, A Tertiary mammalian fauna from the Avawatz Mountains, San Bernardino County, California: Carnegie Institute of Washington Publication 514, p. 1-30.

Jahns, R. H., and Engel, A. E. J., 1949, Pliocene breccias in the Avawatz Mountains, San Bernardino County, California (Abstract): Geological Society of America Bulletin, v. 60, p. 1940.

Jennings, C. W., Burnett, J. L., and Troxel, B. W., 1962, Trona Sheet, Geologic Map of California, 1:250,000, California Division of Mines and Geology.

Kistler, R. W., and Peterman, Z. E., 1973, Variations in Sr, Rb, K, Na, and initial Sr 87/Sr 86 in Mesozoic granitic rocks and intruded wall rocks in central California: Geological Society of America Bulletin, v. 84, p. 3489-3512.

Kistler, R. W., and Peterman, Z. E., 1978, Reconstruction of crustal blocks of California on the basis of initial strontium isotopic compositions of Mesozoic granitic rocks: U. S. Geological Survey Professional Paper 1071, 17 p.

Lamey, C. A., 1945, Iron Mountain and Iron King iron-ore deposits, Silver Lake District, San Bernardino County, California, Part C, p. 39-58 in Jenkins, O. P., editor, Iron resources of California: California Division of Mines and Geology Bulletin 129, 304 p.

Longwell, C. R., 1951, Megabreccia developed downslope from large
     faults: American Journal of Science, v. 249, p. 343-355.

McKenna, M. C., 1955, Paleocene mammal, Goler Formation, Mojave
     Desert, California: American Association of Petroleum Geolo-
     gists Bulletin, v. 39, p. 5112-5115.

McKenna, M. C., 1960, A continental Paleocene vertebrate fauna from
     California: American Museum Novitates, n. 2024, 20 p.

Muehlberger, W. R., 1954, Geology of the Quail Mountains, San Bernar-
     dino County: California Division of Mines and Geology Bulletin
     170, v. 2, Map Sheet n. 16.

Mutti, E., and Ricci Lucchi, F., 1978, Turbidites of the northern
     Apennines: Introduction to Facies analysis: International
     Geology Review, v. 20, n. 2, p. 125-166.

Nilsen, T. H., and Clark, S. H., Jr., 1975, Sedimentation and tec-
     tonics in the early Tertiary continental borderland of central
     California: U. S. Geological Survey Professional Paper 925,
     64 p.

Noble, L. F., 1941, Structural features of the Virgin Spring area,
     Death Valley, California: Geological Society of America Bulletin,
     v. 52, p. 941-1000.

Noble, L. F., and Wright, L. A., 1954, Geology of the central and
     southern Death Valley region, California: California Division
     of Mines and Geology Bulletin 170, Chapter II, p. 143-160.

Plescia, J. B., 1981, The nature of the Garlock fault east of the
     Death Valley fault, eastern California: Geological Society
     of America Abstracts with Programs, v. 13, n. 2, p. 101.

Prodhel, C., 1979, Crustal structure of the western United States:
     U. S. Geological Survey Professional Paper 1034, 74 p.

Shreve, R. L., 1968, The Blackhawk Landslide: Geological Society
     of America Special Paper Number 108, 47 p.

Smith, G. I., 1962, Large lateral displacement on the Garlock fault,
     California: as measured from offset dike swarm: American
     Association of Petroleum Geologists Bulletin, v. 46, p. 85-104.

Smith, G. I., and Ketner, K. B., 1970, Lateral displacement on the
     Garlock fault, southeastern California, suggested by offset
     sections of similar metasedimentary rocks: U. S. Geological
     Survey Professional Paper 700-D, p. D1-D9.

Snyder, W. S., Dickinson, W. R., and Silberman, M. L., 1976, Tectonic
     implications of space-time patterns of Cenozoic magmatism in the
     western United States: Earth and Planetary Science Letters,
     v. 32, p. 91-106.

Stewart, J. H., 1967, Possible large right-lateral displacement along
    fault and shear zones in the Death Valley-Las Vegas area, Cali-
    fornia and Nevada: Geological Society of America Bulletin, v.
    78, p. 131-142.

Troxel, B. W., and Butler, P. R., 1979, Tertiary and Quaternary fault
    history of the intersection of the Garlock and Death Valley fault
    zones, southern Death Valley, California: unpublished report
    submitted to U. S. Geological Survey, Menlo Park, California,
    29 p.

Troxel, B. W., Jahns, R. H., and Butler, P. R., 1979, Quaternary
    and Tertiary history of offsets along the easternmost segment
    of the Garlock fault zone: Geological Society of America
    Abstracts with Programs, v. 11, n. 3, p. 132.

Wright, L. A., and Troxel, B. W., 1967, Limitations on right-lateral
    strike-slip displacement, Death Valley and Furnace Creek fault
    zones, California: Geological Society of America Bulletin,
    v. 78, n. 8, p. 933-949.

Wright, L. A., and Troxel, B. W., 1970, Summary of regional evidence
    for right-lateral displacement in the western Great Basin: A
    Discussion: Geological Society of America Bulletin, v. 81,
    p. 2167-2174.

PLATE 1A

GEOLOGIC MAP

OF THE

AVAWATZ MOUNTAINS,

SOUTHERN DEATH VALLEY REGION,

CALIFORNIA   by JON E. SPENCER   Scale 1:15000

N

GEOLOGIC MAP OF THE AVAWATZ MOUNTAINS

KEY AND CROSS SECTIONS

by JON E. SPENCER

PLATE 1B

CENOZOIC

Qal — Alluvium

Qo — Older alluvium

QTcg — Conglomerate of late Tertiary or early Quaternary age

Avawatz Formation
Tav₄ — fourth member
Tav₃ — third member
br — sedimentary breccia
Tav₂ — second member
Tav₁ — first member

Arrastre Spring Group
Tas < Tasv   Tas — conglomerate, sandstone, with sparse limestone, thin volcanic flows, tuff, and breccia
Tasv — volcanic flows and dikes

MESOZOIC

Mgm — granodiorite of Mormon Canyon
Md — Avawatz diorite complex
Mg₂ — younger granite of Avawatz Peak
Mg₁ — older granite of Avawatz Peak

Mmc — quartz monzonite of Cave Spring wash

Mi — Intrusive rocks, undifferentiated

Mv — Metavolcanic rocks

PALEOZOIC

PCu — Bird Spring Formation and Monte Cristo Limestone undifferentiated

Dsc — Crystal Pass member of Sultan Limestone

Dsiv — Ironside and Valentine members of Sultan Limestone, undifferentiated

€Du — Bonanza King Formation, Nopah Formation, and Sultan Limestone, undifferentiated

IPc — Bonanza King Formation through Bird Spring Formation, undifferentiated

€n — Nopah Formation

€bb / €bp   €bk
€bb — Banded Mountain member of Bonanza King Formation
€bp — Papoose Lake member of Bonanza King Formation
€bk — Bonanza King Formation undifferentiated

€c — Carrara Formation

€z — Zabriskie Quartzite

€wc — Wood Canyon Formation

PRECAMBRIAN

p€s — Stirling Quartzite

p€j — Johnnie Formation

p€gn — Gneiss

p€q — Quartzite

p€d — Dolomite, may include Noonday Dolomite

p€u — Quartzite, phyllite, dolomite, and limestone

mc — pre-Mesozoic metacarbonate

pm — pre-Cenozoic metamorphic and plutonic

CONTACTS

Intrusive or sedimentary contact: definite, approximate

FAULTS

High angle fault: definite, approximate, concealed and inferred

Low angle fault, teeth on upper plate, arrow indicates dip of fault plane

Gouge zone along fault

FOLDS

Anticline: upright, overturned   Syncline: upright, overturned

Axial trend of small syncline, anticline

BEDDING

Strike and dip of bedding: inclined, vertical, overturned, horizontal

FOLIATION

Strike and dip of foliation, inclined, vertical

SAMPLE LOCATION

.738

LANDSLIDE

COMPILATION GEOLOGIC MAP OF THE AVAWATZ MOUNTAINS SOUTHERN DEATH VALLEY REGION, CALIFORNIA

by JON E. SPENCER

PLATE 2

CENOZOIC

Qal — Alluvium

TQc — Conglomerate

Tcm, Tsc, Tv —
- Tcm — Conglomerate, breccia and sandstone
- Tsc — Siltstone, sandstone, breccia, and gypsum of the central Death Valley fault zone belt
- Tv — Andesite and rhyolite

Tsnf, Tsnc — Conglomerate (Tsnc), siltstone, gypsum, and halite (Tsnf) of the northern Death Valley fault zone belt

Tcs — Conglomerate of the southern Death Valley fault zone belt

Tc — Tertiary clastic sediments

Tmx — Tertiary breccia

Ta — Arrastre Spring group, includes conglomerate, volcanic conglomerates and flows, sandstone, and limestone

Tav₄ — Avawatz Formation, 4th member
Tav₃ — Avawatz Formation, 3rd member
Tav₂ — Avawatz Formation, 2nd member
Tav₁ — Avawatz Formation, 1st member

MESOZOIC

Mgm — Granodiorite of Mormon Canyon
Md — Avawatz Diorite complex
Mg₁ — Younger granite of Avawatz Peak
Mg₂ — Older granite of Avawatz Peak
Mg — Granite of Avawatz Peak, undifferentiated

Mmc — Quartz monzonite of Cave Spring Wash

Mv — Metavolcanic rocks

Mi — Intrusive, undifferentiated

PALEOZOIC

Pc — Paleozoic carbonate rocks

PRECAMBRIAN

pЄq — Late Precambrian and early Cambrian, dominantly clastic rocks. Includes Noonday through Zabriskie

pЄk — Kingston Peak Formation

pЄc — Crystal Spring Formation

pЄg — Gneiss

pЄu — pЄq, pЄk, pЄc, and possible older metasedimentary rocks, undifferentiated

FAULTS: high angle, low angle, inferred. Arrow shows dip of fault.

CONTACTS: definite, approximate.

K-Ar sample locality

Scale 1:62500

0   1   2   3   4   5
MILES

N

SOURCE OF MAP DATA

TROXEL AND BUTLER (1979)

JENNINGS AND OTHERS (1962)

JENNINGS AND OTHERS (1962)

SPENCER (THIS REPORT)

GROSE (1959)