

San Bernardino County Museum Association

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Ancient
Surfaces of
the East Mojave Desert

Storm over the Providence Mountains, East Mojave Desert.



edited by Robert E. Reynolds and Jennifer Reynolds

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Preface

The eastern Mojave Desert holds a 1.8 billion year record of events recorded in stone: Proterozoic igneous and metamorphic rocks; Paleozoic and early Mesozoic clastic and carbonate sediments; Cretaceous granitic intrusive rocks; Tertiary volcanic rocks; and Tertiary and Quaternary sediments.

The intervals of time between these intrusive, extrusive, structural and depositional events are important. These intervals record periods of relative calm in the history of the earth's surface: tectonic quiescence, long term weathering and erosion, and the development of extensive stable surfaces.

The oldest surface along our route lies at the base of the Paleozoic sequence. This erosional surface was created on granitic and metamorphic rocks 1.15 billion years old. These rocks are overlain by clastic and carbonate rocks that range in age from the earliest Cambrian and unconformably through the lowest Triassic Moenkopi Formation from the southern Marble Mountains to the Providence Mountains. Although complicated by deformation, these rocks may be also present in the New York Mountains and in the Clark Mountains.

An extensive erosional surface formed over the Mojave Desert in early Tertiary time. This surface is a pediment: a broad, gently-sloping bedrock surface with low relief developed by erosion and situated at the foot of much steeper mountain slopes. Erosional processes stripped early Mesozoic and Paleozoic clastic and carbonate rocks off the Cretaceous Teutonia pluton. This surface was developed before the deposition of the Peach Spring Tuff, 18.5 million years ago. The Peach Spring Tuff commonly overlies the erosional surface or sits on sediments deposited on that surface. We will cross the erosional surface developed on the Teutonia Batholith at the Miocene volcanic rocks of the Van Winkle Mountains, in western Lanfair Valley below the volcanic rocks of the Woods Mountains, and below the Hackberry-Vontrigger volcanic rocks.

The Peach Spring Tuff is an important marker of the ancient surface. It was deposited when a superheated cloud of lava fragments spread over a wide area of the western United States during a volcanic eruption near the Newberry Mountains, Nevada, 18.5 million years ago. The tuff covered 12,350 square miles from Seligman, Arizona to Daggett Ridge near Barstow. It was deposited in a matter of hours or days and thus marks a moment in geologic time that signals the beginning of crustal rifting in southern Nevada, western Arizona, and southern California. The surface defined by the Peach Spring Tuff has been tilted and offset by faults since late Miocene times.

Tertiary extensional tectonics and volcanism in the Woods Mountains, Castle Mountains and Piute Range occurred between 19.8 and 8.0 million years ago. The dips of volcanic flows and the restriction of distinctive flows to discrete basins help us picture the topography during this 10 million years of the late Miocene.

Latest Miocene basalts and "Pliocene" soils define periods of structural stability and constrain late Tertiary faulting. Ringed by mountain ranges, Lanfair Valley was sheltered from surrounding drainages that would otherwise have caused erosion. Modern climatic events and topography are reflected by the late Pleistocene development of the Kelso Dune complex and by a system of drainages now starting to breach the perimeter of Lanfair Valley.

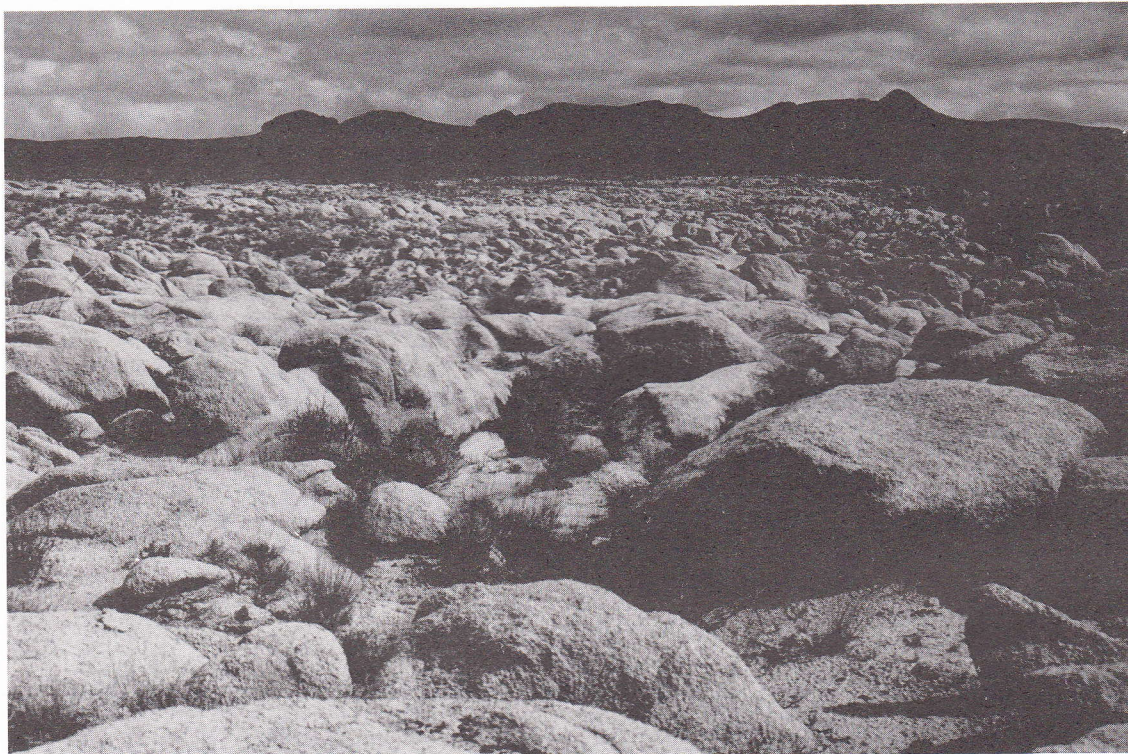
In "Ancient Surfaces of the East Mojave Desert" we will visit the early Tertiary surface, the late Tertiary and Quaternary surfaces, and look for events, both tectonic and erosional, that have impacted these surfaces.

Robert E. Reynolds, San Bernardino County Museum

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Ancient Surfaces of the East Mojave Desert

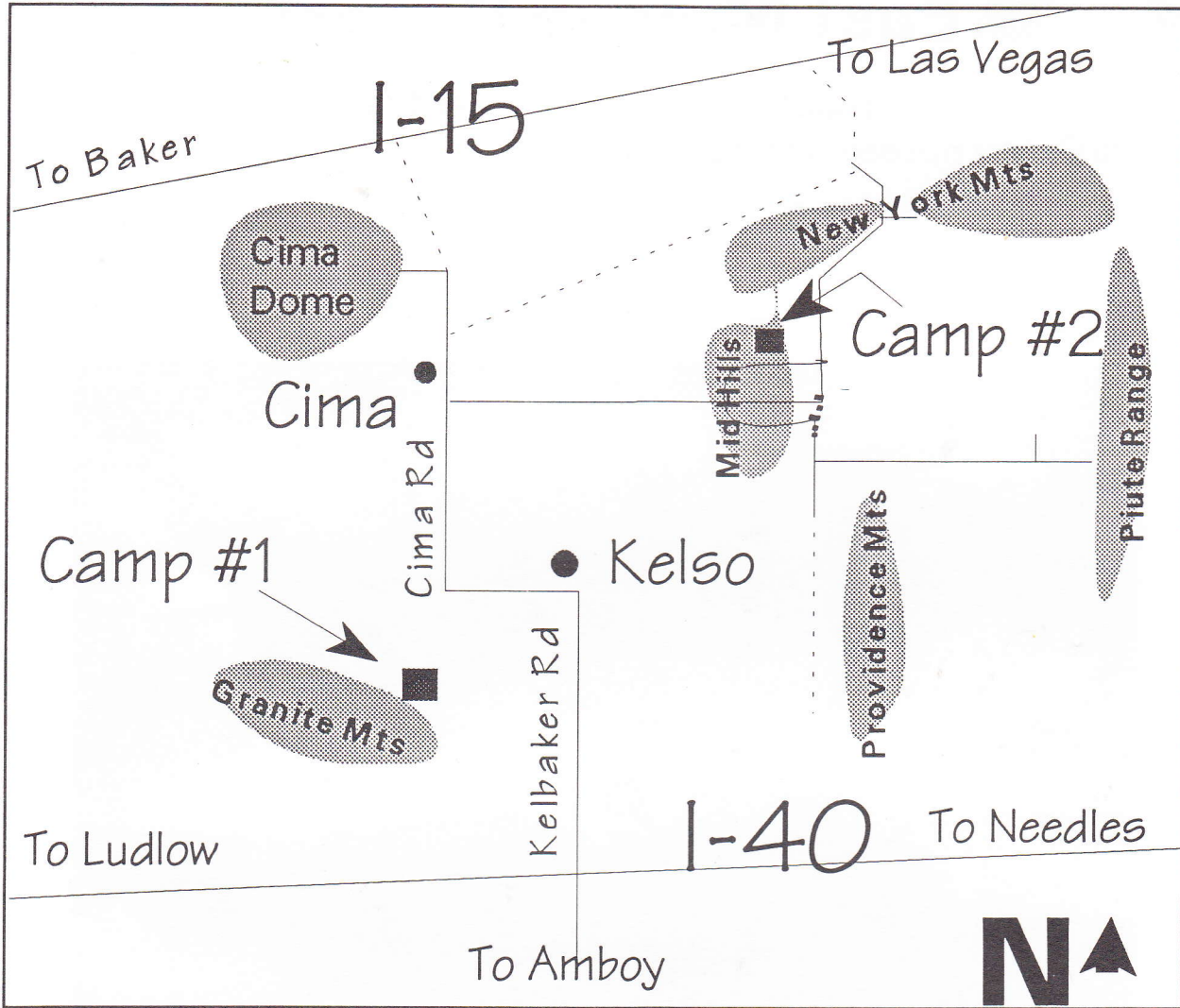
A volume and field trip guide prepared in conjunction with
the 1995 Desert Research Symposium



Erosion has exposed the ancient surface of this pediment, developed on granitic rocks north of the Vontrigger Hills and east of the Hackberry Mountains. R.E. Reynolds photograph.

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and
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Sketch map (not to scale) of field trip route in the East Mojave Desert.

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Field Trip Guide: Ancient Surfaces of the East Mojave Desert

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There are about 275 miles in the following two-day trip without reliable sources of gasoline. It is essential that you fill up your tanks in Twentynine Palms or Ludlow prior to the trip, and if necessary at the intersection of Cima Road and I-15 in the middle of the first day. Gas is intermittently available before 5 p.m. at Amboy. The trip ends at Ivanpah, where the closest gas is east on I-15 at Stateline, Nevada or west on I-15 at Cima Road.

En Route to Day One

Start at the San Bernardino County Museum in Redlands. Proceed west on Interstate 10, north on I-215 to Barstow, and then take Interstate 40 from Barstow to Ludlow (make sure to gas up at Ludlow). From Ludlow, take historic Route 66 east-southeast to Amboy. This route is best if you are concerned about gasoline mileage.

If you prefer highways to freeways, an alternate route from the Museum is east on Interstate 10, exiting north on Highway 62 and proceeding through Morongo Valley and Yucca Valley, to gas up in 29 Palms. In 29 Palms, go north on Utah Trail, then east on Amboy Road to Amboy. If you want to supplement this guide, the route from 29 Palms to Amboy is discussed in the guidebook, "Old Routes to the Colorado" (Reynolds and Reynolds, 1992). The area near the end of the trip is discussed in "Crossing the Borders," (Reynolds, 1991).

Please note: National Park Service assumed the administration of the East Mojave Desert Reserve as this guide was being prepared. As NPS integrates wilderness areas into the reserve, road closures are anticipated, and areas discussed in this guide may no longer be accessible by vehicle. Be flexible, bring your hiking boots, and obey all road closures. Due to variations in vehicles, your mileage may differ from that given in the cumulative and interval mileage.

0.0 0.0 Junction of Amboy Road and Route 66. TURN RIGHT (east) on old Route 66 toward Amboy. Amboy Crater is to the southwest.

0.3 0.3 Cross railroad tracks.

0.8 0.5 Amboy. This town, born in 1883 on the Atlantic and Pacific railway (now Santa Fe) (Gudde, 1969) was orphaned during the late 1960s when I-40 bypassed it to the north, and has remained nearly frozen in time. CONTINUE on Route 66 to Kelbaker Road.

6.7 5.9 TURN LEFT (north) onto Kelbaker Road. The range on the left is

the Bristol Mountains, here mostly Jurassic granitoids and pendants of Paleozoic marble. On the right across the valley are the Marble Mountains (see Glazner and Barkley, this volume).

7.7 1.0 The Blackjack Iron Mine and Snowcap Limestone Mine, two miles to the west, have ore bodies emplaced in marble and dolomite (Wright and others, 1953).

10.0 2.3 The Hope-New Method mine is at 10:00 to the west (see Jenkins, this volume). This mine, like many others in the southern Bristol Mountains, was explored for uranium.

10.8 0.8 Cross Orange Blossom Wash, which drains from farther north in the Bristol Mountains and western Granite Mountains.

11.4 0.6 Well-developed desert pavements are visible in the saddle to the west at 10:00.

13.4 2.0 Pedogenic carbonates and reddish arkosic sands are exposed in a low road cut. View of the northern Marble Mountains to the northeast at 2:00: note the angular unconformity and white ash beds. The Peach Springs Tuff



Figure 1. View from near Interstate 40 north to the Granite Mountains, with pediment in mid-ground. R.E. Reynolds photograph.

(18.5 Ma) occurs in this portion of the volcanic section (Glazner and Barkely, this volume). The thick section of volcanoclastic debris underlying the Peach Springs Tuff at Brown Buttes may be older than 20 Ma. Similar strata in the Bristol Mts are ~21 to 18 Ma (Miller, 1993).

18.2 4.8 Proceed under Interstate 40. Dacite domes near the highway are lithologically similar to 20-21 Ma dacite to the west (Miller, 1993). We now enter the Mojave National Preserve, the newest addition to the National Park Service's showcase lands.

18.7 0.5 The pediment at 10:00 to the northwest was produced by middle Tertiary erosion and is being re-exposed and dissected by erosion (Figure 1). A pediment is an erosional surface of low relief developed on resistant rock or crystalline bedrock. They commonly occur at the base of a bedrock mass with steep slopes that are almost devoid of fanglomerates. Excellent pediment surfaces form on the granitic rocks of the Teutonia Batholith, but here lie on younger granite (Howard and others, 1987). This pediment is overlain by ~21 Ma volcanic flows in the Van Winkle Mountains to the east. Elsewhere, the Peach Springs Tuff (18.5 Ma) is commonly deposited directly on these surfaces. Tectonic events, erosion and sedimentation, and volcanic eruptions in the last 18 million years have buried this surface in places with sands, gravels, and volcanic rocks. Burial was followed by erosion which repeatedly cut back to this surface, apparently to expose it approximately to its former extent. Generally, erosion does not go much deeper into the surface unless it is near a fault scarp.

21.0 2.3 View at 9:00 of granitic fanglomerates at base of Granite Mountains. Fans and fanglomerates in the southern Granite Mountains are rare because granitic rocks weather rapidly to sand-size debris that are removed by erosion. In contrast to the bare pediment seen on the southeast margin of the Granite Mountains, the perimeter of the northern Granite Mountains is covered by large volumes of fanglomerate dipping steeply north (Howard and others, 1987).

22.3 1.3 Pass left turn to microwave station; continue on Kelbaker Road.

22.9 0.6 The VanWinkle Mountains are at 2:00 (see Miller and others 1985). Miocene volcanics overlie a Miocene erosional surface developed on granitic and metamorphic rocks. Despite structural activity and eruptive events that caused debris to collect on this erosional surface, erosion over the last 20 million years (starting prior to deposition of the Peach Springs Tuff) has everywhere cut back to or close to this pediment surface. This late Tertiary pediment cut ~10 m into granite beneath the 21 Ma pediment.

24.8 1.9 Pass the intersection with Hidden Hills Road

25.4 0.6 Slow for 30 m.p.h. curve.

26.1 0.7 Microwave station at Granite Pass. Prepare to turn left across traffic.

26.2 0.2 TURN LEFT (west) just north of cattle guard and drive toward Snake Springs. The road bears west.

26.9 0.7 Continue past first left turn.

27.0 0.1 TURN LEFT (southwest) and proceed 0.3 mile into public camping area. Plants in this area include Mojave yucca, cats claw, juniper, cholla, ephedra, and piñon. The

spectacular boulders are quartz monzonite cut by aplike dikes. They contain feldspar phenocrysts larger than we will see in the granitics elsewhere on this trip. CAMP overnight.

Day One

On Day One our route takes us on the west side of the steep Providence Mountains (primarily Jurassic granite in the south and Paleozoic carbonate rocks in the north) past the Kelso Dunes to the granitic rocks of Cima Dome. We then return to Cima and go east through Cedar Canyon and the Mid Hills into buttes and mesas of Miocene volcanic rocks deposited around the Woods Mountains caldera, and end the day in the piñon pines of the Mid Hills Campground. During the day we will pass through Precambrian gneiss, Paleozoic carbonate rocks, and Mesozoic granitic rocks. Mid-Miocene erosional surfaces developed on these older rocks and were covered by middle to late Miocene volcanic rocks related to crustal thinning and extensional tectonics.

RETRACE route to Granite Pass and meet in the large turnout on the west side of Kelbaker Road across from the microwave station.

From the parking area, hike southwest up a graded dirt road to top of hill.

Stop 1. Due west are the Late Cretaceous granitic rocks of the Granite Mountains. We are well east of the Eastern California Shear Zone (ECSZ) (Richards, 1992) marked by the Bristol-Granite Mountains fault and the Soda-Avawatz fault zone near the Old Dad Mountains to the north-northwest. (Brady, 1992; Miller and others, 1985; Skirven and Wells, 1990; Skirven and Wells, 1992). To the north across the Kelso Dunes are the Marl Mountains and to the north-northeast, Cima Dome (Reynolds, 1991). Charleston Peak, elevation 11,918', is visible in the far distance above Cima Dome. Clark Mountain, elevation 7929', is in the middle distance just east of Charleston Peak. To the northeast is the thick section of Paleozoic carbonate rocks that make up the northern Providence Mountains. Jurassic granitoids make up the southern Providence Mountains. View S50°E is of Pliocene(?) fanglomerates in contact with the granitic rocks of the Horse Hills. View S70°E of Miocene volcanic rocks of Van Winkle Mountains that were deposited on a granitic pediment that runs north to the steep face of the Granite Mountains. View S70°W of Granite Mountain pediment and inselbergs.

We are at a drainage divide. The drainage to our north, from the north slope of the Granite Mountains and the west slope of the Providence Mountains, enters the Mojave River system and Soda Lake. The drainage from east to south comes from Lanfair Valley, the New York Mountains, and the eastern Providence Mountains, and runs along Fenner Valley into the Bristol-Danby trough at Cadiz (Reynolds and Reynolds, 1993). HIKE BACK TO VEHICLES.

0.0 0.0 Go north on Kelbaker Road.

2.6 2.6 Continue past junction at right with Arrowweed Springs Road. White rocks in the nearby hills are Late Cretaceous granite that is part of the same intrusive mass forming the eastern Granite Mountains. Darker rocks surrounding the Cretaceous granite are middle Jurassic quartz monzodiorite, quartz syenite, and more felsic granitoids (Miller and others, 1985).

- 4.6 2.0 Look back at 9:00 to fans coming north from the Granite Mountains.
- 5.8 1.2 Kelbaker Road bears right (north).
- 6.5 0.7 Proceed through junction with power line/gas line road. Lancaster (this volume) discusses buried paleosols six miles to the west.
- 6.8 0.3 Proceed past junction to Kelso Dunes. The Vulcan mine is at 2:00 (Shleman, 1995).
- 9.8 3.0 Road bends to the right.
- 10.2 0.4 CAUTION: watch for oncoming traffic and prepare to turn left and park on west side of Kelbaker Road.
- 10.3 0.1 TURN LEFT and PARK on section of abandoned roadway.

Stop 2. Walk west to the cut bank in the wash.

Fanglomerates derived from the Providence Mountains to the east interfinger with aeolian sand from the west (see Lancaster, this volume, for a discussion of the Kelso Dunes). To the southwest, deposition of large volumes of granitic conglomerate on the north slopes of the Granite Mountains is in sharp contrast to the lack of fanglomerates preserved on the pediment on the south margin of the Granite Mountains (Cahn and Gibbons, 1979). As proposed by Miller and Jachens (1995) a latest Cenozoic thrust fault may be elevating the Granite Mts. Return to vehicles and CONTINUE NORTH on Kelbaker Road, being careful of oncoming traffic.

- 10.7 0.4 Kelbaker Road bears left (west).
- 10.9 0.2 Slow, TURN SHARP RIGHT onto Vulcan Mine Road.
- 11.2 1.3 Caution: proceed across wash: rough crossing.
- 11.7 0.5 PULL OFF to right and PARK.

Stop 3: The Bishop Ash is found in exposures of deeply dissected fanglomerates in the third wash north (Figures 2, 3). The presence of this ash indicates that fans began deposition prior to 0.75 Ma, and that deposition continued



Figure 2. Stop 3. Bishop Ash is visible in the bank between and above the cacti. R.E. Reynolds photograph.



Figure 3. Detail of Bishop Ash preserved in alluvial fan at Stop 3. R.E. Reynolds photograph.

until the Late Pleistocene. It is uncommon for volcanic ashes to be preserved in coarse fanglomerates. The terraced surface of the fan has been developed in the last 27,000 years (McDonald and McFadden, 1994). The Bishop Ash itself defines an ancient surface at a point in time 75,000 years ago.

The Providence Mountains contain an interesting section of Early Paleozoic, Late Paleozoic, and Triassic sediments with invertebrate fossils (Hazzard, 1954; Merriam, 1954). Trilobites and other invertebrates (Mount, 1980) indicate an Early Cambrian age for the lower part of the section. Rugose and tabulate corals from the C&K mine section (Wilson, 1994) indicate a lower Permian (Wolfcampian) age for a portion of the Bird Spring group on the east side of the range. Half a dozen invertebrate taxa are known from the Lower Triassic Moenkopi Formation.

This area contains a spectacular flora. Our walk north takes us through barrel cactus, echinocactus, pencil cholla, golden cholla, Mojave yucca, buckwheat, and creosote bush (see Presch, this volume).

The U.S. Geological Survey (U.S.G.S.) has been measuring the rate and composition of modern dust deposition at 55 sites in southern Nevada and California since

Table 1. Modern dust deposition at test sides in the Providence Mountains area (Reheis and Kihl, in press).

Modern dust content and flux (Reheis and Kihl, in press) and soil accumulation rates (Reheis and others, in review) in the Providence Mountains area, 1984-1991. Silt and clay component of dust flux includes CaCO₃. LH, late Holocene; LP-EH, late Pleistocene to early Holocene.

Trap No.	Location	Weight % of dust		Sediment flux (g/m ² /yr)																	
		CaCO ₃		CaCO ₃		Salt		Silt		Clay											
		ave.	s.d.	Dust	Soils	Dust	Soils	Dust	Soils	Dust	Soils										
23	E side McCullough Mtns.	13.8	1.4	9.5	3.8	1.8	0.7	1.3	0.0	1.2	0.5	8.9	4.3	5.8	0.0	3.0	1.0	2.9	2.1		
27	NE of Cadiz Lake	10.9	1.3	11.0	4.3	1.5	0.3	1.1	0.0	1.3	0.4	5.9	0.9	0.4	1.7	3.1	1.1	0.9	2.3		
28	N side Providence Mtns.	9.6	2.8	9.1	2.1	1.3	0.4	0.3	1.5	1.2	0.3	8.1	3.8	1.1	2.2	4.3	1.9	0.1	4.0		
29	Cima volcanic field	9.3	3.4	11.9	6.4	1.6	1.0	10.8	0.7	1.7	0.5	13.0	4.1	23.9	1.3	3.7	0.7	5.3	3.8		
30	0.5 km W of Silver Lake	9.3	5.1	12.1	6.7	1.6	0.5	1.7	4.8	1.4	0.4	0.3	0.3	8.5	3.2	5.1	6.6	4.6	0.5	2.8	10.7
31	3 km W of Silver Lake	12.5	1.8	11.3	5.0	1.9	0.7	0.7	5.1	1.4	0.3	3.3	-0.7	9.6	4.1	8.2	16.1	3.9	1.0	5.1	8.1

1984, and is using the information to assess the content of dust in Quaternary soils and to provide information on paleoclimate and soil-forming processes. In addition, other researchers have used the dust data in studies of the origin of fault-cementing CaCO₃ in the area of Yucca Mountain, Nevada, and in modeling ground water compositions in the area of Ward Valley, California. The U.S.G.S. has seven dust traps in the eastern Mojave Desert and deposition of aeolian silt is being studied in this region. Data for sample sites in the vicinity of the Providence Mountains are given in Table 1, drawn from Reheis and Kihl (in press) and Reheis and others (in review). Several of the major points discussed in these two papers are illustrated by this small data set. (1) The modern dust flux is similar for sites both near and distant from large potential dust sources, such as the Silver Lake and Cadiz Lake playas; this implies that alluvial plains and bajadas presently contribute as much dust as do playas. (2) Major-oxide compositions show that aeolian materials are important components of the Av and upper B soil horizons. (not shown in Table 1; see Reheis and others, in review). (3) Modern dust fluxes are generally more than large enough to account for rates of accumulation of pedogenic materials in late Holocene soils. (4) Latest Pleistocene soil-accumulation rates are faster than late Holocene rates for sites near former pluvial lakes, reflecting a large dust-deflation and deposition event when the pluvial lakes desiccated (M.C. Reheis, U.S.G.S. Denver, writ. comm. to Reynolds, 1995). RETRACE route to Kelbaker Road.

13.5 1.8 TURN RIGHT (north) onto Kelbaker Road.
17.2 3.7 STOP, cross tracks at Kelso. The station was named for a railroad official when it was established on the San Pedro, Los Angeles and Salt Lake railroad (later Union Pacific) in 1906 (Gudde, 1969). Kelso Depot, on the right, is to become the interpretive center for the new Mojave National Preserve. TURN RIGHT immediately past the depot.

We are crossing the trace of the Cedar Canyon Fault as extrapolated by Jennings (1961), although no evidence exists for its location here. This fault is relatively young (Miller and Jachens, 1995) and may play an important role in the topographic evolution of western Lanfair Valley and the New York Mountains. We will cross the Cedar Canyon Fault later today.

21.1 3.9 Continue past Globe Mine Road.

25.9 4.8 Dawes. Continue past Macedonia Canyon Road.

The view northwest at 10:00 is toward a fault mapped by Jennings (1961) that strikes southwest along the low part of Cima Dome. Outcrops of the Peach Springs Tuff next to this fault raise the possibility that it cut the tuff. Observations by D.M. Miller suggest that the fault predates the Peach Springs, and that it may not exist in the orientation shown by Jennings. Three exposures of the Peach Springs Tuff can be found along this part of Cima Dome, and in one location the tuff lies on highly fractured and slightly mineralized Teutonia Adamellite (Beckerman and others, 1982). However, the fractures in the granite strike N. 15° W., nearly perpendicular to the fault shown by Jennings. The Peach Springs Tuff overlies fractured granite, indicating that fracturing is older than ~18.5 Ma. Several nearly east-striking lineaments cut granite and dikes upslope from the Peach Springs outcrops. These may be faults—they are poorly exposed—but, if so, they only have offset vertical dikes by a few tens of meters at most. No faults have been observed in this area by Miller that can be demonstrated to be Miocene or younger.

30.8 4.9 The Ivanpah Valley Fault runs across Cima Dome at 10:00.

31.7 0.9 Intersection with Cedar Canyon Road

36.3 4.6 Cima store and Post Office to west. Hills to east are underlain by ~1.7 B.Y. gneiss and granite.

36.4 0.1 Cross the first set of railroad tracks.

36.5 0.1 Cross second tracks and immediately TURN LEFT (northwest) onto Cima Road. Don't take the north-northeast Morningstar Mine road toward Nipton.

37.1 0.6 Cross the trace of the southwest-striking Ivanpah Valley Fault (Jennings, 1961).

37.9 0.8 TURN LEFT onto power line road and proceed west-southwest.

38.2 0.3 Cross the cattle guard just past the booster station. The road forks as you cross the cattle guard: TAKE RIGHT FORK (to west). We are driving on a thin sheet of arkosic sand obscuring the pediment surface of Cima Dome.

39.4 1.2 Cross second cattle guard.

40.2 0.8 Pass corral and road to Cut Springs.

41.7 1.5 **Stop 4** We have stopped on the south slope of Cima Dome in a forest of Joshua trees (*Yucca brevifolia*) with sparse juniper, creosote, beavertail cactus, large *Yucca*

schidigera, *Yucca bacatta*, and cholla (Cornett, this volume; Presch, this volume). Cima Dome superficially appears to be featureless, but we've have stopped where there are many small hills and contour irregularities. These hills are supported by dike rocks, aplites, and pegmatites of white quartz and pink feldspar. The Kelso Dunes are south-southwest, the Granite Mountains lie to the south. Kelso Wash drains the Providence and Granite Mountains and runs toward the Mojave River to the west. The Providence Mountains are to the south, the Mid Hills south-southeast, and the New York Mountains east. Notice the northeast-striking, southeast dipping monzonite porphyry dikes on the west slope of the Mid Hills that support ridges and cause ground water to rise near the surface. Historically the Providence, Mid Hills, and New York mountains were all called the Providence Mountains. Table Mountain is visible above the Mid Hills. On a clear day you can see Mt. San Jacinto and the Little San Bernardinos to the southwest. To the east-northeast are the Castle Peaks; Ivanpah Valley drains into the closed basin of Ivanpah and Roach playas.

Beckerman and others (1982) described complex Jurassic through Cretaceous lithologies of the Teutonia Batholith, now restricted to only Cretaceous rocks (Miller and others, 1991). Plutons of the Teutonia Batholith (Mid Hills adamellite, Live Oak Canyon granodiorite, Kessler Springs adamellite, Black Canyon hornblende gabbro, Teutonia adamellite, and Rock Spring monzodiorite) are ~95-97 Ma (DeWitt and others, 1984; Miller and others, 1991, 1994). The Ivanpah granite is Jurassic, and now not considered to be part of the Teutonia batholith (Miller and others, 1991).

We are standing on Teutonia adamellite. We will see all these plutons on this trip except the Kessler Springs adamellite, which occurs to the north-northeast in the Ivanpah Mountains.

We are on the southeast side of Cima Dome, a pediment cut into granites of the Teutonia adamellite. Like many plutons in the Teutonia batholith, it is intruded by numerous dikes. Virtually all of the small hills that stand above the Cima Dome pediment are underlain by resistant dike rock.

The pediment surface of Cima Dome may be in part older than the Peach Springs Tuff, as indicated by the deposition of that tuff on weathered granite a few km south of our position. The tuff dips about 7° downslope, a slightly greater dip than the present pediment surface. The tuff may have been deposited against some hills or on a pediment that was steeper than the present one. It is possible that the surface has been tilted down to the southwest since the tuff was deposited (Miller and Jachens, 1995).

Looking east, the gentle pediment surface can be traced toward the town of Cima. Quaternary alluvial sediment buries the pediment about half way to the town, although no perceptible change in slope marks the change from pediment to alluvium. A few km west of Cima is a small light-colored hill that was mapped by Hewett (1956) as Precambrian. The hill is composed of bouldery gravel, in which all clasts look like bedrock types up slope on Cima Dome. Cutting this gravel deposit on the northwest is a curving fault of generally northeast strike. The fault appears to have uplifted generally on the southeast with respect to the northwest. As described by Miller and Jachens (1995), this fault is probably part of the system of faults that extends

from Ivanpah Valley southwest to the Cima area, and then on to the Kelso Dunes area. Out of sight to the north, the fault offsets a prominent magnetic gradient, probably caused by the Kokoweef and Slaughterhouse faults (grouped as the Clark Mountain Fault by Hewett, 1956), by about 15 km in a left-lateral sense (Swanson and others, 1980; Miller and Jachens, 1995). Similar offset is observed farther north on a fault zone near the town of Nipton (Miller and Wooden, 1993).

Turn your attention to the southeast. This same fault probably is responsible for the change in topography between the gently-dipping pediments of Cima Dome and the >400 m rise in elevation along steep slopes and escarpments bordering the Mid Hills farther east. Granites underlying much of this escarpment are virtually identical to those underlying Cima Dome, and the rocks of the Mid Hills also have a well-developed Miocene pediment. Miller and Jachens (1995) propose that this escarpment is relatively young because it has not been beveled off, and that it indicates an up-on-the-southeast component of offset on the fault system extending down the axis of the valley. Late Pleistocene alluvium on the surface extending west from the Mid Hills indicate that uplift was early Quaternary or older (see Stop 10). The asymmetry across the valley holds from the Cima area, east of here, southwest to the Granite Mountains.

RETRACE route to Cima.

45.4 3.7 Stop, TURN RIGHT (south) on Cima Road. (If you needed to, you could turn left on Cima Road and reach I-15 and gas in 17 miles).

46.8 1.4 Slow, TURN RIGHT across railroad tracks.

46.9 0.1 Cima store

51.4 4.5 Cedar Canyon Road. Looking ahead you can see monzonite porphyry dikes cutting Teutonia quartz monzonite (Figure 4). PARK near the monument on the right side of the road. **Stop 5**, at the intersection of the Old Government Road. The road evolved from the Mojave Trail, a well-traveled trading route in California that joined the Colorado River with the Pacific Coast near Los Angeles and the central valley of the state. Fr. Francisco Garces was escorted over the trail by Mojaves in 1776, and Jedediah Smith followed it from the Mohave Villages on the Colorado River to the Mission San Gabriel in 1826. On his return in 1827, an Indian attack left most of his party massacred as they crossed the river; in part because of these hostilities, the trail was bypassed by immigrants in favor of the longer Spanish Trail, although mountain men Kit Carson and Ewing Young made the trip in 1829-30. The route was traveled again by railroad survey parties in 1853-54 (see Haenszel, this volume), and Beale took camels across Lanfair Valley and over the Piute Range in 1857. With the establishment of Fort Mohave at Beale's Crossing on the Colorado River, the army needed a supply route that was less expensive and faster than steamboats on the river, and they contracted with Phinneas Banning to take supplies along the route. Banning took 68 days to the journey with seven teams and wagons. (Banning was under contract for \$210/day, and probably saw little need for haste). Later in 1859, Winfield Scott Hancock took ten 8-mule wagons over

the trail, leaving Los Angeles on October 21 and arriving at Beales Crossing on November 5. This trip established the Government Road as a practical wagon route. So it remained until the coming of the railroad in 1883. Routed further south, wagon and later automobile traffic followed the tracks, and the Old Government Road was left essentially undisturbed (Casebier, 1975, 1987). TURN LEFT (east).

53.6 2.2 View south at 3:00 at series of monzonite porphyry dikes along west side of Mid Hills between Cedar Canyon Road and the Providence Mountains (Figure 4).

54.5 0.9 Pass the turn north to Dolly Varden/Death Valley mine (Figure 5). This mine followed mineralized rock along a moderately inclined fault. The fault may be part of a system of faults in the Ivanpah- Kelso valleys (Miller and Jachens, 1995).

55.2 0.7 Caution: cross Cedar Canyon Wash.

55.3 0.1 Cross the buried trace of the Cedar Canyon Fault which runs approx-imately east-west. The East Providence Fault runs south from this point past Mid Hills camp toward Foshay Pass. The East Providence Fault does not cross the Cedar Canyon Fault, giving further evidence for the right lateral sense of movement on the latter (Miller and others, 1991).

56.5 1.2 Caution; slow to 20 mph around a right turn. For a short while Cedar Canyon Road parallels the trace of the Cedar Canyon Fault, which here is oriented roughly east-



Figure 4. Monzonite porphyry dikes form resistant ridgebacks in the Mid Hills. R.E. Reynolds photograph.

northeast.

57.6 1.1 Black Canyon Road intersection. Proceed straight on Cedar Canyon Road.

57.8 0.2 Pinto Mountain is ahead at 12:00.

58.5 0.7 TURN LEFT (north) onto dirt track.

58.8 0.3 Look east to Pinto Mountain. On the south side of Pinto Mountain, the brown Peach Springs Tuff (18.5 Ma) caps low hills on the right (south). The steep south face of Pinto Mountain contains the light colored, columnar, jointed Wild Horse Mesa Tuff. (17.8 Ma). (McCurry, this volume).



Figure 5. The Death Valley (Dolly Varden) mine opened in 1907. Photograph courtesy of Larry Vredenburg.

The Peach Springs Tuff defines an older surface prior to the Woods Mountain Caldera event. The Peach Springs Tuff in places sits directly on the granitic pediment. The Wild Horse Mesa Tuff defines a slightly younger surface cut onto granite in some places and the Peach Springs Tuff in others. This suggests a significant amount of erosion or significant changes in relief in only 0.7 million years between deposition of the two tuffs.

CONTINUE NORTH on dirt track

60.0 0.2 Drop off terrace into wash.

60.4 0.4 **Stop 6.** PARK in sand wash and leave vehicles to view the complex intrusive relations between the Mid Hills adamellite and the Rock Spring monzodiorite (Beckerman and others, 1982). The Mid Hills adamellite is medium- to coarse-grained, porphyritic to equigranular, light tan, leucocratic monzogranite. It contains biotite as the sole mafic mineral in most places and commonly is cut by aplite and pegmatite dikes. The Mid Hills adamellite crops out widely over much of the Mid Hills and southern New York Mountains. It was dated by Ed DeWitt as about 93 Ma by U-Pb on zircon (Miller and others, 1994). The Rock Spring monzodiorite is quite different than the Mid Hills adamellite, being a porphyritic, dark-gray to brown, compositionally variable rock. It forms a zoned pluton in the Mid Hills southeast from this point. Common rock types are hornblende-biotite monzodiorite, quartz monzodiorite, and quartz monzonite; mafic hornblende-clinopyroxene diorite is present along the north edge of the pluton. It contains abundant mafic inclusions. Although considered by Beckerman and others (1982) to be Jurassic in age, a U-Pb date on zircon is about 97 Ma (Ed DeWitt, 1985, oral commun.).

Walk up the wash a few hundred meters to water-polished exposures of the two rock units. The darker Rock Spring monzodiorite is quite variable, both in grain size and composition, and contains many xenoliths. It even includes some hornblende gabbro. Light-colored dikes and veins of the Mid Hills adamellite cut across the Rock Spring monzodiorite. In a few places a pink granite is foliated and intermixed with the Rock Spring monzodiorite, and all are cut by dikes of the Mid Hills adamellite. This leucocratic phase related to the Rock Spring is uncommon and of uncertain significance. Intrusive contacts in the Teutonia batholith vary from smooth gradations to abrupt changes. All intrusive contacts between the Rock Spring monzodiorite and adjacent plutons are complex, like those we see in Cedar Canyon.

Returning down wash toward the vehicles, explore up the short, steep washes on the left (east). The Cretaceous granitoids are highly broken in these washes as they near the volcanic rocks of Pinto Mountain (Figure 6). Although a discrete fault plane is not exposed, these breccias are evidence of the Cedar Canyon fault. The fault is exposed both east and west of here. At this location, the Miocene Peach Springs and Wild Horse Mesa Tuffs have been faulted down to the south along the fault and now butt against granitoids. The volcanic rocks are 200 to 300 m thick here; that thickness records the minimum vertical offset on the Cedar Canyon fault. To the west, the Cedar Canyon fault crosses the nearly vertical East Providence fault zone (Miller and others, 1991).

RETRACE route to Cedar Canyon Road.

61.3 0.9 TURN RIGHT onto Cedar Canyon Road toward Black Canyon Road.

62.2 0.9 Intersection of Black Canyon Road. TURN LEFT (south) onto Black Canyon Road.

64.9 2.7 Pass corral.

65.1 0.2 Pass right turn to Mid Hills Campground.

65.8 0.7 We are crossing a divide separating three drainages. The Cedar Canyon drainage to the north runs to Kelso Wash and into the Mojave River. The Lanfair drainage to the east runs southerly and splits, part going to Fenner Valley and part to Sacramento Wash and the Colorado River. The drainage to the south joins drainages from the eastern Providence Mountains, runs to Fenner Valley and then into the Bristol-Danby Trough. We can see Miocene volcanic rocks of Pinto Mountain to the north, Table Mountain to the east, Woods Mountains to the south-southeast, and Wild Horse Mesa to the south. Tuffs and flows younger than 18 million years were deposited during the eruption of the Woods Mountains caldera. This portion of the Black Canyon Road cuts through the Cretaceous Black Canyon gornblende gabbro (Beckerman and others, 1982). This rock is dark because of hornblende and pyroxene and gives Black Canyon its name. The gabbro is intruded by the roughly coeval Mid Hills adamellite and Rock Spring monzodiorite. We will be driving through the adamellite on route to the Mid Hills campground. The Black Canyon drainage runs south to Fenner Valley.

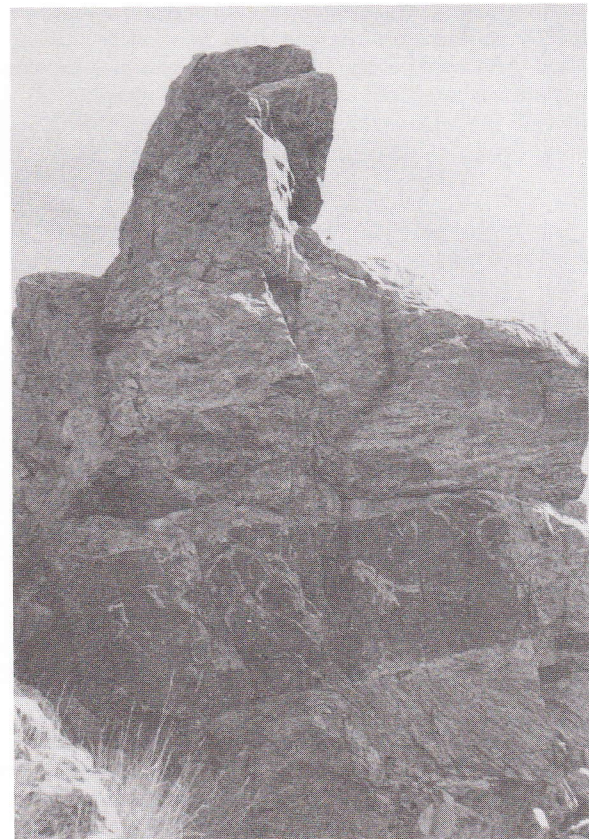


Figure 6. Tectonic breccia at Stop 6. R.E. Reynolds photograph.

66.7 0.9 Cattle guard

68.1 1.4 At the second cattle guard we are on the top of a Quaternary terrace that approximately parallels the Mid-Miocene pediment. Hackberry Mountain is at 9:30, the Woods Mountains at 10:00, the Old Woman Mountains at 12:00, and the Ship Mountains across Fenner Valley at 12:30.

68.4 0.3 Continue past the turn to Gold Valley.

68.8 0.4 View to west at 2:00 shows pediment and granitic inselbergs. This pediment and volcanic rocks on it dip shallowly south-southeast toward Fenner Valley. The dip may be the result of down-to-the-south tilting of a huge block comprising the New York and Providence Mountains and the Mid Hills.

70.3 1.5 View southwest of ashes and welded tuffs.

Prominently layered mountains and mesas to the southwest and east of Black Canyon Road are underlain by the Wild Horse Mesa tuff (abbreviated WHMT; McCurry, 1988). This tuff unit consists of a sequence of ignimbrites produced by major explosive eruptions at ~17.8 Ma (McCurry, and others, in prep.) from the Woods Mountains volcanic center, located several miles further east. Explosive eruptions occurred in three major cycles over a period of time of less than ~0.1 m.y., producing ~80 km³ of ash flow tuff. The tuff extends over an area of ~600 km², extending from Pinto Mt. on the north to the Blind Hills on the south, and from Hackberry Mt. on the east to the Providence Mts. on the west. Thickest deposits of the tuff, at least 320 m, overlie lacustrine sedimentary rocks at Wild Horse Mesa, having ponded there into a depression.

The upper third of the range to the southeast (western Woods Mountains) is capped by several rhyolite lava flows extruded from ring fractures bounding the western side of the Woods Mts. caldera (see McCurry, this volume). Erosion of the flows has produced a more ragged topographic outline than the mesas which are capped by the WHMT.

71.6 1.3 Pass the entrance to Hole-in-the-Wall campground.

71.8 0.2 TURN RIGHT into Hole-in-the-Wall visitors information center.

72.1 0.3 **Stop 7.** PARK in lot. Medium to light grey colored cliffs eroded within the oldest member of the Wild Horse Mesa tuff dominate the base of Barber Mesa at Hole-in-the-Wall campground. The tuff is rhyolitic in composition, sparsely sanidine phyric, generally glassy, and varies from unwelded to lightly welded. Excellent examples of standard ignimbrite flow unit features and textures are apparent in cliff exposures. Variations in erosional resistance of this unit, and variegation in color resulted from combinations of incipient devitrification, vapor phase alteration and crystallization of the ignimbrite during cooling. Anastomosing yellowish-brown veins of intensely altered tuff criss-cross many cliff faces, and resulted from channeling of hot gasses. Cavernous weathering is typical of the unit, and is splendidly expressed at the visitors center.

Note that blocks in colluvium contain strongly phyric, densely welded tuff. These were derived from the upper two members of the WHMT — they have largely been removed by erosion from the top of Barber Mesa. The uppermost is compositionally zoned, and has a highly alkaline chemistry

(i.e. it is mildly peralkaline) which is rare for Tertiary volcanic rocks in the Mojave Desert (McCurry, 1988; McCurry, and others, in prep.; Musselwhite, and others, 1989). The upper fifth of this unit also contains distinctive black flamme consisting of quartz trachyte, a product of magma mixing which took place late during the eruption. Best examples of these are in the western Woods Mountains and Wild Horse Mesa — they have largely been removed by erosion from the top of Barber Mesa.

A hike to the top of Barber Mesa will reward one with excellent and scenic views of the eroded ignimbrite plateau formed by the Wild Horse Mesa tuff. Just below the summit area one can also find exposures of pre-ignimbrite Plinian fall deposits ~1.5 m thick beneath the resistant densely welded tuff cap rock. Asymmetries in isopaches for Plinian fall deposits — which preceded all three ignimbrite eruptions — suggest a strong westerly or southwesterly prevailing wind at the time of the eruptions.

Ignimbrites exposed at Hole-in-the-Wall campground were derived from the Woods Mountains volcanic center (McCurry, 1988), about 3 miles east. Details of the evolution of the center are discussed elsewhere in this field guide (McCurry, this volume).

Evidence of Native American habitation in this area is abundant. Rustlers Rock Shelter (Sutton, this volume) is to the southeast. Wildhorse Canyon to the south contains interesting petroglyphs and pictographs.

RETRACE route to Black Canyon Road.

72.4 0.3 Stop, TURN RIGHT onto Black Canyon Road.

72.6 0.2 TURN RIGHT into Wild Horse Canyon.

73.2 0.6 Cattle guard.

73.7 0.5 Hiking trail to Mid Hills campground.

75.1 1.4 Wild Horse Canyon narrows.

75.9 0.8 Second cattle guard. The rhyolite dome on the right side of the road predates the outflow of the Wild Horse Mesa Tuff.

77.0 1.1 Continue past junction and PARK.

Stop 8. Mesas to the southeast and southwest of Wild Horse Canyon road are underlain by a sequence of six sedimentary and volcanic lithostratigraphic units. A sketch profile of the Opal Butte area (Figure 7) illustrates the stratigraphic sequence (Figure 8): from bottom to top: fluvial arkosic sandstone ("Summit Spring formation"), Peach Springs Tuff (18.5 Ma), unnamed basalt, an unnamed endogenous rhyolite dome, lacustrine sediments ("Winkler formation"), Wild Horse Mesa Tuff (17.8 Ma). Summit Spring and Winkler formations are described below; descriptions of volcanic rocks are in McCurry (1985). A short hike to the base of Opal Butte reveals excellent exposures of the contact between the WHMT and lacustrine sedimentary rocks. In some places weight of the ignimbrite deformed upper parts of the weak underlying sediments into gentle to strong upright folds a few meters across. Low temperature aqueous chemical interaction between the rhyolitic tuffs and carbonate sediments (limestone>dolomite) resulted in silicification of some parts of the lacustrine rocks. Silicification produced abundant opal deposits, most commonly in a pinch-and-swell pattern along bedding

planes. Many of these have been quarried for their fine multi-colored opal.

“Summit Spring Formation”. Tertiary? epiclastic rocks unconformably overlie Mesozoic granitoids and Precambrian metamorphic rocks from the northern Providence Mountains to the northeast of Wild Horse Canyon. Thus far unnamed, these rocks are referred to here informally as the Summit Spring Formation. They are weakly to moderately friable, undeformed, and are conformably overlain by the 18.5 Ma Peach Springs Tuff, suggesting they are late Oligocene to early Miocene age. Two lithostratigraphic units have been observed:

1) In areas to the west and east of the northern part of Wild Horse Canyon the deposits are light to medium gray, well to moderately bedded, moderately friable arkose, lithic arkose and conglomerate. Beds vary from massive to strongly cross-bedded. Boulders up to 80 cm across occur in some of the beds. Lithic fragments are similar to those occurring in nearby exposures of Mesozoic granitic and Precambrian metamorphic rocks. Volcanic clasts are absent. Deposits vary in thickness from a few meters (e.g., ¼ mile NNW of Opal Butte) to 40 m thick (e.g., ¾ mile ESE of Globe Mine) having been deposited in paleovalleys and also upon pediment slopes which appear to have trended and drained roughly south.

2) In the northern Providence Mountains, southeast of Summit Spring, beds of light to medium gray, poorly sorted, massive carbonate conglomerate and breccia are deposited unconformably upon Precambrian metamorphic rocks. Beds consist of subrounded to angular clasts of dolomite, limestone and sparse granitic clasts up to 1 m across in a light gray, fine grained, well indurated matrix of dolomite(?). Thickness of the beds decreases systematically from west to east. Thickest exposures are about 10 m thick a few hundred meters south of Summit Spring; thickness and grain sizes decrease to the east, and the unit wedges out completely a few hundred meters west of Barber Canyon. The wedge-like geometry of this unit suggests it was part of an east dipping alluvial fan on the flank of a north-trending range much like the existing Providence Mountains.

No direct time-stratigraphic correlation of the epiclastic rock units is possible because they do not come into direct contact and fossils have yet to be



Figure 7. Opal Butte (right); the Mid Hills pediment extends across photo center. R.E. Reynolds photograph.

found. However, they occur in similar structural and stratigraphic settings, suggesting they are of similar age. Preliminary analyses of the thickness and grain size distributions of the deposits suggests sediment transport from moderately rugged highlands to the west and north towards a paleotopographic depression centered on Wild Horse Mesa.

“Winkler Formation”. The Peach Springs Tuff is conformably overlain by 1 to 60 meters of lacustrine sedimentary rocks. These in turn are conformably overlain by the Wild Horse Mesa Tuff, constraining their age to between 18.5 and 17.8 Ma. Although apparently widespread — similar rocks are exposed at Pinto Mt. (10 miles north) and at Hackberry Mt. (14 miles east) — these rocks are thus far unnamed. They are informally referred to here as the “Winkler Formation,” in honor of the long-term caretaker of a nearby cabin which served as temporary shelter to many

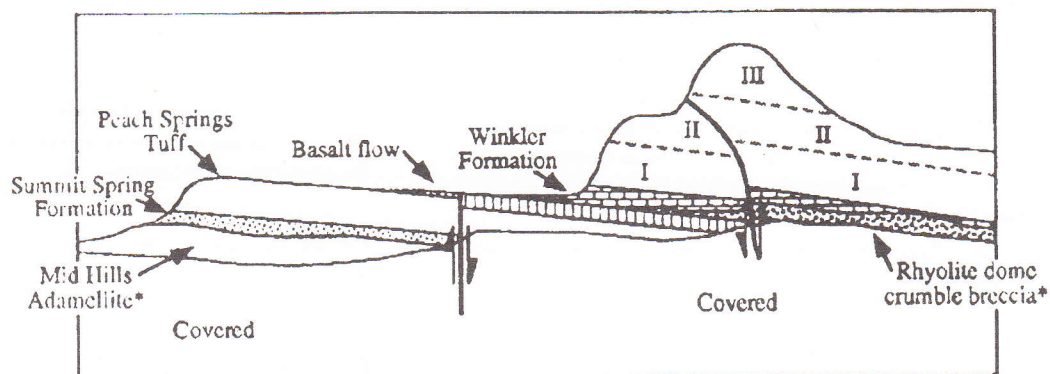


Figure 8. Geologic profile of Opal Butte area as viewed to the southeast from the Wild Horse Canyon Road. Roman numerals I-III represent crystallization and welding zonal units within the lower member of the Wild Horse Mesa Tuff (McCurry, 1985, 1988). Mid Hills adamellite is generally covered by a thin screen of colluvium at this location. Rhyolite crumble breccia is derived from the northern flank of a prominent endogenous dome; the Wild Horse Canyon Road passes through the southern end of the dome further to the south. Sketch by Michael McCurry.

travelers to the northern Providence Mountains.

Best exposures of the Winkler Formation rocks occur between the northern end of Barber Canyon and the north flank of Wild Horse Mesa. They form the prominent white slopes of Wild Horse Mesa to the southwest of our location, as well as the lower part of Opal Butte (to the southeast). They consist primarily of thin bedded, well sorted, very fine grained limestone and dolomite. Individual beds vary from 1 to 20 cm thick. They are interlayered in some areas with lithic arkose and conglomerate. The limestone and dolomite layers vary from massive to finely laminated, and are commonly fossiliferous. Numerous fragments of petrified wood and metasequoia-like leaf fossils occur in some places near Wild Horse Canyon.

Thicknesses of the formation vary spatially in a pattern which suggests deposition within scattered basins. Periodic subaerial exposure of the deposits is suggested by the occurrence of fenestral pores within some samples, suggesting a playa-like depositional environment. The idea that the lakes were often dry is consistent with lack of evidence for hydrovolcanic interactions between the overlying WHMT and top of the formation. In addition, the presence of laminations within some other samples implies a lack of bioturbation.

Dolomite and calcite were identified by X-ray powder diffractometry. Some layers consist of almost pure dolomite. Although calcite is most typical of playa carbonate deposits, dolomite also forms in these environments.

Faults on the north fronts of the mesa and buttes elevate older Miocene gravels and lacustrine sediments that sit on the surface of the granitic and metamorphic terraces to nearly the elevation of the Wild Horse Mesa Tuff. Before we reach Barnwell (Day 2) we are in terrain crossed by northeast-striking faults with north side up movement. To the west of Kelso are the northwest striking faults of the Eastern California Shear Zone and Mojave Desert. To the north are the north striking faults of the Basin and Range Province. The faults we have seen today – the Wildhorse Mesa Fault, the Cedar Canyon Fault, and the Ivanpah Valley Fault – strike northeast, most with a component of north side up offset. This offset has elevated the granitic and metamorphic rocks of the Mid Hills and New York Mountains, and possibly Cima Dome.

Wild Horse Mesa is a southerly occurrence of porcupine in the eastern Mojave Desert and Basin and Range Province. CONTINUE, but drive slowly and notice in 0.1 mile the outcrop on the east side of the road. This is an excellent example of a tectonic breccia: the Mid Hills adamellite with tumbled blocks of gneiss.

77.7 0.7 Pass third cattle guard. Willow Well windmill is to the left (west).

78.3 0.6 Summit. SLOW to 15 MPH, start downhill and prepare to stop for left turn into Macedonia Canyon. We are driving through the Mid Hills adamellite. Macedonia Canyon drains westerly into Kelso Wash and then to the Mojave River at Soda Lake.

78.6 0.3 Watch for oncoming traffic and TURN LEFT into Macedonia Canyon. Last one through closes the gate. Proceed with a firm foot; there are patches of soft sand on this road.

At this point we cross the buried trace of the East Providence Fault Zone of Hazzard (1954). Similar Proterozoic gneiss lies on both sides of the fault here, but elsewhere evidence for large offset is obvious. To the north, the Mid Hills adamellite is juxtaposed with Proterozoic gneiss, and to the south Paleozoic strata are juxtaposed with Proterozoic gneiss. The fault zone has had a complicated history, including reversal in the sense of offset. The net offset is down to the west, but in Miocene time offsets were the reverse. Peach Springs Tuff ramped up against a proto-Providence Mountains near the fault, suggesting that the fault may have created some of that topography. Splays from the East Providence Fault Zone that cut the Wild Horse Mesa Tuff are down to the east (Goldfarb and others, 1988). The East Providence fault zone runs north past Mid Hills camp and south past Columbia mine, Summit Spring, Bonanza King mine, and Winding Stair Cave to Foshay Pass (Miller and others, 1991).

79.2 0.6 Proceed past canyon to right.

79.7 0.5 Pass abandoned mines and rock foundations.

79.9 0.2 **Stop 9.** Exposed along the south wall of Macedonia Canyon is prominently striped, felsic gneiss and several black amphibolite masses. The gneiss is peraluminous, coarse- to medium-grained, biotite monzogranite with stripes defined by thin zones of more abundant garnet and biotite. Dismembered layers of amphibolite are scattered through it. Leucocratic segregations or veins in the gneiss are deformed in the highly folded complex. Blebs of biotite, about 0.5 to 1.0 cm diameter, are a conspicuous feature of the gneiss. Minor relic garnet is preserved in some of these blebs. Textural similarity with garnet-rich granite gneisses in the northeastern New York Mountains suggests to us that this rock once was garnet-biotite granite gneiss of igneous origin. Most garnet was retrograded when the nearby Mid Hills adamellite was emplaced. The roof of the Mid Hills pluton is exposed in several places along canyon walls near here, where it is a gently south-dipping surface (Goldfarb and others, 1988; Miller and others, 1991). The highly deformed to migmatitic character of the gneiss of Macedonia Canyon indicates involvement in an orogenic event, and regional information indicates the event was the Ivanpah Orogeny at about 1710 Ma (Wooden and Miller, 1990). Zircon studied in three samples yielded similar data indicating a U-Pb age of about 1.71 Ga for the gneiss of Macedonia Canyon, which is compatible with an early- to syn-orogenic origin. Amphibolite bodies at Macedonia Canyon are deformed similarly to the enclosing gneiss. They are basaltic (alkalic tholeiite) in composition.

The gneiss of Macedonia Canyon crops out over a 10 x 15 km area (Goldfarb and others, 1988), but shows only minor chemical, petrologic, and textural variation. The uniform granitic composition implies igneous origin. Two possibilities are: (1) a thick pile of rhyolite with a few basalt flows, or (2) a large plutonic complex cut by basalt dikes. We feel that locally crosscutting, distinctive textural phases reminiscent of porphyritic textures and otherwise remarkably uniform composition for the gneiss of Macedonia Canyon point strongly to a plutonic origin. Similar metaplutonic rocks crop out to the west in the Kelso Mountains and nearby parts of Cima Dome, making this

mass of gneiss one of the largest pre- or syn-Ivanpah orogeny masses of granite known in the Mojave Desert. Voluminous plutonism and mafic diking were part of the Ivanpah orogeny in the Providence Mountains. Considerable deformation took place during and (or) after intrusion.

CONTINUE down Macedonia Canyon.

As we leave this stop, look westward down Macedonia Canyon to Old Dad Mountain, marking the Eastern California Shear Zone (Richard, 1992).

80.8 0.9 **Stop 10.** Carbonate-cemented alluvial fan sediments define the 5°W slope of this drainage on the west face of the Mid Hills, approximately the dip of the active wash. These sediments and this slope postdate uplift of the Mid Hills and late Tertiary exposure of the pediment that we saw north of Hole-in-the-Wall. These sediments contain locally derived clasts and were interpreted by Goldfarb and others (1988) to be Pleistocene in age. RETRACE route to Wild Horse Canyon road; close the gate.

83.1 2.3 TURN LEFT onto Wild Horse Canyon Road. Proceed north-northeast.

83.7 0.6 Cattle guard.

85.0 2.3 Another cattle guard. Look north to Eagle Crags, large granitic tors of Mid Hills adamellite just east of the East Providence fault zone. We are on the Mid Hills divide that separates Lanfair Valley on the east from the Kelso Valley on the west. Continue past left turn to Eagle Crags.

85.8 0.8 TURN LEFT into Mid Hills Campground. A hiking trail runs south to Opal Butte in Wild Horse Canyon. CAMP.

End of Day One

Day Two

Our route will take us from the Mid Hills Campground north to Cedar Canyon, east across Lanfair Valley to the historic site of Lanfair, south to the rugged volcanic rocks of the Hackberry Mountains, then back north to Lanfair. We will go east on the Old Government Road across the exceptionally flat surface of Lanfair Valley to the dark-colored lava flows of the Piute Range, then return to Lanfair and go north on Ivanpah Road, take a short side trip to the New York Mountains, and continue north on Ivanpah Road to Barnwell. We will see outcrops and boulders of Peach Springs Tuff in Cedar Canyon, the Vontrigger Hills, and at Barnwell. Our emphasis will be on structural events, young sediments, and surfaces developed after the Miocene volcanic eruptions. As we drive from 5,000' to 3,000' elevation, we will see piñon and Joshua tree plant communities. The area has long been used by Native Americans and was settled, farmed, and ranched in the early 1900s after access was improved by railroad. It had a

particularly dense population in the teens and twenties during a wet series of years, when farms and orchards dotted the Lanfair Valley landscape.

Leave Mid Hills campground.

0.0 0.0 Stop, PROCEED LEFT (northeast) on Wild Horse Canyon Road.

1.3 1.3 View east across western Lanfair Valley. Notice the middle Tertiary pediment and granitic inselbergs. Miocene volcanic rocks of Pinto Mountain are at 10:00; Table Mountain is at 2:00.

2.0 0.7 TURN LEFT (north) onto Black Canyon Road.

4.0 2.0 Downhill grade

4.8 0.8 Stop, TURN RIGHT onto Cedar Canyon Road, paralleling the trace of the Cedar Canyon Fault.

5.7 0.9 Continue past turn to monzonite/adamellite discussion stop.

5.8 0.1 **Stop 11.** Stop, PARK on right side of Cedar Canyon Road where well-developed soils and pedogenic carbonates (Figure 9) are exposed at 5080' elevation in cuts on north side of road. Further west, sediments exposed in terraces cut by Watson Wash do not contain pedogenic carbonates and thus may be much younger than those in the eastern two-thirds of Lanfair Valley. We'll be traveling across the very flat terraced surface of Lanfair Valley controlled by well-developed pedogenic carbonates at elevations of 5080' in Cedar Canyon, 4575' in Barnwell, 4000' at Lanfair, and 3400' at Piute Gorge (Stop 13). Miller (this volume) discusses tilting of the Mid Hills structural block.

6.5 0.7 Cattle guard.

7.5 1.0 Continue past junction of private road on left (east)

7.7 0.2 SLOW for sharp left turn.

8.0 0.3 Cattle guard

8.3 0.3 Pass right turn towards Government Holes (at 2:00 near the large cottonwood tree). This public water source was developed in 1859 by Phinneas Banning when,



Figure 9. Indurated sediments north of Rock Springs. R.E. Reynolds photograph.

during his leisurely contract wagon trip to the Colorado River, he found Rock Springs dry and dug a well two miles east at a small seep. First known as Bannings Well, it soon became known as Government Holes. It served as one of only two public sources of water for Lanfair Valley homesteaders after the turn of the century when the Rock Springs Cattle Company denied access to water sources on their property. The farmers and the cowboys were not friends. Cattle raised havoc in the cultivated fields, and homesteaders were happy to help themselves to "free cattle." Disputes escalated through the 'teens, and continued until the homesteaders had for the most part abandoned their holdings in 1925. It was on November 8, 1925 that the last shoot out in the "Lanfair Valley Wars" took place at Government Holes. between two gunmen hired by the Rock Springs Cattle Company (Government Holes was public water, but the cattle company controlled it by intimidation). One gunman remained loyal to the cattle company; the other had changed coats with the lure of free cattle. They killed each other inside a cabin at the well (Casebier, 1987). Bullet holes could be seen in the cabin as late as the 1950s.

9.7 1.4 Continue past rock cabin turnoff. SLOW for descent into Watson Wash, named for a local homesteader (Casebier, 1987).

10.2 0.5 TURN RIGHT toward Fort Rock Springs. Bear right; avoid left turns. We are south of the Cedar Canyon Fault and are passing terraced gravels that do not appear to contain well-developed soils or pedogenic carbonates. This suggests that debris of ancestral Watson Wash filled a depression that formed later than the pedogenic carbonate seen at Stop 11, seven miles to the east. Clasts in cross-bedded terrace sands seen on west side of Watson Wash include Mid Hills adamellite, mylonitized Mid Hills adamellite, and quartz and monzonite dike rocks, all representative of the middle New York Mountains west of 4th of July Canyon, and Pinto Mountain volcanics.

10.3 0.3 TURN RIGHT into Fort Rock Springs and PARK.

Stop 12. This is the type locality of the Rock Spring monzodiorite of Beckerman and others (1982) and can be examined in excellent water-polished exposures. It is much more uniform in composition than the example we saw in Cedar Canyon. It is porphyritic, dark-gray to brown, and is a hornblende-biotite quartz monzodiorite containing abundant mafic inclusions. Quartz is about 15% and mafic minerals greater than 25%, which is an unusual composition for Cretaceous plutons in the eastern Mojave Desert (Miller and others, 1982). Beckerman and others (1982) showed that the xenoliths are cognate—that is, they are part of the magma lineage that produced the monzodiorite.

Camp Rock Spring was an official army camp from December 30, 1866 to January 2, 1868. Attacks on travelers by Paiutes in the middle 1860s caused great concern about the safety of the mail being routed along the Old



Figure 10. Rail bed of the 1893 Nevada Southern railway near the Hackberry Mountains. R.E. Reynolds photograph.

Government Road between Hardyville and Camp Cady. Locally, miner Moses Little, was killed by Indians in the Macedonia mining district in 1866. The army established a garrison at Rock Spring, where there was already a mail relay station, with twenty cavalymen to provide escorts for the mail carriers (Casebier, 1973). Abandoned mine tunnels from the old Macedonia district were used to store supplies (Vredenburg and others, 1981). Within nine days, five of the twenty had deserted. The camp commander sent men in pursuit; they deserted, too. By January 20 only eleven men were left at Camp Rock Spring; despite replenishment of troops from Camp Cady, fourteen men deserted the post during January, 1867. Casebier (1973) suggests that it was not just the harsh desert conditions that caused the soldiers to desert their post: many found the army a successful way to obtain an outfit with which to go prospecting in the Arizona mines. Camp Rock Spring was dissolved in January, 1868, but remained as an outpost of Camp Cady through May, 1868. (Casebier, 1973).

RETRACE to Cedar Canyon Road.

10.6 0.3 TURN RIGHT onto Cedar Canyon Road.

10.9 0.3 Pass Watson Wash tributary and road on the left to Fourth of July Canyon. Lithologies, imbrication, and cross-bed orientation of cemented gravels 1.5 miles north up this tributary suggest a source in the Colton Hills and Providence Mountains. The absence of Peach Springs Tuff clasts suggests that this deposit formed prior to 18.5 Ma. This deposit hints at the oldest Tertiary surface of the Mid Hills block when clasts from the south and west were traveling over a surface with a shallow, northeast dip.

11.0 0.1 SLOW: road turns to the right. Note the pediment formed on Rock Spring monzodiorite.

12.2 2.2 Road bends sharply right (south).

12.5 0.3 To the south-southwest at 1:00, notice basalt flows dated at 10 Ma (McCurry, 1985). These flows dip 8°E and may reflect the topography and slope of the surfaces on which they flowed, or may reflect subsequent regional

tilting.

13.1 0.6 Slow, road turns to the left.

15.2 2.1 Pass left turn to Caruthers Canyon.

16.4 1.2 Caution: road winds around north side of steeply-dipping, unmetamorphosed Paleozoic limestone promontory. This Joshua tree forest burned in late 1994 (see Minnich, this volume).

18.0 1.6 Road cuts through first exposures of well-developed pedogenic carbonates in western Lanfair Valley.

19.4 1.4 Stop, TURN RIGHT (south) towards Goffs at intersection of Cedar Canyon and Ivanpah Road. A convenient telephone just to the south was blown up 10 years ago; the currently available Lanfair Valley telephone is 0.2 mile north and is reached by turning left (north) Ivanpah Road and then right onto the telephone road. .

21.6 2.2 Proceed through intersection of road that runs westward to Ford's Lake.

22.5 0.9 The 1893 Nevada Southern (by 1895 the California Eastern, and by 1902 Santa Fe) (Myrick, 1963) railroad grade is visible on the east side of the road at 9:00. It has been cut through by local drainages (Figure 10).

23.3 0.8 Continue past a turn to the True Blue mine.

24.1 0.8 TURN RIGHT into parking area above Hackberry Wash.

24.2 0.1 Park. **Stop 13.** (see Reynolds, Hunt and Albright, this volume) Walk across Lanfair Road to view a flat pediment north of the Vontrigger Hills (Figure 11 and title page). To the west, at the east base of the Hackberry Mountains, are lacustrine sediments ponded by volcanic debris (Figure 12). The fossils contained in these sediments suggest a middle Hemingfordian Land Mammal Age of no younger than 17.8 Ma. These appear to interfinger with or be ponded in a thick section of debris from a rhyolite dome. Higher in the section are dacite tuffs and rhyolite flows. White sediments to the northwest may overlie and be younger than the Wild Horse Mesa Tuff, 17.8 Ma. The sediments we have seen in the Miocene section at Wild Horse Mesa and Hackberry are limited in thickness and areal extent and may be the result of local ponding due to Mt. St. Helens style damming of drainages (McCurry, this volume; Reynolds, Hunt and Albright, this volume). RETRACE to Ivanpah Road.

24.3 0.1 Stop, TURN LEFT and proceed north on Ivanpah Road.

26.7 2.4 Pass intersection with turnoff to Fords Dry Lake.

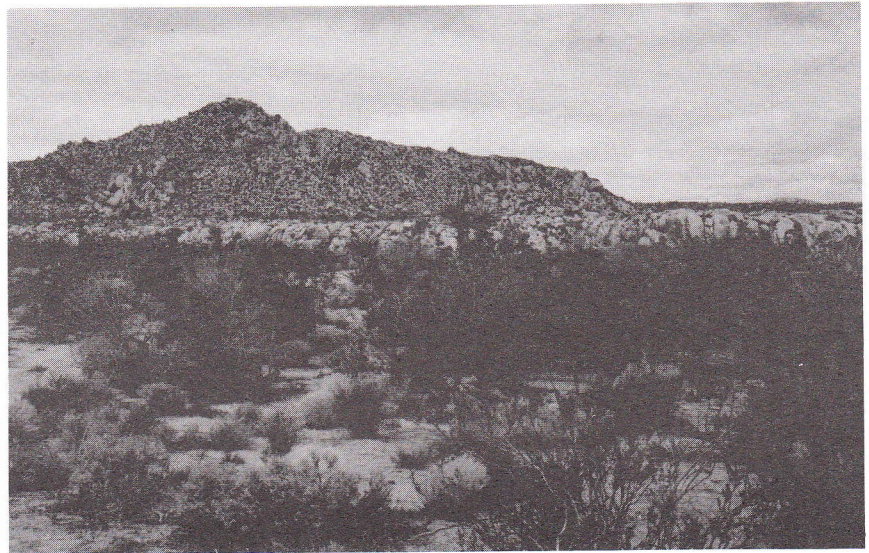


Figure 11. Erosion cuts headward into inselbert east of Hackberry. R.E. Reynolds photograph.

27.0 0.3 Pedogenic carbonates.

28.0 1.0 Road cuts through pedogenic carbonates.

29.0 1.0 TURN RIGHT (east) on Cedar Canyon Road (Pacific Telephone & Telegraph) toward Piute Range. PARK.

Stop 14. This is the townsite of Lanfair. The California Eastern railway supported several homesteading colonies, and Lanfair was the first. Established in late 1910 by Ernest C. Lanfair of Searchlight, it "boomed" in 1912, with a post office, a general store, a box car depot, and a school with as many as 29 students. Lanfair sunk two 550' deep wells, built a concrete reservoir, and laid eight miles of pipe from Lanfair Spring on Watson Wash. One mile away, Dunbar (1911-1914) was established as a colony for African Americans. Five miles north at Ledge, a general store opened in 1912 and a post office, Maruba, was established in



Figure 12. Light-colored, platy, fossiliferous lacustrine sediments at the Hackberry locality. R.E. Reynolds photograph.

1915. A combination of factors mitigated against the success of the colonies. Cattle trampled the crops, and the Rock Springs Cattle Company actively prevented access to water. The end of World War I in 1918 caused crop prices to drop. The railroad cut back service, and weather ranged from drought to downpour. As homesteaders gained title to their land, they sold out and gradually moved away. Only three families were left in Lanfair Valley by 1926 when the Maruba post office closed; the Lanfair post office closed early in 1927 (Casebier, 1987). CONTINUE on Cedar Canyon Road.

30.4 1.4 SLOW for dip.

31.2 0.8 At this intersection we are driving for a short distance on the trace of the Old Government Road.

31.5 0.3 SLOW for dip and curve in road.

32.7 1.2 BEAR SOUTHEAST (right) at fork in the road. The Old Government Road continues straight, toward Fort Piute.

33.1 0.4 Tin shed on left.

34.4 1.3 Continue along telephone road past diagonal junction.

34.8 0.4 Pass junction with road to Rattlesnake Mine. Watch out for dips.

35.0 0.2 WATCH OUT for severe dip in road. A large, subrounded boulder of Peach Springs Tuff is about 25 feet south of road. This probably was originally deposited as a clast in an upper Miocene boulder conglomerate unit. The size of this boulder suggests that, both when deposited in the conglomerate, and when reworked as part of this Quaternary deposit, its source was located relatively close by.

35.0 0.1 Road will be passing through deeply-weathered red clays on fans from the Vontrigger Hills to the south.

35.6 0.6 Caution: bad dip.

36.4 0.8 Continue through intersection.

37.5 1.1 Pass through intersection and proceed past tin telephone shed.

37.8 0.3 CAUTION: dips in road. Holocene drainage at this point runs south into Fenner Valley.

38.8 1.0 Weathered terraces contain clasts from the New York Mountains; clasts in the wash are derived from east of Barnwell.

39.4 0.6 TURN NORTH just prior to cattle guard.

39.9 0.5 Pass the Old Government Road which runs east up the slope of the Piute Range. On the east side of the range, wagon wheel ruts are still visible where they cut the scoria. Fort Piute is the only fort in this chain that had the benefit of being designed by a military engineer (see Haenszel, this volume). Look northeast at 2:00 to white pedogenic carbonates exposed on inclined slope along the west face of the Piute Range in dark volcanic debris of basalt to andesite

composition. These soils must be younger than the roughly 8 Ma basalt.

40.2 0.3 Proceed north past east side of corral.

40.5 0.3 Caution; road bears left.

40.8 0.3 Caution, road bears left again. From this turnout, a trail can be followed into Piute Gorge and on to Fort Piute.

41.3 0.5 TURN RIGHT to promontory.

41.4 0.1 PARK at Lovers Leap.

Stop 15. We are at the west side of Piute Gorge, rim elevation 3400'. The bands of tan and red-brown below us are sediments deposited in a playa overlying dark flows of the Piute Range and eroded into "badlands" topography. The playa sediments formed in late Tertiary (Pliocene) and early Quaternary time, due to ponding of Lanfair Valley drainages by the buttress of the Piute Range; the very linear range front indicates a fault boundary.

The stream in the canyon bottom flows to Piute Valley through dark-colored lava flows and volcanic debris deposits of the Piute Range in Piute Gorge. The antecedent stream probably represents eastward-trending drainage of the area in late Miocene time. Piute Gorge became entrenched into the faulted rocks during uplift of the Piute Range, possibly beginning in the latest Miocene or early Pliocene time. Although the canyon trends generally east, it takes several sharp north-south bends, following faults with north- to northwest trends. It was this uplift event that ponded the streams of Lanfair Valley and probably initiated reorientation of area drainages, which now flow southward.

Below the rim on which we are standing, you can see pedogenic carbonate (Figures 13, 14) representing a long-term, stable surface. (Katzenstein and others, this volume). The youngest stable surfaces present are the late Pleistocene desert pavement and associated playa sediments near the corral and the last cattleguard. Subsurface drainage today runs eastward and some appears as Piute Spring which flows 50 gallons/minute (Hewett, 1956).



Figure 13. Pedogenic carbonates in the sedimentary section at Grandview Gorge. R.E. Reynolds photograph.



Figure 14. Detail of cross-bedded sediments and gravels cemented by carbonates at Grandview Gorge. R. E. Reynolds photograph.

Looking clockwise over Lanfair Valley, Table Mountain is due west; Cedar Canyon is north of Table Mountain to the west-northwest; the New York Mountains are north-northwest; dissected fans at the east end of the New York Mountains are to the northwest; Castle Peaks are north; and the Piute Range is to the northeast. Nielson and Nakata (1993) divided the volcanic rocks of the Piute Range into a



Figure 15. Petroglyph at Telephone Pass. R.E. Reynolds photo.

lower Miocene (19.8-12.2 Ma) and upper Miocene (10.7-8.0 Ma) sections. The upper section is similar in age to basalts we passed southwest of Rock Springs. RETRACE route to cattle guard.

42.4 2.0 Cattle guard. TURN LEFT (east) and proceed to top of Telephone Pass.

42.8 0.4 Pull off road to the left and PARK. **Stop 16.** This pass was used historically and prehistorically (Figures 15, 16) (Haenszel, this volume; Turner, this volume). We are at a drainage divide between Lanfair Valley and Piute Valley to the east. South of our present location the surface drainage of Lanfair Valley splits into three directions near the town of Goffs: one part of the drainage flows southwestward into Fenner Valley, one drains southeastward, around the south end of Signal Hill and Homer Mountain and into Sacramento Wash, where it joins the drainage of Piute Valley and flows to the Colorado River. A third part drains due south into Ward Valley; this last part of the Lanfair Valley drainage is observed on air photos, but is not depicted on some topographic maps. This complex drainage pattern may have been created by crustal deformation in the Quaternary.

We are standing at elevation 3600'. To the west, in the Vontrigger Hills at approximately the same elevation, Proterozoic metamorphic rocks underlie the early Tertiary erosional surface. Near us are 1400 Ma granitic rocks. Because the Miocene volcanics are thin over these rocks, we may be near the same early Tertiary erosional surface. The view east is toward the Proterozoic metamorphic core complex. The Dead (Spirit) Mountains to the north are 16 Ma Miocene granitic rocks. In Piute Wash to the east, pre-Paleozoic metamorphic rocks and Mesozoic granitic rocks crop out at 2000', forming pediment bases around isolated



Figure 16. Petroglyph at Telephone Pass. R. E. Reynolds photo.

promontories (called inselbergs, or "island mountains") that once were high elevations on the early Tertiary erosional surface. The discrepancy in elevation between exposures of the erosional surface in Piute and Lanfair Valleys is due to progressive offset by north-trending Miocene to Quaternary(?) faults; aeromagnetic measurements indicate that the major offset is along the west side of the Piute Range. The latest fault movements elevated the Piute lavas enough to dam drainages in Lanfair Valley (see Stop 15, above) and carve out two east-trending canyons (Fort Piute at the south end, and "Homestead Canyon" at the north end of the range).

Some of the Miocene faults must have developed during accumulation of Tertiary rocks 1500' to 2000' thick that form the Piute Range; exposures of these thick deposits are narrow in an east to west direction (3 mi.), but extend for about 16 miles to the north of here. In this area the Tertiary erosional surface, defined by the base of the lowest Tertiary rocks, gradually changes in elevation from about 2800' east of Fort Piute, to about 3600' on Signal Hill (to the south of us), where lava flows are less than 400' thick. This change in elevation on the erosional surface defines the southern limit of the thick Piute Range lavas, which probably represent accumulation in a fault-bounded trough between about 19 and 8 million years ago. In a similar time frame (18.5 to 13 million years ago), andesitic to light-colored rhyolitic rocks and minor basalt flows were being erupted nearby, in the Castle Mountains. Limited lateral overlap of the Piute Range and Castle Mountains volcanic rocks suggest that they collected side-by-side in fault-bounded troughs.

We do not see the 18-8 Ma volcanic section of rocks to the east that we have traveled through for two days. This area to the east is a highly extended terrane of middle Miocene age, where substantial uplift of once-deep rocks and accompanying erosion have resulted in the stripping of most shallow rocks. Most visible rocks are Proterozoic metamorphic rocks and Mesozoic and Tertiary granitic rocks. Some of the latter have been eroded to form pediments. Our view east includes Homer Mountain, which contains the closest structural evidence of extensional tectonism (Duebendorfer and others, 1993). We are at the eastern margin of a stable area with minor extensional faulting. The Colorado River Extensional Corridor to our east runs north to the Black Mountain Accommodation Zone which is truncated by the Las Vegas Valley - Lake Mead Shear Zone (Duebendorfer and others, 1993). RETRACE route west to cattle guard.

43.3 0.5 Cattle guard.

45.1 1.8 Telephone shed. Prepare to turn right (north).

45.4 0.3 TURN RIGHT at intersection.

45.6 0.2 PARK at tank. **Stop 17.** Walk east along the Old Government Road to inspect wash. The wash bank contains bedded and cross-bedded sands with clasts from the Vontrigger Hills to the southwest. Notice platy carbonates and underlying carbonate-cemented burrows and root casts. These sediments may be comparable to the dissected sediments that overlie the dense carbonates at Grandview Gorge. Return to vehicles and RETRACE to Telephone Road.

45.8 0.2 TURN RIGHT (west).

47.0 1.2 CAUTION; dips.

47.3 0.3 CAUTION, dips.

47.9 0.6 Watch the road; note boulders of Peach Springs Tuff.

48.9 0.9 Telephone shed.

49.3 0.4 Proceed through intersection.

50.5 1.2 CAUTION: bend and dip in road.

53.0 2.5 Stop, TURN RIGHT onto Ivanpah Road.

53.2 0.2 Continue past right turn to **telephone**.

54.5 1.3 Volcanic rocks of the Grotto Hills are to the west. (The hills are an official misspelling of "Guirado," for local homesteader Mary Ann Guiardo who filed this property in 1911 and patented it in 1921. The land is still owned by the family (Casebier, 1987)). These Tertiary rocks (probably dacite or andesite) probably formed in the Miocene, like the volcanic rocks that ring Lanfair Valley. However, structures in the Grotto Hills are difficult to interpret in terms of volcanic eruptions and the rocks probably represent near-surface intrusive domes. Outcrops in the Lanfair Buttes (AKA Eagle Mountain) to the east (see Turner, this volume) are intrusive rocks of similar composition, overlain by airfall tuff and unambiguous lava flows.

The historic Nevada Southern railroad grade (to west) is now used to pond water for cattle.

56.0 1.5 Cattle guard. Railroad grade to the west continues alongside of the road.

58.3 2.3 TURN LEFT (west) onto New York Mountain Road at the OX cattle company ranch headquarters.

60.9 2.6 Pedogenic carbonates visible in road.

61.4 0.5 Pass through intersection.

61.6 0.2 Pedogenic carbonates in road.

63.0 1.4 TURN RIGHT into Caruthers Canyon (Figure 17).

63.9 0.7 PARK at tank. **Stop 18.** We are parked near the trace of the Cedar Canyon Fault. The trace runs to the south-southwest cutting through Miocene Wild Horse Mesa Tuff and dropping it down to the south with respect to the New York Mountains. Near here, pre-late Pleistocene gravel is cut by the fault, but late Pleistocene deposits are not. Blocks of resistant Peach Springs Tuff are perched at elevations of 5800' in Fourth of July Canyon to the east suggest that the New York Mountains were once lower elevation and define the old 18.5 Ma Miocene surface. To the northeast, the trace of the Cedar Canyon Fault is enigmatic. It may be hidden beneath very large slide blocks of unmetamorphosed Paleozoic limestone which appear to be intact. The source of this limestone is puzzling, since most of Paleozoic limestone in the eastern New York Mountains was metamorphosed with the intrusion of the Mid Hills adamellite.

The bulk of the New York Mountains consists of the Cretaceous Mid Hills adamellite (Beckerman and others, 1982) (Figure 18). The pluton is 95-97 Ma (Miller and others, 1994). Younger K-Ar dates come from a 400 m wide, north-northwest striking mylonite zone in the western part of the range "which may correlate to the Morningstar Thrust in the southern Ivanpah Mountains" (Beckerman and others, 1982). Work by G.A. Davis (Burchfiel and Davis, 1977)

indicates the Morningstar Thrust is Jurassic. However, recent study by D.M. Miller indicates the New York Mountains mylonite is late Cretaceous. Furthermore, the Morningstar Thrust has reverse-sense motion and the mylonite zone shows normal sense motion. Recent studies suggest a coincidence for the Teutonia Batholith, extrusion of the Delfont Volcanics, and continued late Cretaceous thrusting of the foreland fold-and-thrust belt (Fleck and others, 1994). At the mouth of Caruthers Canyon and along the east side of the canyon are metamorphosed Paleozoic and Mesozoic strata typical of the eastern New York Mountains (Brown, this volume; Burchfiel and Davis, 1977). A post plutonism stress regime is suggested by the roughly east-northeast strike of hydrothermal veins, porphyry dikes (Hewett, 1956) and pegmatite dikes (Reynolds and Housley, this volume). Similarly oriented dikes cut much of the Teutonia batholith, although local variations are common. The dikes appear to only be present in the upper parts of the batholith, near its roof.

New York Peak, elevation 7535', hosts fir, 3-needle piñon, and oak (Presch, this volume). At the tank you can see Nevin's barberry, 2 species of oak, scrub oak, *Yucca bacatta*, *Y. brevifolia*, piñon, juniper, wild almond, and mistletoe (Presch, this volume).

The New York Mountains have a long history of mining (see Reynolds and Reynolds, this volume, and Vredenburg, this volume).



Figure 17. Mr. and Mrs. Caruthers ("Carruthers"). J. Riley Bemby collection.

RETRACE route to New York Mt. Road

- 64.3 0.6 TURN LEFT (east) onto New York Mountain Road.
- 65.0 0.7 Take the right fork at the junction.
- 65.5 0.5 Cattle guard. At 9:00 cemented fans may mark the northeast trace of Cedar Canyon Fault.
- 65.7 0.2 Pedogenic carbonates.
- 66.8 1.1 Proceed across junction.
- 69.9 3.1 Stop at intersection with Ivanpah Road. We are again at the OX cattle company headquarters, the site of Ledge siding (later, Maruba) on the Nevada Southern/Eastern California railroad (Myrick, 1963; Sharp, 1984). TURN LEFT (north).
- 72.4 2.5 To the right, pedogenic carbonates cement clasts of Paleozoic limestone in the bank of the wash crossing road.
- 72.8 0.4 To the west, carbonates are visible in the wash bank.
- 73.8 1.0 Proceed past historic road southwest to Caruthers Canyon and northwest to Sagamore Canyon at historic Purdy siding (Myrick, 1963). From here north, we are driving on the surface of carbonate-cemented fanglomerates derived from the New York Mountains.
- 75.8 2.0 The pyrophyllite mine is visible at 11:00. Southeast of this mine are metavolcanic rocks possibly equivalent to the Delfont Volcanics in the Mescal Range (Burchfiel and Davis, 1977). The Live Oak granodiorite, part of the Teutonia batholith, is located in this area

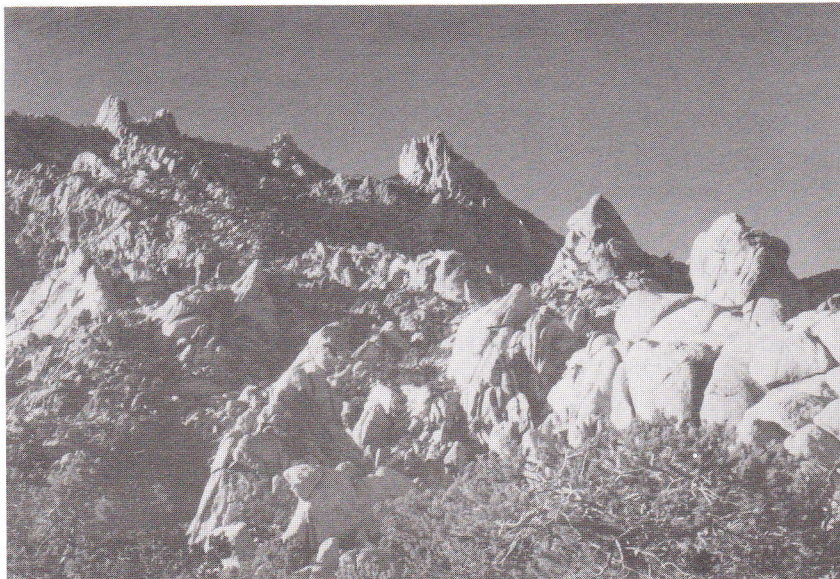


Figure 18. Peaks in the New York Mountains. R.E. Reynolds photograph.

(Beckerman and others, 1982).

76.0 0.2 Cross cattle guard.

77.1 1.1 **Stop 19.** On the east side of the road (right) is an exposure of the late Miocene or Pliocene gravel which overlies volcanic rocks in the Castle Peaks, Castle Mountains and Piute Range, and filled stream channels within lava flows of the Piute Range. Here it is coarsely-bedded pebble sandstone, containing trains of rounded-to-subangular cobbles of granite (Mesozoic and older), Paleozoic limestone, calc-silicate rocks, and various volcanic rock types. The unit has a sandy matrix containing a high proportion of crystals (quartz, plagioclase, biotite), and generally displays greater rounding of the clasts compared to clay-rich matrix and more angular clasts of overlying Quaternary deposits. Most exposures are poorer in sandstone than this one, and some deposits contain clasts that are armchair-size boulders (like the Peach Springs Tuff boulder, described earlier). Upsection in this outcrop, Paleozoic limestone and marble clasts increase in abundance. These gravel fans head toward a point to the northwest where no source can be seen. The sequence in the hill (first granitic clasts, then carbonate clasts) may reflect paleotopography when the Live Oak granodiorite was high. Subsequent to stream capture would have shed Paleozoic carbonate debris eastward into Lanfair Valley at this point. Climate may have played a part in making the Live Oak pluton now a topographic low when it was previously a high.

77.3 0.2 Pass left turn to Sagamore Canyon and Keystone Canyon. CAUTION: drive slowly; curves ahead. To the west, granitic basement is overlain by a distinctive sequence of Paleozoic and early Mesozoic rocks in the New York Mountains. There is a very similar sequence to the northwest across Ivanpah Valley: Paleozoic and early Mesozoic rocks in contact with the Teutonia Batholith. We are on the east side of the Slaughterhouse Fault in terrain which is underlain by Early Proterozoic amphibolite generally similar to the metamorphic rocks east of the Clark Mountain Fault in the Ivanpah Mountains and the Mescal Range to the north. In the New York Mountains, the metamorphic, granitic, and Paleozoic rocks were exposed and stripped by erosion before the middle Miocene. The erosional surface was then covered with sediments and volcanic rocks. These deposits contain the 18.5 Ma Peach Springs Tuff, which is the only time-marker to constrain the underlying erosional surface. We are leaving the area of northeast-striking faults and entering an area of northwest-striking faults that offset Proterozoic and late Tertiary volcanic rocks. Many of the northwest-striking faults have small separation and they are hard to date. Most postdate late Miocene gravel and predate Late Pleistocene alluvium.

78.5 1.2 Continue past the Barnwell/Hart Mine road.

78.9 0.4 Road cut. CAUTION: proceed slowly downhill. We are driving along the old California Eastern railroad grade, built from Manvel (Barnwell) to Leastalk (today, Ivanpah) in 1901-1902. The road bed was graded by up to 150 Mojaves, mostly by hand with occasional low grade glycerine powder (Myrick, 1963).

79.5 0.6 The road diverges from the railroad grade, which runs on the east side of the canyon.

79.7 0.2 CAUTION: watch for oncoming traffic and

prepare for left turn.

79.8 0.1 TURN LEFT into wash. PROCEED south-southwest along historic road to Barnwell that lies westerly of the railroad grade.

80.2 0.4 Historic road is crossed by wash. PARK and walk westerly to **Stop 20.** Peach Springs Tuff sits on and protects bright red, deeply-weathered middle Tertiary soils developed on Proterozoic amphibolites. Notice the fiamma (flattened pumice) and breccia in the PST, and look for adularic feldspar. Look in the wash for cobbles of Proterozoic garnet-rich gneiss that is characteristic of the metamorphic terrane east of the Slaughterhouse Fault and also east of the Clark Mountain Fault in the Mescal Range.

This outcrop of the Peach Springs Tuff underlies locally-erupted volcanic rocks of the Castle Peaks. The basal contact of the tuff is not well exposed here, but northeast of Ivanpah Road the Peach Springs Tuff overlies an arkosic (derived from granite) conglomeratic sandstone unit that filled stream channels in the Tertiary erosional surface. The Peach Springs Tuff is alkali rhyolite in composition, and is distinctive because of abundant feldspar crystals (adularic sanidine) that display a blue flash as the specimen is rotated (the blue flash can be seen by thenaked eye in good sunlight conditions), and yellow-brown lozenges of a titanium-bearing mineral called spene; also present may be plagioclase, hornblende, biotite, allanite, pyroxene, and quartz.

In addition to minerals, the tuff contains rock fragments, some picked up from surface debris as it flowed. Such contamination by sand and fragments derived from older rocks has made the unit difficult to date using bulk samples; the accepted age of 18.5 million years was determined on minerals from lumps of pumice (Nielson and others, 1990). Elongate light-colored streaks are flattened pumice lumps; the coarse size of these suggest that the eruptive source is relatively near; measurements of various kinds indicate a source either in Lanfair Valley (Buesch, 1993) or near the conjunction of Nevada, California, and Arizona (Hillhouse and Wells, 1991). On the rock, note shiny, planar areas about parallel to the creek — these are small fault surfaces. The outcrop above the tuff (uphill) is the light-colored, coarse lower breccia unit of the Castle Peaks. Myriad faults have sliced these two units and repeated them numerous times between this outcrop and Barnwell.

RETRACE route to Ivanpah Road.

80.6 0.4 TURN RIGHT (south) onto Ivanpah Road and proceed to Barnwell.

81.7 1.1 On the right (north) side of the road, notice blocks of Peach Springs Tuff with drill holes. The resistant tuff was quarried for railroad berms and bridges.

81.9 0.2 TURN LEFT (south) at Barnwell onto the Hart Mine road. The historic 7IL Ranch headquarters building still stands.

82.3 0.4 Pass windmill on right side of road.

82.6 0.3 PARK at second windmill with water tank.

Stop 21. Hike south to top of hill.

From this point we can see the skyline formed by locally-erupted volcanic rocks of Miocene age that dominate outcrops of the Castle Peaks (north), Castle Mountains (east,

site of the big gold mine), and Piute Range (south). Compositions of volcanic materials in each of the mountain ranges are mostly types of andesite varying to rhyolite; minor basalt flows are found in the Castle Mountains and piute Range. The locally-erupted volcanic rocks of each range overlie the regionally-widespread Peach Springs Tuff (see Stop 20). The ages of the youngest dated units indicate that volcanism ended about 15 to 16 million years ago in the Castle Peaks, but continued until 13 million years ago in the Castle Mountains, persisting to 8 million years ago in the Piute Range.

The Castle Peaks volcanic rocks are a series of very coarse to moderately coarse volcanic breccia that erupted mostly from narrow sheet-like feeders (called "dikes"). The lowest and coarsest breccia is light-colored dacite, which probably fared explosively during near-surface intrusion of sticky magma domes (like Mount St. Helens). Less coarse dark andesitic flow breccia deposits overlie the lower unit and breccia deposits that represent mixtures of these two compositions are found locally. Interbedded with the dark flow breccia are ash flow tuff units. One of them (informally called the Barnwell Tuff) is exposed near the corral; this may be the equivalent of the (15.8? million year old) Tuff of Wild Horse Mesa (McCurry, 1988). We will walk up a slope underlain by the dark andesite breccia unit, which is overlain in turn by Tertiary gravel (not well exposed here) and then by Quaternary deposits.

From the ridge we can see some of the spiky buttes of the Castle Peaks, and the steep-sided intrusive rhyolite domes that form the castellated backbone of the Castle Mountains. The domes are mantled by light-colored tuff and breccia deposits that represent violent eruptions of pulverized lava from central volcanic vents (or dikes); the domes likely mark the sites of central vents. The tuff and breccia units are thickest in the south part of the Castle Mountains and gold is being mined by the Viceroy Company (of Canada) from dense swarms of veinlets within thick tuff and breccia that were altered and mineralized by fluids from the intrusions. The thick rhyolitic rocks of the Castle Mountains overlie dark lava flows and an ash flow tuff that are similar in composition and age to the breccia units and Barnwell Tuff in the Castle Peaks.

We have already viewed the long, flat ridge of the Piute Range from southern perspectives, and discussed the volcanic deposits. Although the compositions and ages of the locally-erupted volcanic rocks are very similar for the Castle Peaks, Castle Mountains, and Piute Range, the actual deposits are surprisingly different in character, as suggested by the very different appearance of each range from this vantage point.

Tertiary gravel, containing rounded limestone derived from the New York Mountains, is found in the central part of the Castle Mountains east of the high ridge of domes.

Transportation of those clasts requires a drainage that flowed generally north and east in late Miocene time, before uplift of the Castle Mountains. Reorientation of the drainage to the present southeast trend probably was accomplished by post-Miocene faulting. Although the section is thin, the lithologies appear to increase upsection and to the southeast as follows: a) metamorphic and Tertiary volcanics; b) granitics and Tertiary volcanics; c) Paleozoic limestone.

Return to vehicles and PROCEED SOUTHEAST on the Hart Mine road.

83.6 1.0 Road cut through carbonate-cemented conglomerate.

84.1 0.5 Continue past intersection on right.

84.6 0.5 PARK. **Stop 22.** Pedogenic carbonates are on the left side of the road. The Castle Peaks are northwest, the Castle Mountains due east, the Hart/Viceroy gold mine east-southeast, and the Piute Range and Piute Gorge south-southeast. The mining community of Hart emerged with gold discoveries in late 1907 by James Hart with Bert and Clark Hitt. The mining camp grew rapidly in early 1908 and the townsite was surveyed. It grew to include stores, hotels, real estate offices, a bakery, two lumber yards, saloons, a book store, and a candy store. Fire destroyed much of the town in late 1910, but gold production continued until 1915. Between 1930 and 1974, clay was the major commodity shipped from Hart. Today, the Hart district has been reactivated with the Viceroy gold mine (Hensher, 1991; Snorf, 1991; Vredenburg and others, 1981).

Return to vehicles and RETRACE to Barnwell Road.

87.3 2.7 Barnwell. TURN RIGHT at intersection with Ivanpah Road.

88.6 1.3 Pass historic road to Barnwell. OK

89.5 0.9 Historic mill and tailings at 3:00.

90.4 0.9 Stop and PARK alongside road near turnoff to Vanderbilt Gold Mine. **Stop 23.**

Gold was discovered in 1891 and by 1893 Vanderbilt boasted a dozen tents and a population of 200. Unfortunately, the promise of the mines was not met, and of the 200 only about 20 were regularly employed. That the promised riches did not materialize contributed in part to the failure of the Nevada Southern to progress from Manvel to Vanderbilt (Myrick, 1963).

A pyrophyllite quarry in marble is to the south; peaks of the New York Mountains (elevation 7353') are to the southwest; the Vanderbilt Gold Mine is to the southeast. CONTINUE on Ivanpah Road. We are in a zone east of the Slaughterhouse Fault where there are slivers of Upper Paleozoic Bird Spring Limestone and Mesozoic carbonates and metavolcanics mixed with slivers of Proterozoic gneiss. This suggests that in this area the Slaughterhouse fault zone is very complex.

90.8 0.4 Start of pavement.

91.5 0.7 PARK alongside road before it bends to the west (left). **Stop 24.** Look approximately to the northeast, just to the right of the Castle Peaks. With good light on the crest of the range and perhaps a good imagination, you can discern a shallow, U-shaped patch of light material set into darker rock. The light colored material is Miocene gravel that fills a paleovalley cut into earlier volcanic rocks younger than 17 Ma (Miller and others, 1985; Miller and Wooden, 1993). The gravel contains abundant clasts of metamorphosed rocks and granite derived from the southern New York Mountains (the high range visible to the southeast), and a few clasts derived from the Proterozoic rocks of Willow Wash (in the foreground of the paleovalley). Lack of clasts derived from gneiss of the eastern New York Mountains and the Castle Mountains indicates a generally



Figure 19. View from the New York Mountains north to Ivanpah and Roach lakes; the Spring Mountains are distant. R.E. Reynolds photograph.

north and east direction of transport. The paleovalley trends southeast and was eroded at least 180 m into volcanic and metamorphic rocks; flow in it must have been toward the southeast. The channel's course leading to the point where we see it in the ridge crest must have been directly above us several hundred meters (Miller, this volume).

A combination of faulting and erosion has dramatically modified the landscape since that paleovalley was active. These changes in landscape took place well before Pleistocene alluvium blanketed slopes leading west from the Castle Peaks, since the slopes are relatively gentle. In contrast, the slopes on the west side of the McCullough Range in the distance to the north, up Ivanpah Valley, are precipitous. Miller and Jachens (1995) propose that Basin-and-Range faulting is active in the area of the McCullough Range but faulting has not taken place for at least a few million years in the area of the Castle Peaks.

Look west to the south slope of Cima Dome, the Ivanpah Range, Mescal Range, and Clark Mountain, to the northwest. Devil Peak is due north, and there is likely snow on Charleston Peak at elevation 11,918'. North and northeast are Spring Mountain, Lucy Gray, and the McCulloh Range. Looking to the northwest we can see the approximate position of the Clark Mountain Fault near the Morningstar Mine. Preliminary magnetic and gravity studies (Swanson and others, 1980) suggest that the Clark Mountain Fault of Hewett, the Kokoweef Fault of Burchfiel and Davis, and associated structures were once continuous with the Slaughterhouse Fault and structures in the northern New York Mountains immediately to our west (Miller and Wooden, 1992). The magnetic contour maps suggest that the Clark Mountain Fault has been left laterally offset 8-9 miles by a fault in central Ivanpah Valley (Figure 19). Miller and Jachens (1995) consider the fault in Ivanpah Valley to be exposed near Nipton, where a broad shear zone offsets Proterozoic rocks about 15 km in a left-lateral sense. The three major northeast-striking faults we have crossed during the last two days might be characterized as being left lateral

and/or normal, down to the south. Miller and Jachens (1995) describe a possible connection of this system with the Eastern California Shear Zone west of Cima Dome. Their model proposes that the Granite Mountains were recently thrust to the north, depressing the Devil's Playground area.

93.7 2.2 Ivanpah, formerly South Ivanpah and before that Leastalk. The California Eastern railroad was constructed from Barnwell to this siding on the San Pedro, Los Angeles, and Salt Lake (now Union Pacific) railroad in 1901-2 to provide a route to ship ores from the Valley Wells smelter (Myrick, 1963).

93.0 0.3 Caution: avoid high speed trains when crossing railroad tracks.

99.4 6.4 Junction of Morningstar Mine Road. If you have enough fuel and want to return via I-40 or Twentynine Palms, turn left here to Cima. Otherwise, proceed along Ivanpah Road to Interstate 15.

102.6 3.2 Stop, TURN LEFT on Nipton Road.

106.0 3.4 Stop on the north side of the Interstate 15 overpass. Go west to Baker; east to Stateline, Nevada.

End of Day Two

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The Hope - New Method Mine, San Bernardino County, California

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The New Method Mine (originally the Hope Mine) is located on the east edge of the Bristol Mountains about six miles northeast of Amboy, San Bernardino County, California. The first published report on this mine was by Chesterman and Bowen (1958). The report listed fluoborite as occurring in association with forsterite, fluorite, clinohumite, diopside, tremolite, calcite, dolomite, brucite, magnetite, uranophane, beta-uranite, and an unidentified uranium-yttrium mineral.

The uranium minerals were almost certainly misidentified by Chesterman and Bowen. Dr. William Wise (Department of Earth Sciences, University of California, Santa Barbara, writ. comm. to author Feb. 28, 1995) completed a study of



Figure 1. Boltwoodite and fluorite, 30x magnification. John Jenkins photograph.

Table 1. Mineral Assemblage, Hope-New Method Mine

Fluoborite	$Mg_3(BO_3)F(OH)_3$	occurs in seams and is similar in appearance to satin spar. Also as prismatic crystals in a matrix of calcite.
Boltwoodite	$(H_2O)K(UO_2)SiO_4$	yellow, flattened blades and hairlike crystals, sometimes with a tuft on the termination
Fluorite	CaF_2	Usually such a deep purple color as to appear black; others lighter in color ranging from pale lilac, some with darker phantoms. It is not unusual to find fluorite crystals impaled on boltwoodite.
Sepiolite	$Mg_4Si_6O_{15}(OH)_2 \cdot 6H_2O$	white hairlike crystals, often more or less matted. Sometimes as tufts on the termination of boltwoodite crystals.
Calcite	$CaCO_3$	occasionally has boltwoodite crystals attached
Quartz	SiO_2	colorless double-terminated and opaque white. Colorless crystals have been found attached to sepiolite, and white crystals impaled on boltwoodite crystals. The white crystals fluoresce bright green.
Sklodowskite	$Mg(UO_2)_2Si_2O_6(OH)_2 \cdot 5H_2O$	found as an overgrowth on boltwoodite (Wise).
Uranophane	$Ca(UO_2)_2(SiO_3OH)_2 \cdot 5H_2O$	Although most reports list uranophane, this is probably a misidentification of boltwoodite (Wise).



Figure 2 (left).
Boltwoodite, 40 x
magnification. John
Jenkins photograph.

Figure 3 (below).
Quartz, 30 x
magnification. John
Jenkins
photograph..

the uranium minerals from the New Method Mine in 1982, using several different methods to verify the results. All samples except one were boltwoodite (a species which had not been described at the time Chesterman and Bowen published their 1958 report). Crystals containing phantoms were identified by Wise as having boltwoodite cores with overgrowths of sklodowskite.

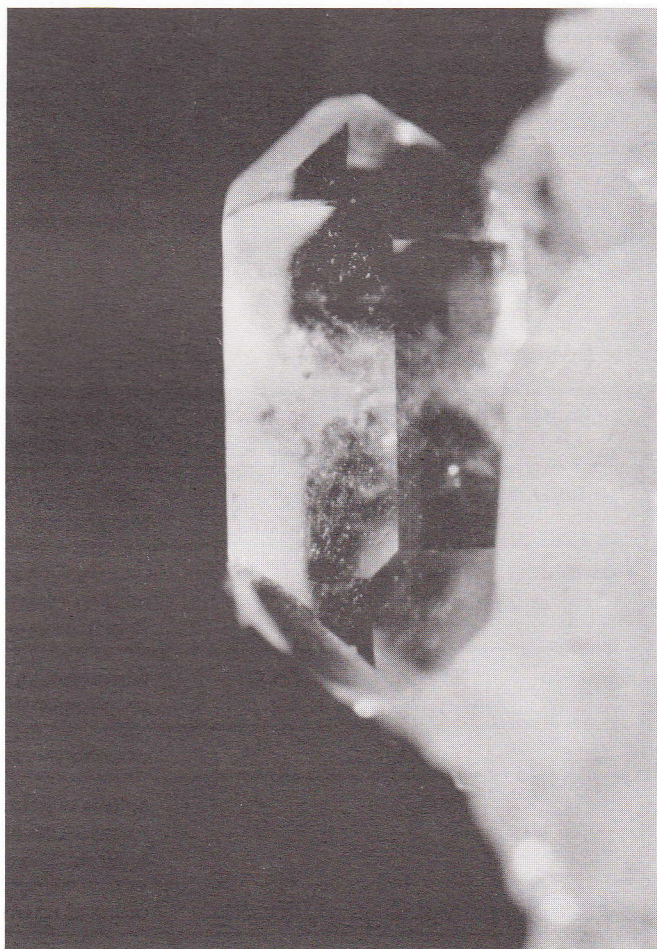
The mine workings include an inclined shaft, a vertical shaft, and a horizontal adit that reached the fault zone about 100 feet from the portal. Fluorborite, boltwoodite, fluorite, calcite, and quartz have been obtained from the inclined shaft. Sepiolite, fluorite, and calcite have been obtained from the horizontal adit. There is no available record of minerals from the vertical shaft.

The workings explored a mineralized sheared contact zone between quartz syenite and Paleozoic dolomite. Uranium, thorium, rare earths, and a number of trace metals were found at this and other prospects in the area (Otton, Glanzman, and Brenner, 1980).

No records of ore shipments from the Hope-New Method mine are known. Some fluorborite specimens may have been shipped to Wards Scientific Supply in New York for sale to mineral collectors.

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Early Miocene Dome Emplacement, Diking, and Limited Tectonism in the Northern Marble Mountains, Eastern Mojave Desert, California

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Abstract

The northern Marble Mountains of the eastern Mojave Desert, California, contain a thick sequence of early Miocene volcanic rocks which was erupted while extension was active in areas to the east and west. From bottom to top, the Miocene section comprises 1) a complex sequence of dacite to rhyolite domes, lava flows, and pyroclastic rocks; 2) basalt flows and scoria; and 3) the Peach Springs Tuff. K-Ar dating of a dacite from near the base of the sequence, coupled with the presence of the Peach Springs Tuff at the top indicates that the entire sequence was erupted between approximately 20.2 and 18.5 Ma ago. During that interval, extreme extension occurred in the Colorado River trough to the east and in the Waterman Hills area to the west. The Miocene sequence in the Marble Mountains is only weakly deformed, and most deformation, including hundreds of meters of dip-slip displacement along a curved fault, can be ascribed to doming of the sequence by hypabyssal dacite plugs. Thus, the Marble Mountains and surrounding ranges formed an unextended corridor between the central Mojave and Colorado River trough extended areas.

Introduction

The Marble Mountains lie in the eastern Mojave Desert, about 150 km east of Barstow, California (Figs. 1, 2). Their tectonic setting is important for two reasons. First, they occupy a relatively unextended region between two areas of extreme Miocene extension. To the east, steeply-tilted Tertiary strata in the Whipple Mountains and surrounding ranges record tens of km of early Miocene extension along the Whipple Mountains detachment and other low-angle normal faults (Davis et al., 1987; Howard and John, 1987); to the west, the Waterman Hills detachment was active at the same time (Glazner et al., 1989; Walker et al., 1990). Several authors have proposed kinematic models which link these highly extended regions (Bartley and Glazner, 1989; Davis and Lister, 1988; Glazner et al., 1989; Hileman et al., 1990). Knowledge of the structural history of the Marble Mountains region is thus critical to testing these models.

Second, the Marble Mountains lie along the eastern boundary of the Mojave block (Fig. 1; the "Mojave-Sonoran boundary" of Fuis, 1981), an important but poorly understood structure which separates an eastern terrane of linear, north-trending ranges with abundant Paleozoic and Proterozoic rocks from a western region which lacks these features (Fuis, 1981; Glazner, 1990). The Marble Mountains contain the westernmost widespread exposures of Paleozoic strata in this part of the Mojave Desert. Studies of the Marble Mountains will yield information about the nature of the Mojave-Sonoran boundary.

In this paper we present new structural and stratigraphic data for the northern half of the Marble Mountains. These data result from 1:20,000 mapping on color air photos, combined with structural analysis of selected faults. Our principal conclusion is that volcanism in the Marble Mountains occurred while extension was active in the Colorado River trough and the central Mojave Desert, but the Marble Mountains record little, if any, extension. Thus,

the Marble Mountains region defines a stable block between two highly extended regions.

Previous Work

Hazzard (1931) and Kilian (1964) mapped the entire range at 1:52,000 and 1:31,250, respectively. They dealt with the Tertiary section in reconnaissance. Much attention has been given to the relatively undeformed and unmetamorphosed Paleozoic strata in the southern part of the range (e.g., Hazzard and Mason, 1936; Stone et al., 1983). Bassett and Kupfer (1964) included the Marble Mountains in their summary of the geology of the southeastern Mojave Desert,

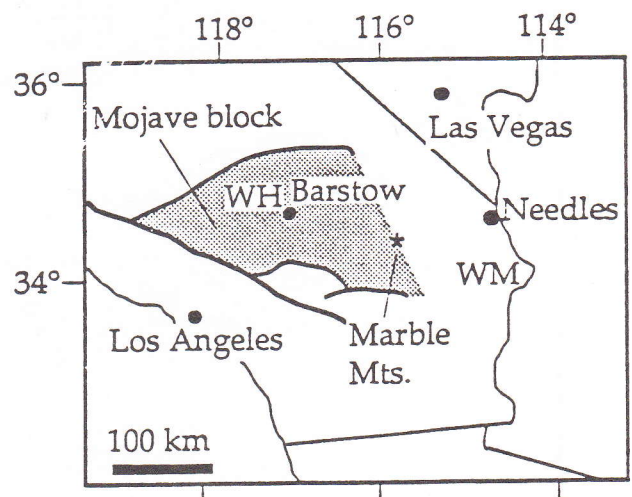


Figure 1. Location of the Marble Mountains, which lie along the eastern edge of the Mojave Block (shaded). WH and WM show the locations of the Waterman Hills and Whipple Mountains areas, both of which were actively being extended when the volcanic section in the Marble Mountains was being erupted.

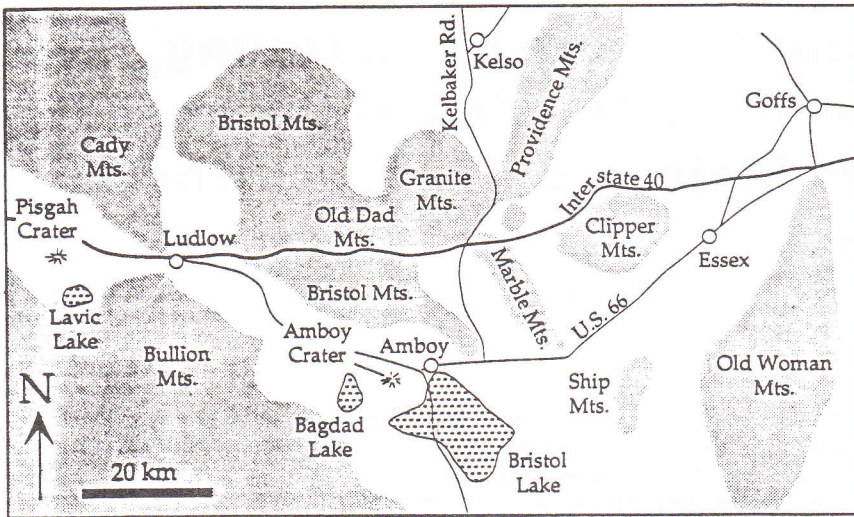


Figure 2. Geology around the Marble Mountains.

as did Miller et al. (1982) in their summary of the geology of the region surrounding Bristol Lake.

Stratigraphy

Map Units

The stratigraphic column in the northern Marble Mountains is conveniently divided into six units, five of which are depicted on Figures 3 and 4. From oldest to youngest, these are: pre-Tertiary metasedimentary rocks;

granitoids of Jurassic, Cretaceous(?), and Proterozoic(?) age; a complex of early Miocene dacite to rhyolite domes, flows, and tuffs; early Miocene basalt to basaltic andesite flows and pyroclastic rocks; the 18.5 Marhyolitic Peach Springs Tuff; and conglomerates of post-Peach Springs age. Minor amounts of volcanogenic sandstone and conglomerate occur throughout the Tertiary section. Hazzard (1931) and Kilian (1964) give descriptions of all units in the Marble Mountains, and Glazner (1990) and Glazner and O'Neil (1989) give chemical and isotope analyses of some of the volcanic units.

Pre-Tertiary rocks

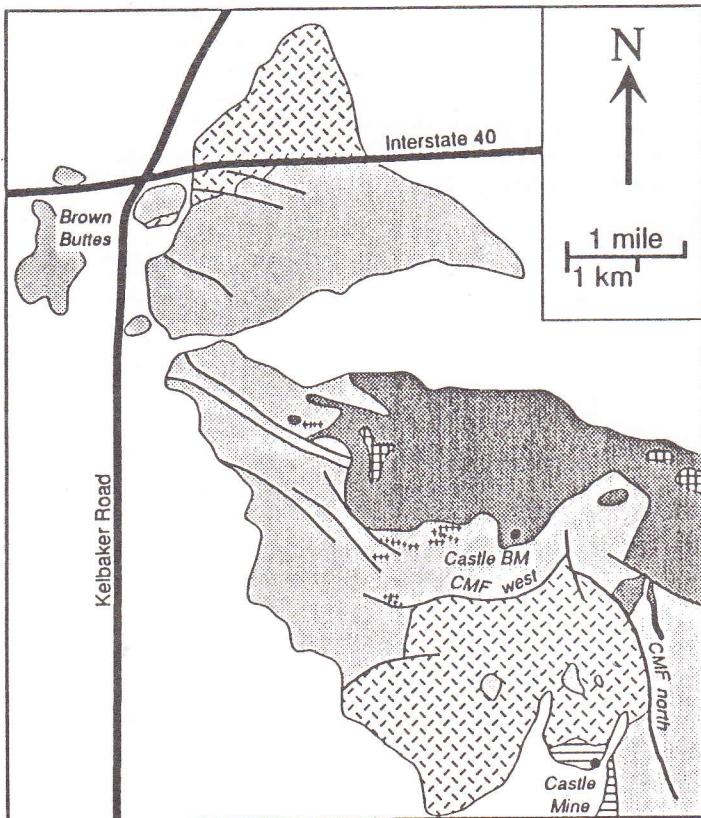
Pre-Tertiary rocks comprise two main groups: granitoids of presumed Jurassic, Proterozoic, and Cretaceous(?) age, and a complex of metamorphic rocks which includes foliated and gneissic diorite, marble, and calc-silicates. Granitoids of presumed Jurassic age (dated mid-Jurassic plutonic rocks crop out in the central part of the range; K.A. Howard, personal communication, 1983) crop out mainly in the southern part of the area and are bounded by the Castle Mine fault (Fig. 3). This group of rocks is distinctly orange in the field. Granitoids north of Interstate 40 are lighter in color and were mapped as Cretaceous by Miller et al. (1982) and

as Proterozoic by Fox and Miller (1990).

Metamorphic rocks crop out in the southernmost part of the area, near the Castle Mine.

Brown Buttes dacite

Dacite and rhyolite flows, domes, and tuffs comprise the lower half of the Tertiary section. These rocks, informally termed the Brown Buttes dacite after exposures along Interstate 40, crop out extensively along the western side of the area and immediately south of Interstate 40. Domes and flows are generally purple



Geology by A.F. Glazner and J. M. Bartley, 1982-83

EXPLANATION









-  Peach Springs Tuff (18.5 m.y.)
-  Basalt flows and scoria
-  Dacite and rhyolite domes, flows, and tuffs (20.2 m.y.)
-  Jurassic, Proterozoic(?), and Cretaceous(?) granitoids
-  Pre-Jurassic(?) plutonic and metasedimentary rocks
-  Fault
-  Contact
-  Basalt dike

Figure 3. Geologic map of the northern Marble Mountains. CMF west and CMF north refer to west- and north-striking segments of the Castle Mine fault.

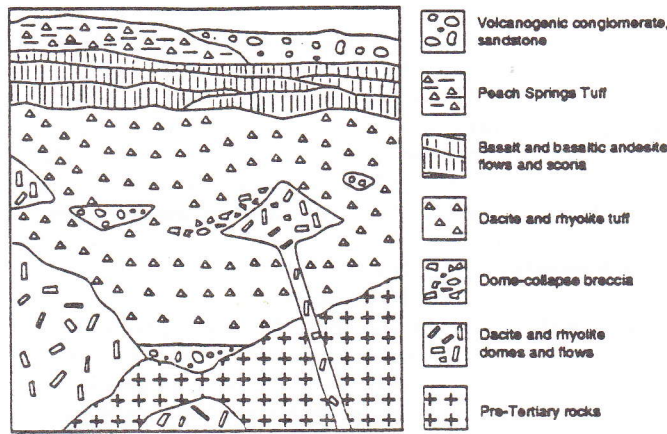


Figure 4. Schematic stratigraphy of the northern Marble Mountains.

to brown porphyritic dacites with phenocrysts of biotite, hornblende, and plagioclase. Rocks of this composition crop out extensively west of the Marble Mountains, in the Bristol, Newberry, Bullion, and Calico Mountains (Glazner, 1990). Locally, impressive monolithic dome-collapse breccias form aprons around domes. Flow bases and dome margins are commonly gray and perlitic.

Tuffs are brilliant white to pale yellow pyroclastic flows and falls. Pyroclastic flows are massive, small pumice flows and block-and-ash flows. Bedding in the pyroclastic falls is prominent and chaotic (see below), and some of the well-bedded tuffs may be planar-bedded pyroclastic surge deposits. Where the base of the Tertiary section is exposed, the tuffs and flows sit upon thin gravel sequences which sit depositionally upon granitoids and metamorphic rocks.

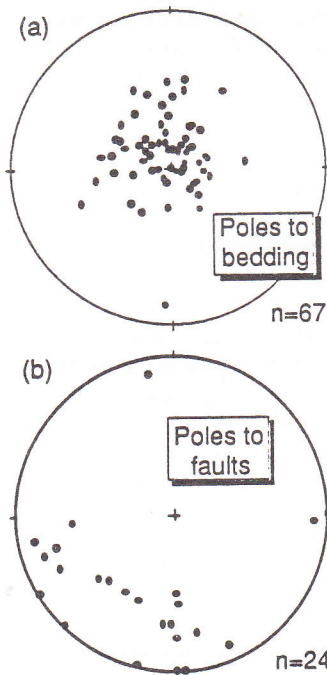


Figure 5. (a) Poles to bedding in tuffaceous units of the Brown Buttes dacite and related epiclastic rocks. (b) Poles to faults (mostly from Brown Buttes dacite).

Castle basalt
Basalt and basaltic andesite flows, informally termed the Castle basalt after excellent exposures along the slopes of the highest peak in the range (Castle benchmark; 3,482'), overlie all phases of the Brown Buttes dacite. However, reconnaissance mapping indicates that basalts are interlayered with dacites southeast of the map area. The basalt flows are typically thin (2-10

m), fluid pahoehoe flows with locally scoriaceous tops. Pods of basalt scoria are common in the cliff exposures west of Castle benchmark, and a cinder cone is exposed in cross section in the central part of the area, in west-facing cliffs under a prominent exposure of Peach Springs Tuff. Basaltic andesite flows form gentler slopes and are commonly platy.

West of Castle benchmark, a swarm of about one dozen basalt dikes cuts the tuff section. These dikes are vertical, trend east-west, and are typically 100-500 m long. Isolated east-west dikes are found in other parts of the area.

Peach Springs Tuff

The Peach Springs Tuff (Glazner et al., 1986) is a prominent cliff-former that is visible to eastbound travelers on Interstate 40. It is 30-40 m thick in the map area. Extensive exposures occur along the eastern slopes of the southern half of the range; at Cadiz Summit, where old U.S. Highway 66 crosses the southern part of the range, the tuff sits directly upon Proterozoic granitoids. Locally, along the eastern slopes of the range, a thin, fresh basalt flow lies directly upon the tuff.

Post-Peach Springs conglomerates

East of the map area, vast exposures of conglomerate cover the volcanic section. Small patches of this conglomerate (not mapped) occur in the map area.

Age control

The age of Tertiary volcanism is constrained by two dates. Gray perlitite from the base of a dacite flow, west of Castle benchmark, yielded a K-Ar age on biotite of 20.2 Ma (J.K. Nakata, written communication, 1984). This flow occurs in the lower part of the tuff section and should give a reasonable approximation of the age of inception of volcanism in the area. The volcanic section is capped by the 18.5 Ma Peach Springs Tuff (Nielson et al., 1990). A sample of Peach Springs Tuff from the Marble Mountains yielded a K-Ar date of 16.2 Ma on sanidine (Nielson et al., 1990). This age is considered spurious, as are many conventional K-Ar ages on the tuff (Nielson et al., 1990).

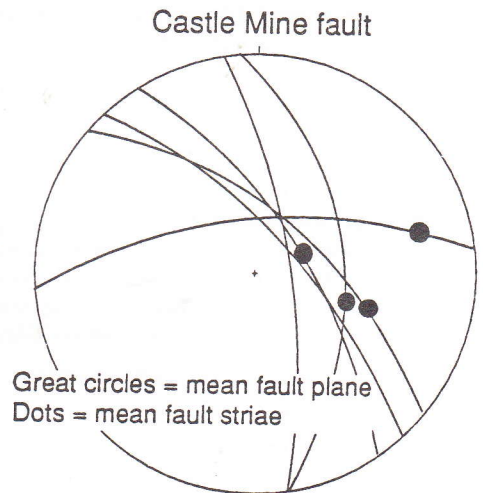


Figure 6. Summary of average fault plane and fault striae measurements from six segments of the Castle Mine fault. Note the anomalously shallow striae on the west-striking segment; these striae cut older, steeply dipping striae (see text).

Structural Geology

Bedding attitudes

Figure 5a summarizes measurements of bedding orientation in the silicic tuffs and associated epiclastic rocks

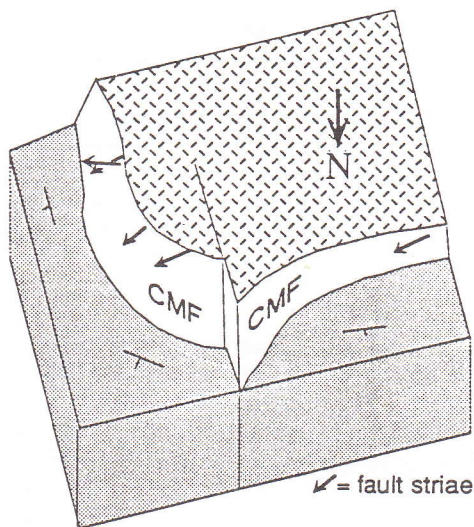


Figure 7. Schematic block diagram of the Castle Mine fault, looking south. Shaded footwall block is dominately dacite; patterned hanging-wall block is composed of Jurassic granitoids.

and tuffites. Bedding is highly scattered in all directions from horizontal, with dips greater than 35° not uncommon. However, virtually all of this tilting is caused either by initial dip or by doming above local dacite plugs. Tuffs commonly dip steeply away from dacite domes, and pronounced angular unconformities are common. The mean bedding attitude in the tuff section is essentially horizontal (Fig. 5a); thus, we infer that there has been little or no tectonic tilting of the Miocene section in the northern Marble Mountains.

Faults

Faulting in the northern Marble Mountains can be divided into three episodes: (1) faulting which formed the western margin of the range, (2) east-west high-angle

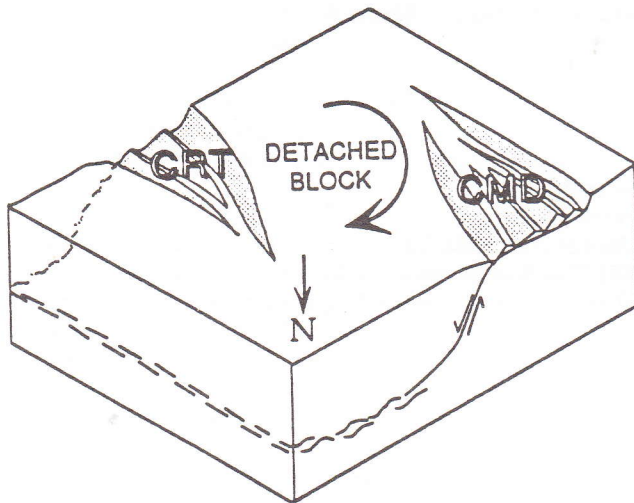


Figure 8. Schematic diagram (after Bartley and Glazner, 1989) showing how coeval extension in the central Mojave Desert (CMD) and Colorado River trough (CRT) might be accommodated by rotation of a panel of weakly deformed rocks. In this model, the Marble Mountains lie in the detached block and should exhibit clockwise rotation.

faulting of pre-Castle basalt age, and (3) faulting related to dacite dome injection. The last two episodes are early Miocene. The age of the first is unknown, but is probably young (middle Miocene or younger?) and possibly ongoing.

Tertiary rocks exposed in the steep western margin of the range contrast sharply with pre-Tertiary rocks in the Bristol Mountains to the west. This juxtaposition probably results from offset along a right-lateral(?), northwest-trending fault which bounds the Marble and Granite Mountains on the west. This fault, although unmapped, is clearly visible on air photos of the Granite Mountains (D.M. Miller, personal communication, 1983), and is presumably part of a series of faults which forms the southern extension of the Death Valley fault zone (e.g., Bishop, 1964; Hamilton and Myers, 1966; Jennings, 1977). Although volcanic rocks are absent in the Bristol Mountains immediately west of the northern Marble Mountains, they reappear farther northwest in the central Bristol and Old Dad Mountains, immediately west of the Granite Mountains. If the basalt dike swarm of the northern Marble Mountains can be found in the Old Dad Mountains, which are unmapped, then a piercing point can be established.

East-west basalt dikes and two west- to northwest-trending high-angle faults suggest that a period of roughly north-south extension occurred around the time of the dacite-to-basalt transition. The faults cut cliff-forming flows of edacite approximately 4 km west-northwest of Castle benchmark, and are overlain by unfaulted basalt flows (Fig. 3). Although sense of slip is not determinable, the faults show north-side-down separation. Similar relationships are seen at Van Winkle Mountain, immediately northeast of the northern tip of the Marble Mountains (D. M. Miller, personal communication, 1983). We infer that these faults are high-angle normal faults which are related to the same stress regime that promoted injection of the basalt dikes. If this scenario is correct, then the faulting probably occurred near the beginning of Castle basalt time. North-south extension of early Miocene age is widespread across the southwestern United States (Best, 1988). Similar relationships (east-west dikes; north-side-down faults) closer to the coast have been explained as flexure of the North American plate above the subducted Mendocino fracture zone (Glazner and Loomis, 1984; Glazner and Schubert, 1985).

Figure 5b shows measured fault attitudes. Most measurements were taken in the Brown Buttes dacite. Most faults are vertical to steeply northeast-dipping and strike north to west. We infer that most of these faults, which commonly have displacements measured in meters, are related to bowing of the section during intrusion.

The largest fault in the area is designated the Castle Mine fault after exposures in the canyon north of the Castle Mine. This intriguing structure forms the northern and eastern boundaries of a block of Jurassic plutonic rocks. It trends due north from the Castle Mine, swings smoothly around to a northwest strike, and then continues westward west a west to southwest strike before losing expression in the dacite. Exposures are excellent, and it is clear that the fault does indeed swing around, changing strike through 90°. Large bodies of intrusive dacite are present along the margins of, and within, the plutonic block. Slickensides and fault-surface corrugations plunge steeply along the north-trending

segment of the fault; along the west-trending segment, shallow-plunging slickensides are superimposed on steeply plunging corrugations.

We interpret the Castle Mine fault to bound a flap of plutonic rocks which was lifted up, trapdoor fashion, by a subjacent dacite pluton. The smooth change in fault strike, coupled with down-dip corrugations, requires psiton-like movement. In this model, the abundant dacite domes that ring the plutonic block are leaks from the subjacent pluton. Along the north-trending segment of nonconformity that bounds plutonic blocks on the west (Fig. 3), tuffaceous sandstones and conglomerates dip 17° to the west. If this region is considered to be a hinge for an eastward-opening trapdoor, then then dip requires a dip-slip displacement of approximately 1.2 km along the north-trending segment of the Castle Mine fault. Geologic relations along this segment require a minimum of approximately 400 m of dip-slip movement.

Regional Relationships

Structural and stratigraphic relationships presented above indicate that the Marble Mountains represent a relatively stable block between two highly extended terrances. Volcanism in the Marble Mountains was roughly synchronous with volcanism and extension in the Whipple and central Mojave terranes, but tectonism was limited to minor north-south extension and local faulting above shallow intrusions. Similar relationships in the nearby Clipper, Van Winkle, Old Woman, and Turtle Mountains areas (Lukk, 1982; Hileman et al., 1990; Hazlett, 1990; D.M. Miller, personal communication, 1983) indicate that a large, continuous block of weakly-deformed ranges separated the Whipple and central Mojave regions in the early Miocene, and that the two were probably not connected at the surface.

The kinematic models have been proposed to link the Whipple and central Mojave terranes. In one (Davis and Lister, 1988; Glazner et al., 1989; Hileman et al., 1990), extension in the central Mojave is slightly older than extension in the Whipple region, and the Marble Mountains region was carried piggyback by the northeast-dipping central Mojave system. In the other (Barley and Glazner, 1989), the Marble Mountains region lies in a panel of weakly deformed rocks between the Whipple and central Mojave terrances, which are lined *en echelon* (Fig. 8). This speclative model accounts for proposed clockwise block rotations in parts of the Mojave block (Ross et al., 1989). Both models are compatible with current data.

Acknowledgments

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The Relative Influences of Climate Change, Desert Dust, and Lithologic Control on Soil-Geomorphic Processes on Alluvial Fans, Mojave Desert, California: Summary of Results

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Introduction

Soil-landscape relations are widely used in desert environments to provide fundamental stratigraphic and age information for Quaternary deposits for studying neotectonics, surficial processes, paleoclimatology, archeology, stratigraphy, and environmental science. Many studies have shown, however, that processes controlling soil development and landscape modification in desert environments are extremely susceptible to the (1) effects of climate change, (2) accumulation of fine-grained eolian sediment (dust), and (3) lithologic control of source rocks. The influence of these variables on soil-landscape processes have not been integrated into one comprehensive study that addresses both the

individual effects, as well as the combined effects, on the origin and development of soils and desert landforms. Several key attributes of the Providence Mountains piedmont (Fig. 1) provide an unprecedented opportunity to learn how the factors listed above control processes of soil-development and landscape modification: (1) a well-defined record of Holocene and Late Pleistocene paleoclimatic and eolian activity in the Mojave Desert, (2) established Quaternary stratigraphy for alluvial fans along nearby mountain piedmonts, and (3) the juxtaposition of four

Figure 1. (a) Location of study area (open box) along the Providence Mountains piedmont (PM) in relation to other areas of Quaternary geomorphic and pedologic studies in the eastern Mojave Desert, California. Map abbreviations: Soda Mountains (SDM), Silver Lake playa (SL), Soda Lake playa (SDL) Mojave River (MR), Baker (B), Union Pacific Railroad (UPRR), Cima volcanic field (CVF), Cima Dome (CD), Cima, (C), Kelso Dunes (KD), Granite Mountains (GM), Kelso (K), New York Mountains (NYM), and the Kelbaker road (KB). (b) General Distribution of Quaternary depositional units along the Providence Mountains piedmont, sample sites for ^{14}C and IRSL age control, and locations of representative soil study sites (Table 1).

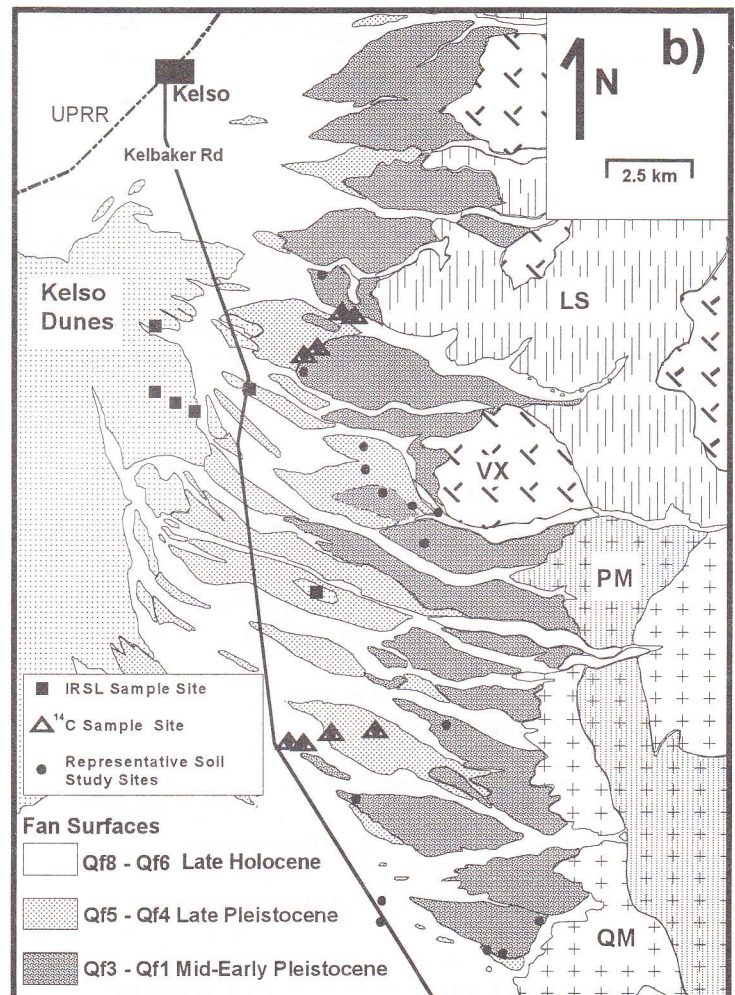
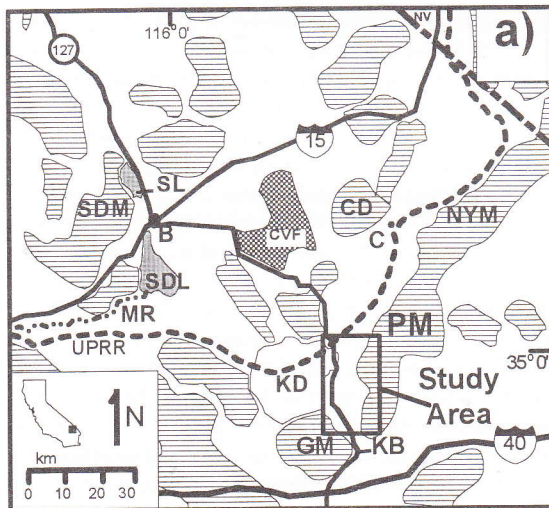


Table 1. Typical soil and pavement characteristics of the four soil-chronosequences.

Fan Unit ^a	Fan-Sequence			
	PM	QM	VX	LS
Qf8 Late Holocene	Horizon: ^b no soil Texture: ^c sand Stage: ^d 0 Pavement: ^e none	no soil sand 0 none	no soil sand to loamy sand 0 none	no soil sand to loamy sand 0 none
Qf7 Late Holocene	Horizon: Avk-AC-Ck-C Texture: loamy sand Stage: I- Pavement: none	Av-A-AC-Ck-C loamy sand I- none	Avk-ACK-Ck-C loamy sand I very weak to weak	Avk-ACK-Ck-C sandy loam I very weak to weak
Qf6 Late Holocene	Horizon: Avk-Bwk-Ck-C Texture: sandy loam Stage: I SB: ^f cambic Pavement: weak	Av-A-Bw-Ck-C loamy sand I cambic very weak to none	Avk-Bwk-Bk-Ck-C loamy sand I cambic weak to moderate	Avk-Bwk-Bk-Ck-C sandy loam I-II cambic weak to moderate
Qf5 Early Holocene to Latest Pleistocene	Horizon: Avk-Bwk-Btk-Ck-C Texture: sandy loam Stage: I-II SB: cambic or argillic Pavement: moderate to weak	fan unit not found	Avk-Bwk-Bk-Ck-C sandy loam I-II cambic/calclac moderate to strong	Avk-Btk-Bk-Ck-C loam II-III calclac moderate to strong
Qf4 Late Pleistocene	Horizon: Avk-Btk-Bk-Ck-C Texture: sandy clay loam Stage: III SB: argillic Pavement: moderate to strong	Av-ABv-Bt-Bk-Ck-C clay loam I-II argillic weak to moderate	Avk-BAv-Bwk-Bkm-Ck-C sandy loam III-IV petrocalclac strong	Avk-Btk-Bwk-Bkm-Bk-Ck-C loam III+ petrocalclac strong
Qf3 Middle Pleistocene	Horizon: Avk-BAv-Bt-Bkm-Ck-C Texture: clay loam Stage: III-IV SB: argillic/petrocalclac Pavement: strong	Av-BAv-Bt-Bck-Ck-C loam I- argillic moderate to weak	Avk-BAv-Btk-Bk-Bkm-Bk-Ck-C sandy clay loam IV argillic/petrocalclac strong	Avk-Btk-Bwk-Bkm-Bk-Ck-C loam IV petrocalclac strong to very strong
Qf2 Mid-Early Pleistocene	Horizon: Avk-Btk/Bkm-Ck-C Texture: sandy clay loam Stage: IV SB: petrocalclac Pavement: moderate to weak	Av-Bt-Btk-Bck-Ck-C sandy clay loam I- argillic weak to moderate	Avk-BAv-Btk-Bk-Bkm-Ck-C sandy clay loam IV argillic/petrocalclac strong to moderate	Avk-Bkm-Bk-Ck-C sandy loam IV-V petrocalclac moderate to strong

PM: mixed-plutonic, QM: quartz Monzonite, VX: mixed-volcanic, LS: limestone

a: Age estimations of fan units in Figure 2. Unit Qf1 not shown because unit is highly eroded.

b: Typical soil horizon sequence. Position of representative soil sites shown in Figure 1.

c: Finest texture of any B or AC horizon.

d: Soil carbonate stage (Gile *et al.*, 1966).

e: General quality of desert pavement where either best developed or where best preserved.

f: Strongest diagnostic B horizon.

lithologically different sequences of Holocene and Pleistocene alluvial fan deposits along the Providence Mountains piedmont. The purpose of this paper is to report on some of our recent findings regarding soil-geomorphic processes that occur on alluvial fans in the Mojave Desert.

Results

Quaternary Stratigraphy

Seven major alluvial fan units and three eolian units were defined based on relative development of soils and desert pavements and stratigraphic relationships among depositional units (Table 1, Fig. 2). We defined four separate fan sequences based upon the dominant rock types that make up the deposits in each sequence: 1) *PM*: leucocratic to mesocratic mixed-plutonic rocks (mostly syenite and syenogranite), 2) *QM*: quartz monzonite, 3) *LS*: limestone and 4) *VX*: volcanic-mixed (rhyolitic tuff and rhyodacite

with lesser amounts of PM and LS).

Age estimates of Quaternary alluvial deposits (Fig. 2) are based on (1) infrared stimulated luminescence (IRSL) ages for eolian sand interstratified with an units (Clarke, 1994), (2) identification of the 0.74 Ma Bishop Tuff near the base of the oldest fan unit, (3) soil-stratigraphic correlations to nearby, relatively well dated fan deposits in the Cima Volcanic Field and Soda Mountain areas (McDonald and McFadden, 1994; McDonald, 1994), and (4) radiocarbon age estimates of pedogenic carbonate (Wang *et al.*, in press). Although they are in correct relative order and generally in the same age range (i.e. Holocene and latest Pleistocene) preliminary radiocarbon dates of pedogenic carbonate do not agree with age estimates based on IRSL dates, current knowledge of soil-forming rates in the Mojave desert, and soil-stratigraphic correlations to nearby and relatively well-dated piedmont deposits. Inconsistencies of radiocarbon dates may reflect the possible influence of older, detrital

Study Area	Providence Mountains ^a				Soda Mountains / Silver Lake and Vicinity ^b				Cima Volcanic Field ^c			
	Fan Surface	Age Estimations (X1000) yr ^d Min Max	IRSL ^e	Numerical Age Estimations 14C ^f Model ^g	Fan Surface	Age Estimations (X1000) yr ^d Min Max	IRSL ^e	Numerical Age Estimations 14C ^f Model ^g	Fan Surface	Age Estimations (X1000) yr ^d Min Max	IRSL ^e	Numerical Age Estimations 14C ^f Model ^g
Quaternary	Late	Qf8	0 2.8 (7) 0.1			Qf6	0 0 0		Qf9	0		
		Qf7	0 4.2 (8) 0.5-1		4410 ± 110 7-8 5110 ± 60 4-6	Qe3	0 0.8 0.6 2.3		Qf8	3.6		
Holocene	Late	Qe3	0 4.2 (8) 3.5	4250 ± 290		Qf4	4.9 3.9 7.9	3400 ± 60	Qf7	6.5		
		Qf6	0 8.4 (11) 4		5380 ± 80 8-11 4010 ± 70 7-8	Qf5 ^m	7.5 5.3 6.6	3620 ± 70	Qf6	6.1		
Early	Middle	Qe2	8.4 10.4 (29) 8.5	3500 ± 220 3700 ± 425 4074 ± 334 8420 ± 795		Qe2						
		Qf5	8.4 16.8 (36) 12	12460 ± 1151 16310 ± 100 27-33	12.1	Qf2/ Qf2	14.3 12.2 19.7	8350 ± 300 9160 ± 400 10333 ± 120 13670 ± 550	Qf5	12.4		
Pleistocene	Late	Qe1	10.4 22 (45) 15	16830 ± 1465 17300 ± 1935		Qe1						
		Qf4	17.3 90 50	28690 ± 440 48-57 26980 ± 290 40-48	25.9	Qf1	19.3 20.3 46.4	14660 ± 260 20320 ± 740	Qv6		18 ± 12 20 ± 10	
Middle	Early	Qf3	>36 130 130		47.0				Qv5		0.06 ± 0.02 0.09 ± 0.04	3 ± 7 4 ± 6
		Qf2	>170 <740 650	0.74 Ma Bishop Tuff		Qf1	?	?	Qf1		0.13 ± 0.02 0.15 ± 0.01 0.16 ± 0.02 0.17 ± 0.02	80 ± 10 65 ± 9 74 ± 7
Neogene	Early	Qf1	>730 4000 1500			Qf1			Qv1		0.24 ± 0.02 0.32 ± 0.05	0.46 ± 0.04 0.51 ± 0.02 0.58 ± 0.08
		Qf1				Qf1			Qf1		3.88 ± 0.05	

a = This study.
 b = Soda Mountains/Silver Lake stratigraphy from Wells et al. (1987, 1990).
 c = Cima volcanic field stratigraphy from Wells et al. (1990); correlations between Cima and Silver Lake stratigraphic sections by Wells et al. (1990); Qv = Quaternary volcanics
 d = Maximum surface ages using modeled radiocarbon results from Wang et al. (1994).
 e = IRSL dates from Clarke (1994).
 f = Oldest radiocarbon age estimates from soil organic matter and soil carbonate (upper PM soils, lower LS soils) reported in Wang et al. (1994) and listed in Tables 7, 8.
 g = Oldest modeled radiocarbon ages (x 1000 yrs; upper PM soils, lower LS soils) from Wang et al. (1994).
 h = Soil development index values calculated according to Harden (1982), and Harden and Taylor (1983).
 i = Values for Soda Mountains piedmont, from McFadden et al. (1989).
 j = Values for Salt Springs Hills area, from McFadden et al. (1989).
 k = Values for Soda Mountains piedmont and Cima piedmont, from Harden et al. (1991).
 l = Age estimations reported in Wells et al. (1990).
 m = Unlabeled late Holocene lake stand, Silver Lake playa, informally defined here for illustration.
 n = Cosmogenic 3He surface-dates on basalt and pavement clasts Wells et al. (1994).

Figure 2. Regional correlations and age estimations among Quaternary fan, lacustrine, and eolian deposits in the eastern Mojave Desert.

carbonate (Pendall et al., 1994) or the IRSL dates may be incorrect. The IRSL technique used has been validated in part, however, based on accurate ages derived for eolian sands bracketing Mount Mazama tephra in Oregon (Clarke, 1994).

The Influence of Quaternary Climate Change and Lithologic Control on Periods of Fan Deposition

Regional correlations and geomorphic relations among Quaternary alluvial fan and eolian deposits in the eastern Mojave Desert indicate that depositional units along the

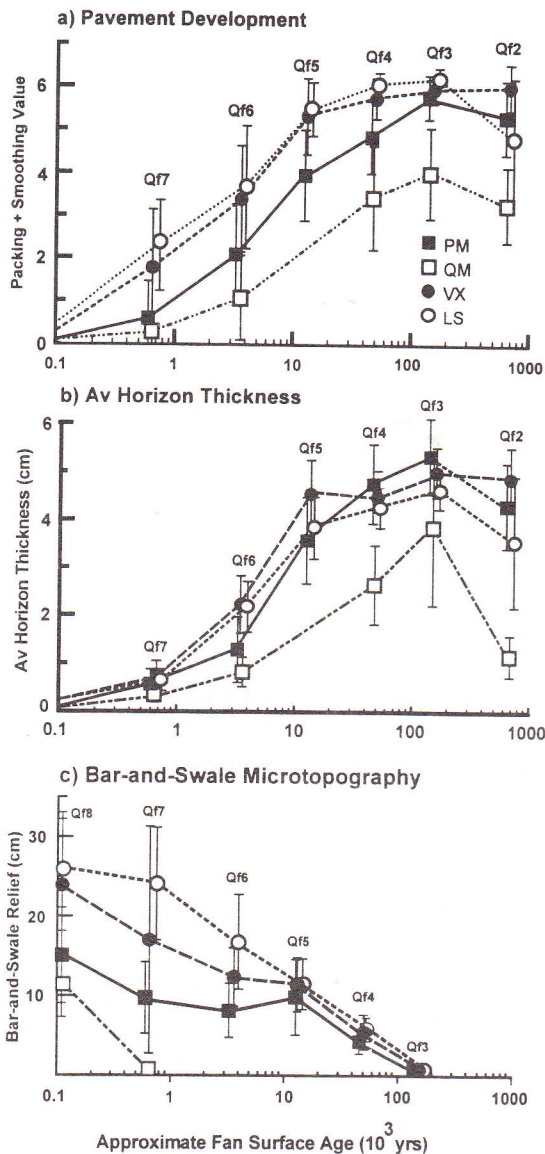


Figure 3. Temporal trends in mean (a) Av horizon thickness in cm, (b) pavement development based on mean alignment (packing) and smoothing (horizontal character) values, and (c) bar-and-swale microtopography. Each point represents between 20 and 50 measurements for each particular fan sequence. Bars represent \pm one standard deviation. Displayed age for each set of fan units represents best age approximations. Ages of individual fan-units within each set of fan units are slightly offset to increase visibility of error bars.

Providence Mountains piedmont appear to have been deposited in response to some aspect of regional climate change. Stratigraphic relations and age estimates among latest Pleistocene and early Holocene fan and eolian units indicate that an extensive period of regional fan deposition occurred between about 8,400 and 16,300 yrs B.P. presumably in response to destabilization of drainage basin hillslopes (McDonald et al., 1994). The regional extent of fan deposition during the Pleistocene-Holocene transition strongly supports models of Bull (1991) and Wells et al. (1987) that an extensive period of fan aggradation can result

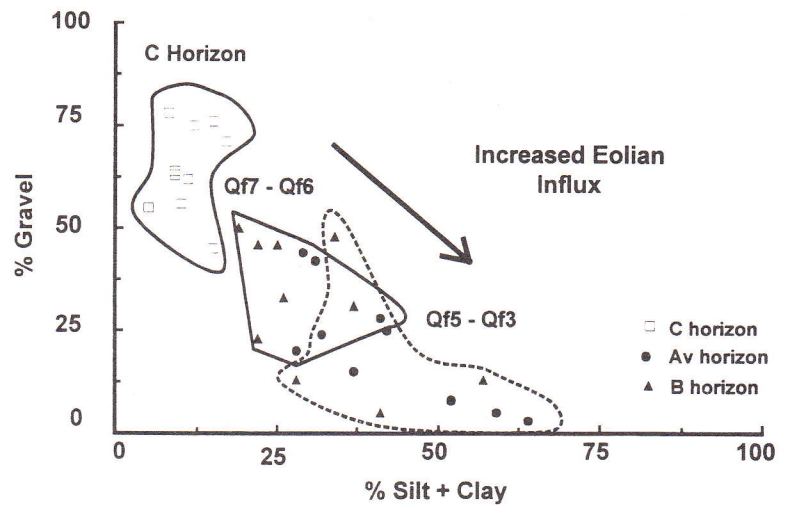


Figure 4. Changes in silt + clay (% wt.) and gravel (% wt.) due to accumulation of dust below pavement stones for 10 soils formed on Qf7 to Qf3 fan surfaces. Soils formed in all four rock types represented. Each soil has a plotted value for an Av, AC or B, and C horizon texture.

from a destabilization of hillslopes during a cycle of glacial-to-interglacial climatic transition.

Other intervals of fan deposition along the Providence Mountains piedmont also appear to be related to some aspect of climate change. Regional fan deposition occurred between about 4,500 to 3,500 yrs B.P. and appears to have been associated with a brief interval of late Holocene pluvial lake activity (Fig. 2). At least two intervals of fan aggradation occurred during the late and middle Pleistocene with one occurring between about 22,000 and 90,000 yrs B.P. and the other occurring between about 36,000 and 130,000 yrs B.P. These range estimates for periods of fan aggradation generally overlap with intervals of glacial-to-interglacial climate change. The most extensive fan deposit along the Providence Mountains pediment is underlain by a tephra correlative to the 0.74 Ma Bishop Tuff. We surmise that the large size and extent of this deposit suggests that (1) deposition may have largely been by perennial streams rather than by ephemeral streams, and (2) that deposition occurred during a sustained interval of significantly wetter climate that was of greater duration than subsequent periods of wetter climate associated with intervals of glaciation. Periods of increased deposition of eolian sand along the Providence Mountains piedmont also appear to have resulted from regional fluctuations in the supply of eolian sediment associated with fluctuations in nearby pluvial Lake Mojave.

The influence of the drainage-basin rock type on periods of fan deposition appears to have been secondary to the influence of regional changes in climate on fan deposition. Similar stratigraphic sequences of fan deposits along the Providence Mountains piedmont that are physically correlative have been derived from drainage-basin source areas consisting of such diverse rock types as coarse-grained plutonic, micro-crystalline siliceous, and massive carbonate rocks. Lithologic control, however, had a strong influence on hillslope response to climate change. Sediments derived

from plutonic rocks (PM, QM) appear to have been derived from the erosion of an extensive cover of soils and colluvium; whereas, sediments derived from microcrystalline siliceous and carbonate rocks (VX, LS) may have largely been derived from talus and colluvium that had accumulated along drainages.

The Influence of Desert Dust and Lithologic Control on the Origin and Evolution of Desert Pavements

Temporal relations among the fan chronosequences show that there is a strong systematic trend between increases in the quality of the desert pavement and increases in the thickness of underlying Av horizons (Figs. 3 a, b). These trends indicate that the formation of pavements and Av horizons: (1) begins soon after deposition of the alluvial deposits, (2) co-evolve simultaneously, and (3) pavements on alluvial fans form by the upward lifting of surface clasts from the accumulation of eolian fines below pavement clasts and not deflation or upward migration of clasts as had been generally hypothesized. These results indicate that the accretionary model developed by McFadden *et al.* (1987) that describes development of desert pavements on basalt flows from eolian accumulation also largely applies to pavement formation of alluvial fans. Results also indicate that Av

horizons on alluvial fans also accrete vertically over time due to the continued accumulation of dust that is translocated through the Av horizon and into underlying soil horizons (McDonald, 1994). As a result, the rising Av horizon enhances pavement formation by creating a stable platform for uplift of pavement clasts.

Our results also indicate that important differences occur among pavements derived from different rock types due to variations in dust trapping efficiency and bar-and-swale microtopography (Fig. 3c). Av horizons and desert pavements develop fastest on Holocene fan-surfaces composed of resistant rock types that typically have features that increase dust trapping efficiency, such as (1) coarse-grained texture (cobble-gravel) with an initially open framework and (2) produce a strong bar-and-swale microtopography. By contrast, Av horizons and pavements form more slowly on fan surfaces composed of poorly-resistant rock types, such as quartz monzonite, that produce finer-grained deposits (grus) and have a poor bar-and-swale microtopography. Pronounced degradation of bar-and-swale microtopography occurs concomitant with increased pavement development leading to the formation of broad, smooth pavement-covered surfaces that are common sights across the Mojave and other desert regions. Temporal increases in the abundance of cracked, split, and spalled cobbles and boulders at the surface, especially in bars,

provide a ready source of additional clasts for forming strong, interlocking pavements.

Influence of the Accumulation of Desert Dust on Soil-Formation

Soil formation on fans composed of all rock types show systematic increases in clay and silt in A and upper B horizons resulting from the accumulation of dust (Fig. 4). In soils formed from granitic rock types (PM and QM fan-sequences), dust provides the primary source of calcium carbonate. In soils formed in all rock types the accumulation of dust provides essentially all temporal increases in clay- and silt-sized particles, with only small amounts of the secondary fine-grained material in these soils being derived from to the physiochemical alteration of primary soil minerals.

Profile distributions of major oxides and trace elements in soils formed on fans composed of all rock types also demonstrate that the accumulation of dust is the primary source of temporal increases in silt and clay. Significant changes in ratios of Zr/Cr and Ti_2/MnO in Av horizons relative to the parent material (Fig. 5) indicate pedogenic mixing of dust within upper soil horizons. The accumulation of dust is also consistent with a noticeable decrease in the variation of ratios in Av horizons relative to ratios for the different parent materials.

Influence of Lithology on Soil Formation

The results of this study provide the first direct comparison of soils developed from different and contrasting parent materials (Table 1). Clay-rich argillic (Bt) horizons and strongly oxidized cambic (Bw) horizons have formed in soils developed from granitic materials (QM and PM) but these soil features

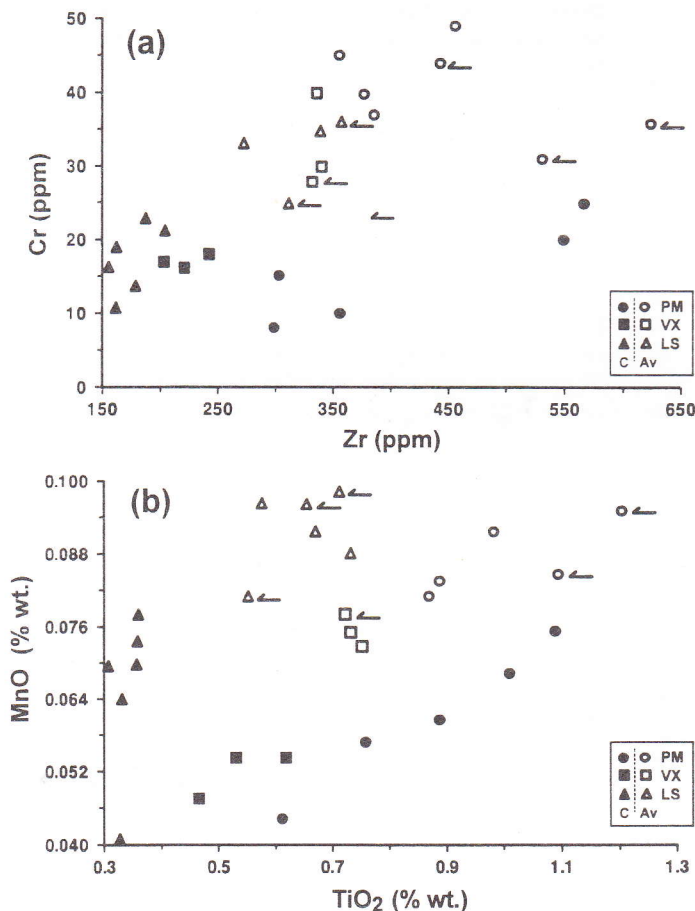


Figure 5. Ratios of trace elements Cr and Zr and major oxides MnO and TiO₂ for Av horizons (open symbols) and soil parent material (C or Ck horizons; solid symbols) for 14 soils formed on the PM, VX, and LS fan surfaces. Arrows indicate late Holocene Av horizons (Qf7 and Qf6 soils).

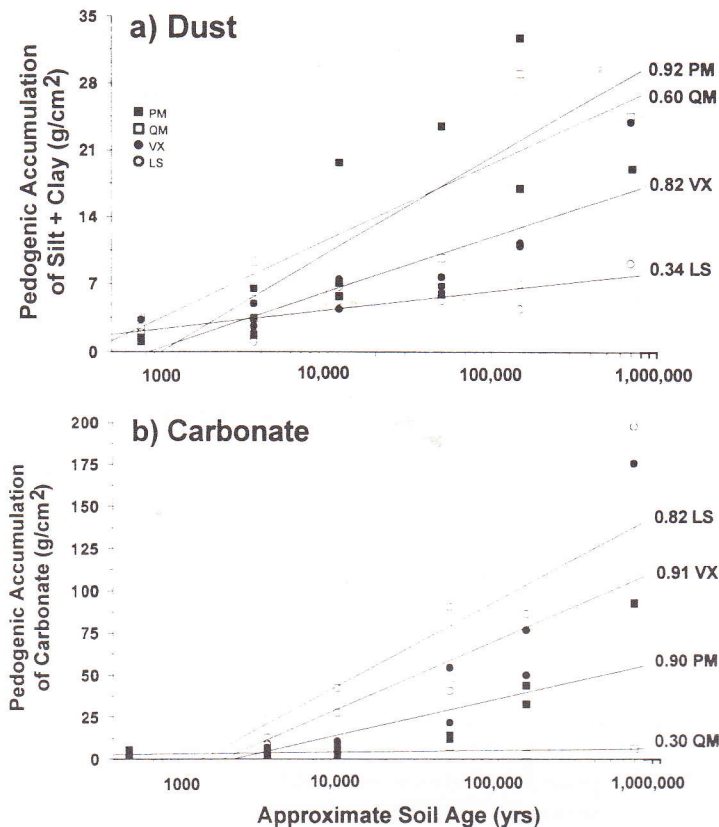


Figure 6. Linear regression estimates of rates of pedogenic accumulations of (a) silt + clay (g/cm^2) and (b) carbonate (g/cm^2) in the four soil-chronosequences. Best age estimates for soils are used in regression analysis.

are virtually absent in soils developed from calcareous materials (LS). By contrast, the accumulation of pedogenic carbonate is strongest in the LS soils and weakest in the PM and QM soils. Soils developed from mixed deposits (VX fan-sequence) generally represent a general half-way point between soil developed from granitic materials and soils from calcareous materials.

Significant variation occurs among the pedogenic accumulation of silt and clay among the four soil-chronosequences, indicating that lithology of the soil parent material must exert some type of control on the rates of dust accumulation (Fig. 6 a). Accumulation of dust-derived silt and clay in LS and VX soils totals about 5 to 10 g/cm^2 in about 50,000 years (Qf4 fan surfaces), whereas 8 to 30 g/cm^2 of silt and clay have accumulated in QM and PM soils. Variations in rates of dust accumulation may be largely due to temporal variations in the rate of desert pavement development. The development of the desert pavement occurs at a relatively faster rate on the VX and LS fan deposits relative to the rate of pavement development on the PM and QM fan (Fig. 3). A reduction in dust-trapping efficiency would coincide with reductions in surface turbulence associated with increases in the packing and smoothing of clasts during development of desert pavements. Recognition of this type of intrinsic feedback between soil formation and the modification of alluvial fan surfaces is critical for interpreting the systematic changes in desert soil formation among soils

formed in the different rock. Reductions in dust accumulation in soils from the development of pavements have been suggested for reg soils in Israel (Amit and Gerson, 1986).

Significant differences also occur among rates of accumulation of pedogenic carbonate among the four soil-chronosequences (Fig. 6 b). Temporal increases in carbonate in the LS and VX soils are nearly 2 to 3 times the rate of carbonate accumulation in the PM soils reflecting the greater abundance of parent material carbonate in the LS and VX soils. Temporal increases in carbonate in PM soils generally reflects temporal increases in dust and the concomitant additions of eolian carbonate. A lack of temporal increases in carbonate content in the QM soils probably reflects the greater leaching of carbonate through the QM soil profile due to greater effective soil moisture, due in part to a combination of higher elevation, sandy-pebble texture, and poor development of pavements and Av horizons all of which favor greater infiltration and leaching of carbonate.

Influence of the Accumulation of Desert Dust on Soil-Water Balance

Perhaps one of the most significant results of our study of dust accumulation in soils is in demonstrating the strong effect that temporal accumulations of dust has on soil-water balance. The development of Av horizons and desert pavements has a profound impact on infiltration rates (Fig. 7). Measured infiltration rates are reduced to about 35% to 1% of the infiltration rate for coarse-textured deposits that do not have Av horizons. Accumulation of desert dust into soil B horizons that underlie the Av horizon also have a strong effect on soil-water balance by decreasing infiltration rates and increasing the soil-water holding capacity. Recognition of the strong impact that temporal increases in texture has on soil hydrology provide a very new direction in which to evaluate soil formation and landscape evolution. For example, increases in surface runoff that coincide with decreases in infiltration probably result in the degradation of

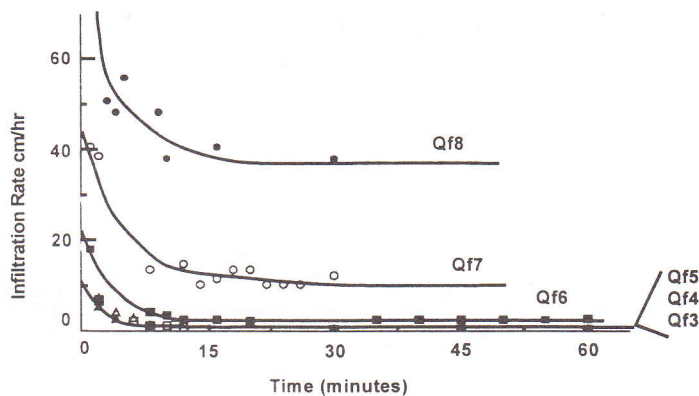


Figure 7. Infiltration rates for soils formed on the PM Qf8 through Qf3 surfaces. Infiltration measured by double-ring infiltrometry with a diameter of 60 cm for the inner ring, a diameter of 100 cm for the outer ring, and 2-3 cm of hydraulic head.

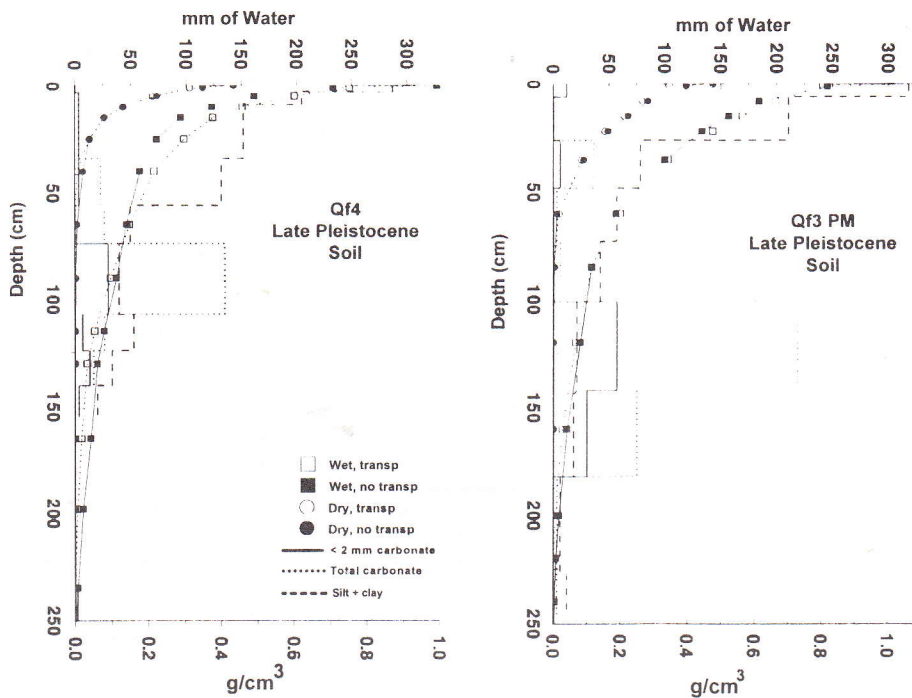


Figure 8. SHAW simulation results of the annual flow of soil-water for wet and dry years and for simulations run with and without transpirational-water loss for Pleistocene soils. The annual flow of soil-water is superimposed over profile distributions of carbonate (fine-earth and whole-soil) and silt + clay.

soils and desert pavements, eventually leading to the establishment of surface drainage and the eventual widespread dissection of alluvial fan surfaces. In other words, the systematic development of soils and desert pavements directly controls long-term surface stability, promoting the eventual self-destruction of alluvial fan surfaces. The soil surface and desert pavements show extensive degradation on the Qf3 and Qf2 surfaces that are underlain by argillic and/or petrocalcic horizons that impede infiltration.

The Influence of Holocene Climate Change on the Formation of Calcic Soils

Carbonate distribution in soils formed on Pleistocene surfaces is commonly bimodal, with carbonate accumulation occurring within the upper and lower parts of the soil profile. The bimodal distribution of carbonate is generally thought to correspond to a general shift to relatively drier climates in conjunction with the Pleistocene-Holocene climatic transition (line A, Fig 8). Paleoclimate records derived from nearby pluvial lakes, however, indicate that the Holocene has been punctuated by brief episodes (50-400 yrs) of significantly wetter climate with a significant period of pluvial activity at about 3600 yrs B.P. (Wells, et al., 1989; Enzel et al., 1989, 1992; line B, Fig. 8). Historic flooding of desert playas in the Mojave Desert appears to be generally associated with significant regional increases in rainfall from Pacific frontal storms as a result of a southern shift in the jet stream and frontal storm track during the winter and spring (Enzel, et al., 1989, 1992).

We used a process-based soil-water balance model (SHAW: Flehinger and Pierson, 1991) to simulate soil-water movement under varied conditions of climate, vegetation, and soil texture to determine if these episodic periods of extreme climatic during the Holocene have influenced soil-water movement, and in turn, the profile distribution of soil carbonate. Simulations were run with actual daily weather data associated with relatively "wet" (historic playa-flooding years with extreme increases in rainfall) and "dry" climate (historic average annual rainfall) to simulate the affects of wet and dry climate.

Results indicate that soil-water flow associated with dry and wet years strongly corresponds with the upper and lower zones of carbonate accumulation in soils that have bimodal distribution of carbonate (Fig. 9). The depth of soil-water flow only reached the

lower zone of carbonate accumulation during a "wet" year when extreme increases in winter/spring storm activity resulted in a nearly 100% increase in annual precipitation. By comparison, the depth of soil-water movement during a "dry" year corresponds to the upper zone of carbonate accumulation in soils formed on pre-Holocene surfaces. Because of the linkage between increases in frontal storm activity and the flooding of desert playas, we believe that this relation suggests that the carbonate distribution in calcic soils in the Mojave Desert may have been strongly influenced by episodic and extreme climatic events that have occurred throughout the Holocene and resulted in the formation of perennial lakes (line B, Fig. 8). If this relation is true, then much of the carbonate in the upper 75 cm of Pleistocene soils may have accumulated during the late Holocene rather than climate change associated with the

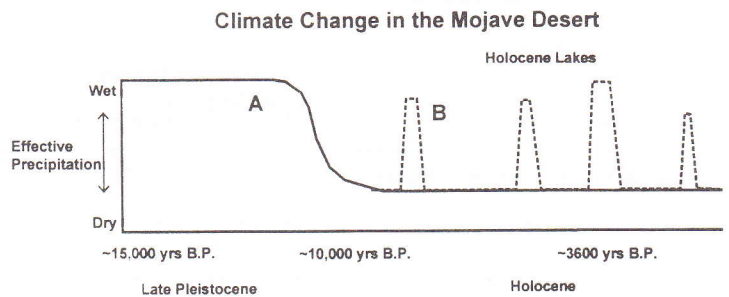


Figure 9. Schematic diagram of models of time-transgressive climate in the Mojave Desert. Line A represents general onset of aridity concomitant with the Pleistocene-Holocene climatic transition, line B represents fluctuations in Holocene precipitation associated with periods of perennial lakes the covered the Silver and Soda playas during the Holocene.

Table 2. Pedogenic carbonate contents in late Holocene and Pleistocene soils.

	Fan Surface	Estimated Soil Age (yrs)	Upper Profile Total g/cm ² a	Profile Total g/cm ² b
Late Holocene	Qf7	750	1.12	
	Qf7	750	1.74	
	Qf7	750	2.90	
	Qf6	3600	2.18	
	Qf6	3600	0.96	
	Qf6	3600	0.80	
Pleistocene	Qf5	12000	1.87	2.42
	Qf5	12000	0.98	1.20
	Qf4	50000	1.10	8.87
	Qf4	50000	1.52	7.15
	Qf3	150000	1.90	26.63

a) all carbonate in late Holocene soils and only carbonate within the upper zone (dry, <75 cm) for Pleistocene soils.

b) Total profile mass of carbonate for Pleistocene soils.

Pleistocene-Holocene transition (line A, Fig. 8). Similar carbonate contents in the upper zones of Pleistocene soils and late Holocene soils is consistent with this hypothesis (Table 2). These results suggests that the distribution of carbonate in calcic soils may provide a proxy record of Holocene climatic change in the Mojave Desert.

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Natural History Notes: East Mojave Desert (Granite, New York, and Providence Mountains, Kelso Dunes, Mid Hills, and Hole-in-the-Wall)

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Introduction

California has three main deserts: from north to south, the Great Basin, Mojave and Colorado (Sonoran) deserts. All lie to the east of the major mountain ranges (Sierra Nevada, Transverse, and Peninsular) and are the combined product of rain-shadow effect and global position. Traditionally, a region that receives less than ten inches of precipitation per year is classified as a desert. The driest place in the state of California is Death Valley, within the boundary of the East Mojave Desert. However, the former town of Bagdad in the East Mojave Desert once went for 767 days without any precipitation. Other characteristics of deserts are temperature extremes, wind, high light intensity, nutrient-poor alkaline soils and low rates of primary productivity.

This trip takes us into the Mojave Desert, specifically the East Mojave Desert within the East Mojave National Preserve. The Mojave Desert is south of the Great Basin desert. Geologically, its northern boundary is the Garlock Fault and its southwestern the San Andreas Fault. The Mojave extends east across the Colorado River into northwestern Arizona and southern Nevada, near Las Vegas. The Transverse Ranges form the southwest boundary. The Mojave is bounded to the northwest by the Tehachapis and the southern Sierra Nevada.

The Mojave Desert is the "high desert" at an elevation of about 3500 feet (1100 meters). Yet it contains the lowest point in the United States (Badwater in Death Valley). The valleys of the Mojave Desert include the Antelope, Apple, and Yucca in the west, Lanfair Valley and the basins (sinks) of Death Valley and Ivanpah to the east.

The Mojave is a hot desert. Rain occurs in the winter, but summer thundershowers are frequent in some parts of this desert. Snow is common at higher elevations in the New York, Providence, Granite and the Clark Mountain Ranges. At elevations above 3800 feet (1200 meters), the most conspicuous plant is the Joshua tree. The distribution of the Joshua tree is essentially an outline of the boundaries of the Mojave Desert. Above 4300 feet (1300 meters), in rocky soils and along ridges, pinyon pines predominate. Below the Joshua tree community lies the majority of the sandy-soiled Mojave Desert dominated by a creosote bush scrub community. The creosote bush is the most abundant shrub in California, covering 80% of the Mojave Desert. Where sandy soil gives way to clay soils of dry lake beds, creosote bush is absent. The alkaline soils are dominated by halophytic (salt tolerant) communities. Cacti are more abundant in the Mojave Desert than they are in the Great Basin Desert, but are mostly confined to rocky slopes and fans. Desert willows (*Chilopsis linearis*), catclaw acacia (*Acacia greggii*), and honey mesquite (*Prosopis glandulosa*) line the washes.

The vegetation and wildlife components in the East Mojave reflect the interdigitation of the three major American deserts: the Great Basin, Mojave and Colorado (Sonoran) Deserts. The East Mojave contains both flora and fauna components of these deserts as well as some elements from the California Coastal and Arizona Interior chaparral zones. The Mojave Desert is considered a unique floristic unit.

Canyons of the New York Mountains have species of manzanita (*Arctostaphylos* spp.), California lilac (*Ceanothus* spp.), oak (*Quercus* spp.), and silktassel (*Garrya* spp.), characteristic of coastal California. The Mid Hills have large stands of Great Basin sagebrush and Utah juniper. Sonoran species such as smoke tree (*Psoralea spinosus*), ocotillo (*Fouquieria splendens*) and desert lavender (*Hyptis emoryi*) are found extending into the East Mojave.

Seven hundred species of plants are known from the East Mojave. Thirteen unusual plant assemblages occur in the area (Table 1). Approximately 60% of the East Mojave Desert is Creosote Bush Scrub, 20% Joshua Tree Woodland, and 15% Pinyon-Juniper Woodland. The remaining 5% are various assemblages.

Thirty-five wildlife habitat types are recognized. Some 300 species of animals are found within these habitats. Thirty-six species of reptiles, 200 species of birds, and 47 species of mammals have been reported. These include the gila monster, desert tortoise, Mojave fringe-toed lizard, regal ring-necked snake, and desert striped whipsnake. Important bird species include prairie falcon, Bendire's thrasher, Crissal thrasher, gray vireo and Lucy's warbler. Rock squirrels, woodrats, porcupines, mountain lions, bobcat, and a very diverse bat fauna (some eleven species) are found within the Mojave Desert.

There are five sensitive wildlife species found in the East Mojave: desert tortoise (*Gopherus agassizi*), desert big horn sheep (*Ovis canadensis nelsoni*), Mohave tui chub (*Gila bicolor mohavensis*), banded gila monster (*Heloderma suspectum*), and the gilded northern flicker (*Colaptes chrysoides*). The desert tortoise and the gilded northern flicker are fully protected under California law and may not be collected. The banded gila monster is observed infrequently, with three of the last five sightings in California in the East Mojave.

Desert Communities

Deserts are divided into communities more on the basis of soil types and latitude than on elevations. Soil types influence the vegetation, while latitude plays a role in temperature and day length. These factors define the biotic conditions of the desert zones. We will pass through several major plant communities on this trip through the East

Mojave Desert: Creosote Bush Scrub, Joshua Tree Woodland, Pinyon-Juniper Woodland, and Sand Dunes.

Creosote Bush Scrub Community

This community dominates the East Mojave Desert to an elevation of about 3500 feet (1100 meters). The dark green leaves and scraggly, dark gray branches make creosote bush easy to identify from a distance. Creosote bush (*Larrea tridentata*) is named for its strong resin odor. The sweet, resin scent is very noticeable after a rain. With two root systems — a tap root that may extend 40 feet (12 meters) or more down into the ground, and numerous small surface roots — the creosote bush is well suited to obtain water. It has been estimated that some creosote bush plants are over 10,000 years old, making them among the oldest living organisms on earth. This is possible due to the cloning of individuals from a parent plant. Each individual of a clone is in contact with a genetically identical individual and with the original parent bush. The shape of such clones is a ring. The so-called "King Clone" is about 60 feet (18 meters) in diameter and is located in the Mojave Desert. *Ambrosia dumosa* (Burro Bush) is the other dominant plant. A variety of other shrubs and annuals such as range ratany (*Krameria parvifolia imparta*), spiny hopsage (*Grayia spinosa*) and desert thorn (*Lycium andersonii*) are associated with this community.

The animal inhabitants of the Creosote Bush Scrub community exhibit a great diversity. Representative mammals include the antelope ground squirrel (*Ammospermophilus leucurus*), black-tailed hare (*Lepus californicus*), deer mouse (*Peromyscus maniculatus*), southern grasshopper mouse (*Onychomys torridus*), several species of pocket mice (*Perognathus* spp.), kangaroo rats (*Dipodomys deserti*, *D. merriami*), desert cottontail (*Sylvilagus auduboni*), kit fox (*Vulpes macrotus*), coyote (*Canis latrans*), and bobcat (*Felis rufus*).

Birds are very numerous in the East Mojave. On a good day some 25 to 30 species can be observed. These may include the greater roadrunner (*Geococcyx californianus*), black throated sparrow (*Amphispiza bilineata*), sage sparrow (*Amphispiza belli*), flycatchers, common raven (*Corvus corax*), red-tailed hawk (*Buteo jamaicensis*), prairie falcon (*Falco mexicanus*), and the golden eagle (*Aquila chrysaetos*). Reptiles are abundant: side-blotched lizard (*Uta stansburiana*), long-tailed bush lizard (*Urosaurus graciosus*), zebra-tailed lizard (*Callisaurus draconoides*), desert iguana (*Dipsosaurus dorsalis*), common leopard lizard (*Gambelia wislizenii*), western whiptail lizard (*Cnemidophorus tigris*), western banded gecko (*Coleonyx variegatus*), collared lizard (*Crotophytus collaris*), chuckwalla (*Sauromalus obesus*), glossy snake (*Arizona elegans*), gopher snake (*Pituophis melanoleucus*), common kingsnake (*Lampropeltis getulus*), coachwhip (*Masticophis flagellum*), Mojave rattlesnake (*Crotalus scutulatus*), and the desert tortoise (*Gopherus agassizi*).

Joshua Tree Woodland

Yucca brevifolia, the Joshua tree, is considered the indicator species of the Mojave Desert. It does occur outside the Mojave in a few areas, such as the Sonoran Desert of west-central Arizona. In addition to Joshua tree, two other species of yucca are found in the woodland: *Yucca baccata*, the Spanish bayonet with long bluish leaves,

Table I. Unusual plant assemblages.

White Fir	Two isolated populations of Rocky Mountain white fir (<i>Abies concolor concolor</i>) occur in the Scenic Area in the New York and Clark Mountains. Along with a population in the Kingston Mountains, these small groves represent the westernmost occurrence of Rocky Mountain white fir. Considered "very sensitive."
Chaparral	An "enriched" pinyon-juniper/oak woodland with affinities to Arizona interior chaparral and California coastal chaparral occurs in Caruthers, Keystone and Live Oak Canyons in the New York Mountains. Considered "sensitive."
Calcareous Scrub	Vegetation associated with limestone and dolomitic outcrops occurring in the Providence, New York and Clark Mountains. Characterized by the occurrence of many uncommon plants. Considered "highly sensitive."
Sagebrush Scrub	Great Basin sagebrush (<i>Artemisia tridentata tridentata</i>) occurs in the Round and Gold Valleys in the Mid Hills area. Considered to be "subject to disturbance, but not sensitive."
Desert Grassland	Characterized by the presence of approximately 20 species of perennial grasses, this assemblage is found in Lanfair Valley. Considered to be "subject to disturbance, but not sensitive."
Shadscale Scrub	This stand of <i>Atriplex confertifolia</i> occurs at Valley Wells and is characteristic of alkaline soils of the Great Basin Desert. Considered to be "subject to disturbance, but not sensitive."
Joshua Tree Woodland	The Cima Dome area supports a very dense and extensive stand of <i>Yucca brevifolia</i> var. <i>jaegeriana</i> . Considered to be "subject to disturbance, but not sensitive."
Kelso Dunes	The Kelso Dunes support vegetation that is highly adapted to life in the sand, including a number of perennial grasses. Considered "very sensitive."
Huge Mojave Yucca	The slopes of the Hackberry, Woods and Providence Mountains support stands of very large <i>Yucca schidigera</i> (up to 10 feet tall). Considered "sensitive."
Succulent Shrub	The slopes of the Hackberry and Woods Mountains support an extensive stand of succulent shrubs, including barrel cactus, silver cholla, buckhorn cholla, hedgehog, pin-cushion, beavertail, and prickly pear cacti. Considered "sensitive."
Seeps and Springs	The relatively scarce occurrence of seeps and springs in the California Desert has resulted in all vegetation associated with seeps and springs being designated as Unusual Plant Asemblages. Considered "highly sensitive."
Riparian	Piute Creek, the only perennial stream in the East Mojave, supports a lush stand of cottonwoods, willows, and other streambank vegetation. Considered "highly sensitive."
Mesquite Thickets	Mesquite thickets indicate the presence of a high water table and occur in substantial numbers in the Scenic Area near Crucero. Considered to be "subject to disturbance, but not sensitive."

and the Mojave yucca, *Yucca schidigera*, with trunks up to several feet tall. Mixed in with the yucca are Utah or California juniper (*Juniperus* spp.). Desert thorn (*Lycium andersonii*), blackbrush (*Coleogyne ramosissima*), Mojave horsebrush (*Tetradymia stenolepsis*), California buckwheat (*Eriogonum fasciculatum*), Great Basin sagebrush (*Artemisia tridentata*), antelope brush (*Purshia glandulosa*), pygmy cedar (*Peucephyllum schottii*), bladder sage (*Salazaria mexicana*), spiny menodora (*Menodora spinescens*), and Mormon tea (*Ephedra nevadensis*). Most of the vertebrates found in the Creosote Bush Scrub community extend into the Joshua Tree Woodland. In addition there are several specific species that utilize this community's resources. The desert night lizard (*Xantusia vigilis*), the desert spiny lizard (*Sceloporus magister*), and the night snake (*Hypsiglena torquata*) use the yucca for shelter and eat associated insects and lizards. The desert woodrat (*Neotoma lepida*) often constructs nests among the trunks of the Joshua trees. Scott's oriole (*Icterus parisorum*) is found in the community along with the ladder-backed woodpecker (*Picoidea scalaris*), common flicker (*Colaptes auratus*), and ash-throated flycatcher (*Myiarchus cinerascens*). Yucca moth (*Tegitigula* spp.) species pollinate the flowers of particular species of yucca. The seed pods and fruit produce the food for the developing larvae. After pupating, the adult moths complete the cycle. This is a classic example of mutualistic symbiosis.

Pinyon-Juniper Woodland

This community in the East Mojave Desert occurs at elevations of between 4500 and 7000 feet (1400 - 2100 meters). The high elevation of this community results in an annual rainfall of 15-20 inches. Pinyon-Juniper Woodland is not a desert community, but rather more mesic elements that are surrounded by desert. *Pinus monophylla*, the one-leaf pinyon pine, is the typical species of pine. Either of two species of juniper, the California juniper (*Juniperus californica*) or the Utah juniper (*Juniperus osteosperma*), are present depending on location. Mid Hills, one of the overnight stops of this trip, is dominated by Utah juniper. Understory plants include Great Basin sagebrush and other shrubs. Most of the reptiles, birds and mammals that are found in this community are the wide-ranging forms mentioned above. Several species are closely associated with the Pinyon-Juniper Woodland: the pinyon mouse (*Peromyscus truei*) feed on juniper berries; the pinyon jay (*Gymnorhinus cyanocephalus*) on pinyon nuts.

Cactus Scrub

At lower elevations a high desert plant association termed the "devil's garden" or Cactus Scrub is characterized by cacti and yucca. Among the cacti are buckhorn cholla (*Opuntia acanthocarpa*), silver/golden cholla (*Opuntia echinocarpa*), pencil cholla (*Opuntia ramosissima*), beavertail cactus (*Opuntia basilaris*), old man cactus (*Opuntia erinacea* var. *utahensis*), Mojave prickly pear (*Opuntia phaeacanthra*), barrel cactus (*Ferocactus cylindricus*), hedgehog cactus (*Echinocereus engelmanni*), and nipple cactus (*Mammillaria tetrancistra*). The vertebrates of this association are those of the Joshua Tree Woodland. The cactus wren (*Campylorhynchus brunneicapillus*) is generally abundant here because its preferred nesting site is in chollas.

Kelso Dunes

At first appearance, the dunes appear to be a hot and very dry place. However, sand can store water at depths of a few

inches to many feet for long periods. Because of this, over 100 species of plants live on or near the dunes. Common perennial species are creosote bush, salt bush (*Atriplex* sp.), desert dicoria (*Dicoria canescens*), plicate tiqulia (*Tiqulia plicata*), sand verbena (*Abronia villosa*), desert lily (*Hesperocallis undulata*), and galleta grass (*Pleucaphis rigida*). Annual species can be numerous, but their appearance and number are dependent on the amount and timing of the last winter rains. Common annuals include primrose (*Oenothera deltoides*) and lax flower (*Baelia pauciradiata*). The dunes themselves are not very hospitable to vertebrates, but the area immediately surrounding the dunes is the home to several species of reptiles and mammals. The Mojave fringe-toed lizard (*Uma scoparia*), zebra-tailed lizard, western whip-tailed lizard, desert iguana, desert horned lizard (*Phrynosoma platyrhinos*), sidewinder (*Crotalus cerastes*), shovel-nosed snake (*Chionactis occipitalis*), glossy snake, desert kangaroo rat, antelope ground squirrel, black-tailed hare, kit fox, and coyote are members of the dune community. Two species, the Mojave fringe-toed lizard and the sidewinder, are sand-dwelling specialists. These two species may be found in the more central parts of dune fields. A number of insects, including several endemic species (Kelso Dunes giant sand treader, *Macrobaenetes kelsoensis*, a camel cricket; Kelso Dunes Jerusalem cricket, *Ammopelmatus kelsoensis*, and the Kelso Dunes shieldback katydid, *Eremopedes kelsoensis*) occur on the dunes, providing a source of food for the lizards.

Geographic Areas

Granite Mountains

The southernmost mountains in the East Mojave Desert, the Granites, are noted for their unusually diverse range of plants (over 400 species) and animals, including 120 bird species, 40 mammals, and 33 reptiles. A 6,720 acre area has been designated a Research Natural Area, jointly managed by the National Park Service and the University of California.

Providence Mountains

The Providence Mountains are one of the most scenic mountain ranges in the East Mojave Desert. Limestone cliffs form the sheer north face of the North Providence Mountains, with volcanic flows providing a striking contrast. Dense stands of cacti and desert shrubs dot the bajadas. These mountains contain rich mineral deposits. The Vulcan Mine in Foshay Pass was a major iron ore producer during World War II. Mitchell Caverns, a National Natural Landmark, is administered by the State of California as a State Recreation Area. Many "wild caves" exist in the limestone areas.

Devil's Playground

Sand from Soda Dry Lake deposited by the Mojave River has been blown southeast by prevailing winds to cover a large expanse of land. The Cowhole, Old Dad, Marl and Kelso Mountains are the home to one of the highest concentrations of Desert Bighorn Sheep (*Ovis canadensis nelsoni*) in the desert. Current estimates place the population at over 3700 individuals in the East Mojave. The region contains important archaeological sites and excellent wildlife habitat.

Kelso Dunes

A National Natural Landmark, sands at the southeastern

edge of Devil's Playground have formed the continent's third highest dune system (940 meters). The area has been closed to vehicle use since 1973 to protect vegetation and animals but is still open to grazing and mining.

Mid Hills

Mid Hills, north of the Providence Mountains, are a series of exposed granite crags and boulders. To the south, volcanic hills form the backdrop to Hole-in-the-Wall campground. Mid Hills, at an elevation of 5600 feet (1700 meters), sits within the Pinyon-Juniper Woodland community. Dense stands of sagebrush and other high desert species reflect the nearby Great Basin Desert. To the south, the Woods and Hackberry Mountains share many characteristics with Mid Hills. Three distinctly-shaped volcanic mountains dominate the area. Table Mountain is a dominant landmark. The Woods Mountains are steep-faced mountains with many canyons noted for several "cactus gardens." To the east, the Hackberry Mountains are noted for significant fossil finds (Reynolds and others, this volume).

Cima Dome

Covered by one of the desert's finest Joshua Tree Woodland, the 70 square mile, gently rounded granite dome is a dominant physical structure in the East Mojave Desert. To the west of Cima Dome several layers of ancient lava punctuated by some 30 extinct volcanic cones are noted for their petroglyph-covered basalt and geologic features.

New York Mountains and Castle Peaks

The central high elevation "spine" of the East Mojave is located in the New York Mountains. Elevations from 4600 to over 7200 feet (1400-2200 meters) characterize this range. To the west, the New Yorks have a steep, sharply eroded north face. A more gently sloping southern flank has carved out long canyons best known for their chaparral and live oak plant communities. Also notable is a small grove of white fir (*Abies concolor*) to the west of Caruthers Canyon. These firs and the chaparral species in Caruthers Canyon are relic populations stranded by Holocene climate changes. To the east, the New York Mountains are dominated by the reddish spires of Castle Peaks.

Ivanpah Valley

This creosote scrub covered bajada, located between the Ivanpah and New York Mountains, contains crucial desert tortoise habitat. A permanent desert tortoise study plot has been located here to assess possible long range effects on tortoise from cattle grazing.

Mescal Range

The Ivanpah Mountains and Mescal Range lie south of Mountain Pass on Interstate 15, and comprise a highly mineralized area. Near Mescal Range is the 480 acre Dinosaur Trackway Area of Critical Environmental Concern, the only known dinosaur footprints in California.

Piute Range

This is a small mountain range located on the eastern edge of the Mojave Desert. It contains the only perennial stream in the East Mojave. The presence of permanent water has created an oasis for plant and animal species. Substantial progress has been made in improving the riparian zone through the removal of tamarisk trees (*Tamarix* sp.) and the planting of cottonwoods (*Populus fremonti*). Plans to transplant the elf owl (*Micrathene whitneyi*) into this area are proceeding.

Fenner Valley

This area in the southeast corner of the East Mojave is the northern portion of the larger Fenner-Chemehuevi Valley crucial desert tortoise habitat. This area also contains a portion of Camp Clipper, one of eleven desert military training camps established by General Geroget Patton to train troops for the North Africa campaign.

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Kelso Dunes

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Kelso Dunes covers an area of approximately 100 km² at an elevation of 500 - 900 m on the piedmont alluvial deposits of the Granite and Providence Mountains (Lancaster, 1993; Sharp, 1966). It forms the depositional sink for a well-defined eolian sediment transport system that extends from the fan-delta of the Mojave River as it exits the Afton Canyon eastwards to the piedmont of the Providence Mountains (Fig. 1).

Dune Morphology

The dune field consists of a 40 km² area of active dunes surrounded on their west, north, and east sides by areas of lower vegetation-stabilized dunes (Fig. 2).

The "core" of the Kelso dune field consists of 3 large WSW-ENE trending complex linear ridges up to 160 m high and 1900-2000 m apart (Units VI, VII, VII, X). The linear

ridges consist of a series of coalesced star dunes. Active crescentic dunes 3 - 8 m high with a spacing of 100 - 150 m, aligned transverse to WNW - W winds are superimposed on their flanks. Linear ridges similar to those at Kelso also occur in other dune fields in the Mojave (e.g. at Dumont and Eureka). On the west and northwest sides of the core of the dune field there are four areas of degraded, vegetated crescentic and linear dunes. The crescentic dunes are of two varieties: straight N - S oriented broad-crested simple transverse dunes with a height of 5 - 8 m and a spacing of 180-190 m (Units XII, XIII); and 5 - 15 m high degraded crescentic dunes with a spacing of 180 m to 250 m and strongly concave lee faces oriented to the SE (Units IX, XIV). These crescentic dunes overlie linear ridges on WSW-ENE trends, creating a complex network of intersecting hollows and ridges. On the southwest side of the dune field there are

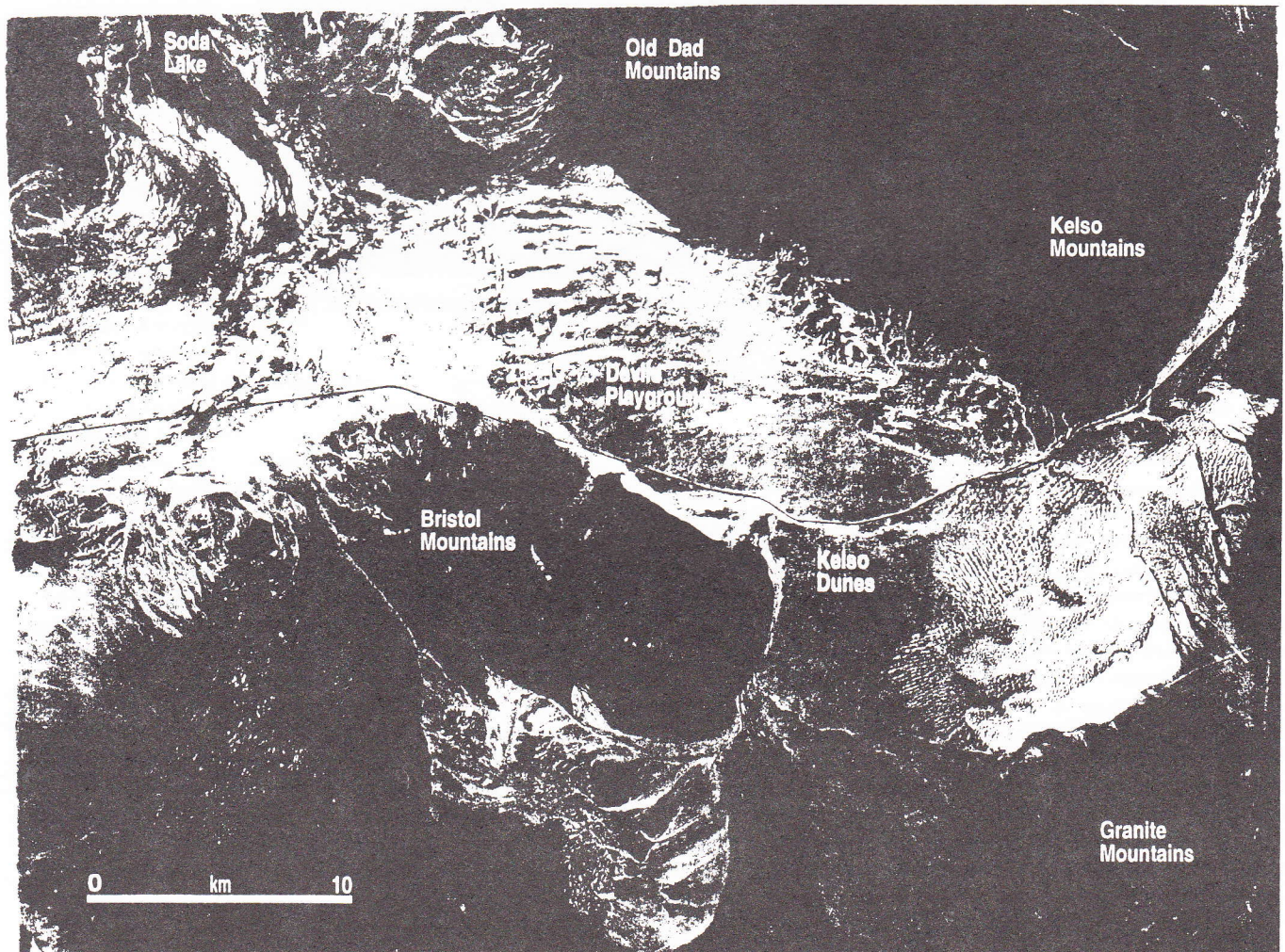


Figure 1. Landsat TM image of the Kelso Dunes eolian sediment transport system. Note active dunes to west of hills in the western Devils Playground, yet no active sand transport to Kelso dune field.

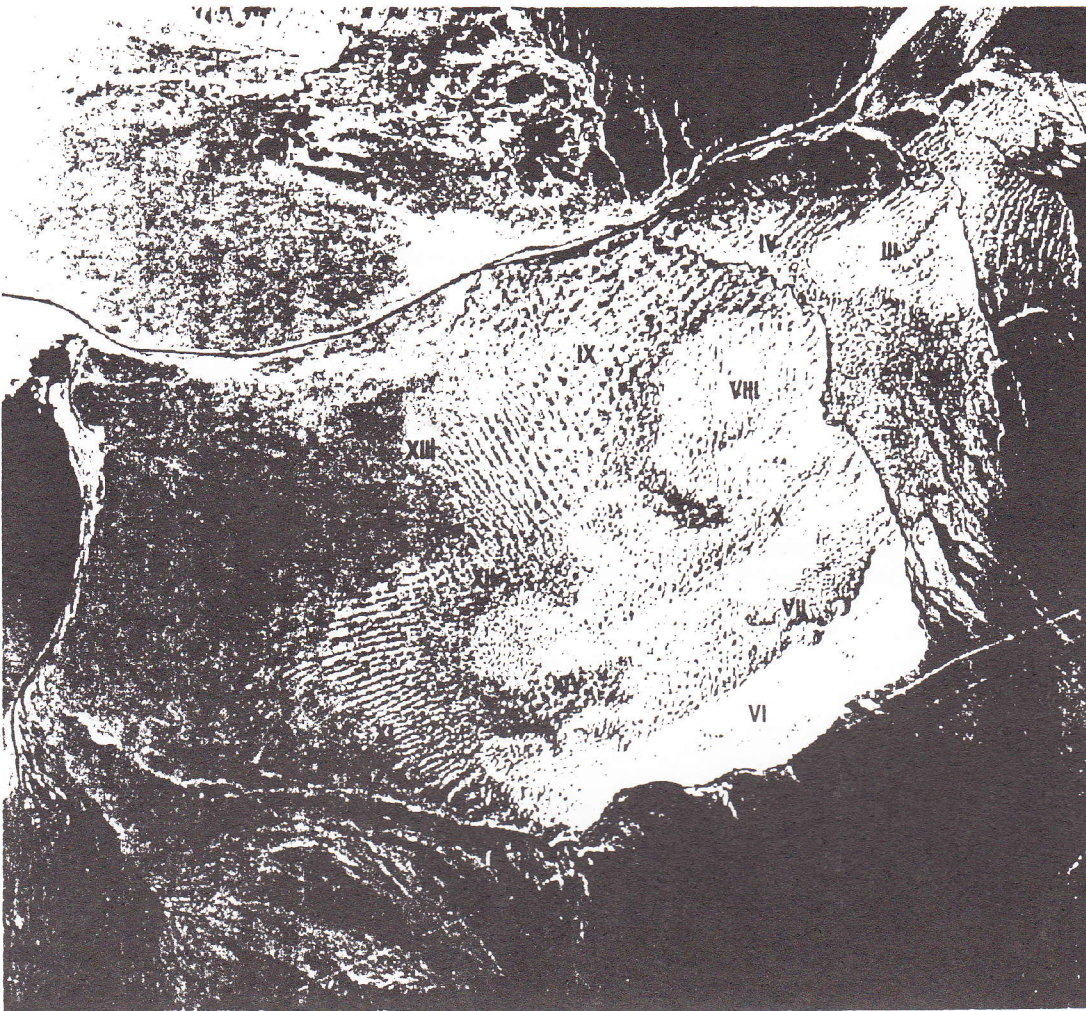


Figure 2. Landsat TM image of Kelso dunefield showing areas of different dune morphology and sediments.

areas of 2 -3 km long very degraded linear dunes, 3 -5 m high and spaced 150 m apart (Unit XI).

The eastern section of the dune field consists of 1 -5 m thickness of sand formed into of five smaller areas of dunes, each cut by washes and separated from the "core" area by Cottonwood Wash, which is incised into eolian and fluvial deposits by as much as 20 m. To the north is a smaller version of the main area of active dunes that consists of three linear ridges up to 50 m high with superimposed 2-4 m high crescentic dunes (Unit III). Areas of low, partially active, linear and crescentic ridges, as well as extensive areas of sand sheets occur on the northeast and southern edges of this part the dune field (Units I, IV, and V). East of Winston Wash are areas of vegetated and degraded SSW - NNE trending crescentic ridges with a height of 8 - 10 m and a spacing of 180 - 200 m (Unit II). Smaller ridges cross the interdune areas on alignments roughly perpendicular to the main ridges, giving rise to a network dune pattern.

Dune Sands

Dune sands at Kelso are composed of 50 - 80 % quartz, as much as 50% plagioclase and potassium feldspar, and up to 1 - 2 % granite fragments in places. Heavy minerals are

dominated by amphibole and magnetite. The magnetite is concentrated in the very fine sand fraction and forms prominent deposits on dune crests in the main active dune area, where it may form as much as 30% of the whole sample. The highest percentages of granite fragments are found along the northern edge of the dune field, close to Kelso Wash, suggesting that dune sands here were derived from this source. The highest percentages of amphibole occur along the eastern edge of the dunes, as well as in small areas to the west. Alluvial sediments of streams draining the Granite and southern Providence Mountains contain significant

quantities of mag-netite and amphi-bole (Yeend et al., 1984) and are likely the source of these components of dune sediments. The total volume of sand present at Kelso (at least 1 km³) suggests however that the major source of sand for Kelso was the Mojave River sink and the Soda Lake basin.

Dunes at Kelso are composed of sands that range in texture from coarse to fine sand, and sorting values that vary from very well sorted to poorly sorted. Three distinct sand populations (Fig. 3) can be distinguished on the basis of their particle size distribution and mean grain size and sorting parameters: (1) moderately-well to moderately sorted relatively coarse sands (s_1 [phi standard deviation] = 0.5 - 1.0) with mean grain sizes ranging from 1.5 to 2.0 phi in the northern and south central areas; (2) moderately well sorted (s_1 = 0.5 - 0.75) sands that are somewhat finer (M_z = 2.0 - 2.4 phi) on the eastern and western margins; and (3) very well to well sorted (s_1 = 0.25 - 0.5) and fine (M_z = 2.0 - 2.6 phi) sand which is restricted to the active dune ridges of the central parts of the sand sea.

Development of the Dune Field

A characteristic feature of Kelso Dunes is the juxtaposition of areas of dunes of distinctly different morphological type,

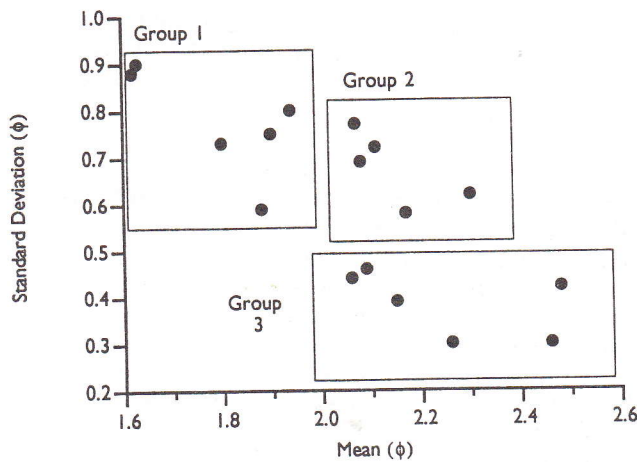


Figure 3. Grain size and sorting characteristics of different sand populations at Kelso dunes.

size and spacing and alignment. Spatial variations in dune types and composition in many sand seas have been regarded as the product of regional variations in wind regimes and sediment sources (e.g. Wasson and Hyde, 1983), but the close juxtaposition of dunes of contrasting morphology, sediments, and activity cannot be explained in these terms (Lancaster, 1992). In many parts of Kelso Dunes, geomorphic and sedimentary relations between different dune morphological units indicate that they are superimposed on one another, and the dune field represents in part a stacked or shingled sequence of dunes of different generations. "Stacking" of dune generations has occurred because the lateral expansion of the dune field in response to sediment inputs is restricted by its location close to the mountain front of the Granite Mountains.

Estimates of the age of Kelso Dunes have varied widely between "several thousand years and possibly 10,000-20,000 [years]" (Sharp, 1966) to "very likely greater than 100,000 years, and quite possibly more than a million years" (Yeend et al., 1984, Smith, 1984). Because the Kelso dune field overlies, and in some cases intercalates with, alluvial fan

deposits derived from the Granite and Providence Mountains, it is possible to use the ages of these alluvial fans (Wells et al., 1990) to constrain the age of the dunes. It appears that the western part of the dune field rests on fans of Early or possibly Middle Pleistocene age. Late Pleistocene fan units underlie the southern margins of this part of the dune field. A well-developed paleosol is developed on eolian sand that lies on the Granite Mountains piedmont alluvial fans south of the main dunefield. This soil appears similar to those developed on sand ramps elsewhere in the region with an age of more than 20,000 yr.

Dunes east of Cottonwood Wash appear to be much younger and mostly lie on fan surfaces of Holocene and latest Pleistocene age (Clarke, 1994; McDonald and McFadden, 1994). Unit Qe1 of McDonald and McFadden (1994) consists of thin discontinuous sand sheets that overlie alluvial fan unit Qf4 which is of Late Pleistocene age. Eolian Unit Qe2 includes much of the eastern part of the dunefield and overlies fan unit Qf5 (early Holocene to latest Pleistocene). In turn, Qe2 is truncated by the late Holocene Qf6 unit.

Luminescence Dating Studies

Luminescence dating of dune and sand sheet units (Edwards, 1993; Wintle et al., 1994) provides minimum age estimates for some parts of the dune field. IRSL ages obtained so far indicate that there is an exponential increase in the age of dune and sand sheet units from northwest to southeast, supporting the hypothesis that the dune field accumulated by stacking or shingling of successive generations of eolian units on the piedmont of the Providence and Granite Mountains. The oldest sands known are those which were deposited as sand sheets on alluvial

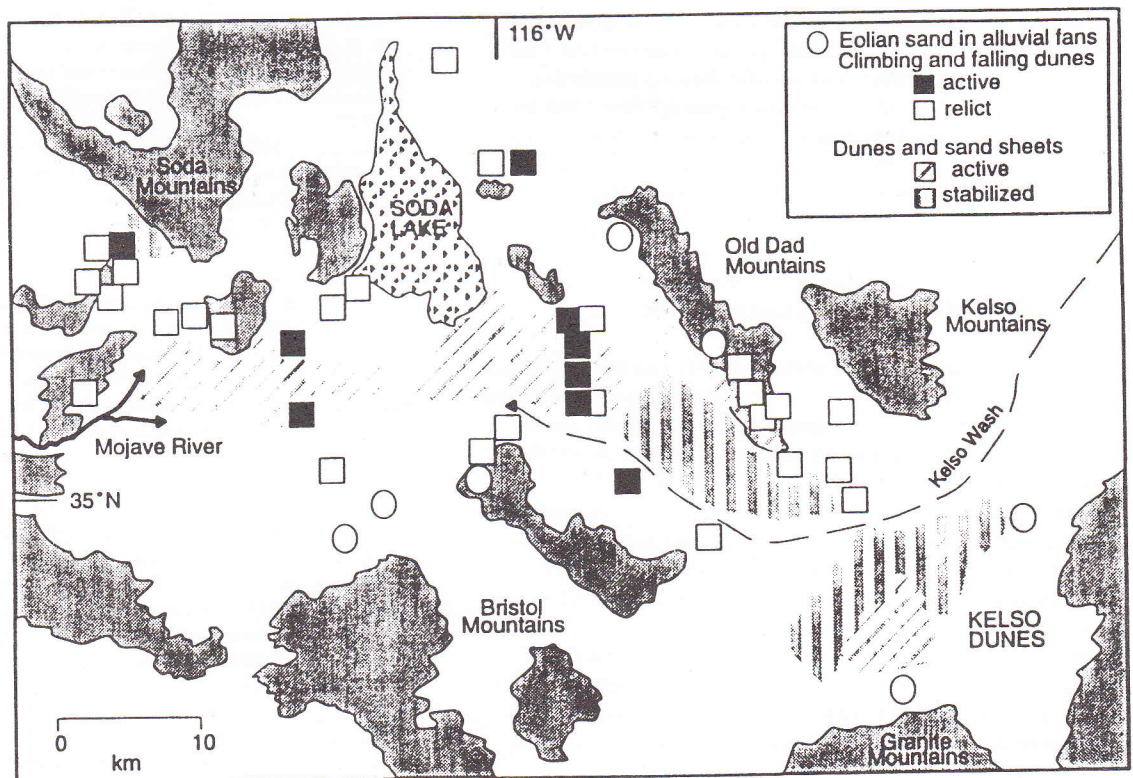


Figure 4. Distribution of active and relict eolian landforms in the Kelso Dunes eolian sediment transport system.

fan surfaces as much as 5 km southeast of the present dune margins between $16,830 \pm 1465$ and $17,300 \pm 1935$ yr. ago (Clarke, 1994). Dunes on the eastern margin of the dune field (Qe2) were accumulating 8,420 - 10,410 years ago, but were stabilized $3,500 \pm 220$ to $3,700 \pm 425$ yr. B.P. in conditions of increased regional rainfall (Wintle et al., 1994). Later periods of dune formation and/or reactivation occurred on the northern and western margins of the dune field around 1,500 yr. and 400 - 800 yr. ago (Wintle et al., 1994).

Discussion

Currently, the Kelso dune field is receiving no sediment from its primary source – the Mojave River fan delta. Active transport of sand is restricted to the Mojave River sink and the western parts of the Devil's Playground and no sand reaches the Kelso dune field from these sources. Relict sand sheets stabilized by gravel lag surfaces and soil development occur in the eastern part of the Devils Playground. Sand ramps that are stabilized by talus mantle the slopes of the Bristol Mountains southwest of Balch, as well as the southern extension of the Old Dad Mountains (Fig. 4).

Stratigraphic and geomorphic studies, together with luminescence dating have established that five main periods of enhanced eolian activity have occurred in this eolian sediment transport system in the past 25,000 yr. (Clarke et al., in press; Rendell, in press). The extent and character of these periods has varied through time. Phase I (25,000 to 16,800 yr. ago) was characterized by the formation of now relict sand ramps at several localities. For example, eolian sand in the Balch sand ramp and the Cronese Mountains (e.g., the Cat Dune) accumulated between about 24,500 and 17,000 yr. ago. This period was coeval with sand sheet accumulation on the fans of the Providence Mountains (Unit Qe1). Phase II (12,500 to 3,500 yr. ago) involved renewed sand ramp accumulation in some localities, including both Balch (7580 ± 1020 to 4270 ± 580 yr. ago) and the Old Dad Mountains where major periods of eolian accumulation occurred around 11,000 yr. ago and from $6,710 \pm 1,110$ to $4,320 \pm 350$ yr. ago. This phase correlates with the accumulation of the dunes and sand sheets on the eastern side of the Kelso dunefield (Unit Qe2). Late Holocene periods of eolian activity (Phases III to V) occurred 3,000 to 1,500, 800 to 400, and 150 to 250 years ago. They involved formation of small dunes in the Cronese basin and local accumulation of sand at Balch and in the Old Dad Mountains.

Some of these periods of eolian activity can be correlated with other sources of paleoclimatic data in the area. The first phase of eolian activity occurred both prior to and coeval with lake highstands (Lake Mojave I: 20,900 to 16,900 B.P.; Lake Mannix 23,500 to 20,800 and 17.6 to 16,500 B.P., calibrated ^{14}C ages) (Clarke et al., in press). A period of much more mesic climates (Spaulding, 1990) and high lake levels and river discharge (Lake Mojave II, 16,500 to 13,400 B.P.) (Brown et al., 1990) was associated with eolian stability and soil formation on many sand ramps in the region. The major part of Phase II of eolian activity was associated with desiccation of Lake Mojave II after 13,400 B.P. with final desiccation at 9,700 B.P. (Brown et al., 1990), but dunes remained active as a result of mid Holocene aridity in the region (Spaulding, 1991). Late Holocene periods of dune formation and/or reactivation reflect both periods of

regional drought (Fritts and Gordon, 1982; Spaulding et al., 1994), as well as enhanced sediment supply from temporary lakes in the Silver Lake - Soda Lake and Cronese Basins at around $3,620 \pm 70$ and 390 ± 90 yr. B.P. (Enzel et al., 1992).

Conclusions

Kelso Dunes is a very good example of a dune field that has developed in response to Quaternary climatic changes in the region. Sedimentary and geomorphic relationships between the multiple areas of different dunes and sand sheet suggests that they have accumulated as a series of small genetically independent dune fields formed during periods of enhanced sediment supply from the primary source area. In the late Pleistocene, this was associated with desiccation of paleolakes in the Silver Lake - Soda Lake basin and is well demonstrated during the early to mid Holocene when much of the eastern part of Kelso dunes probably formed.

Acknowledgements

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A Piedmont Landscape in the Eastern Mojave Desert: Examples of Linkages Between Biotic and Physical Components

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Introduction

Soil variation exerts a major influence on ecological patterns in the deserts of the American Southwest. Over the last 40 years, various ecological studies in the Mojave, Sonoran, and Chihuahuan Deserts have attempted to describe ways in which vegetation and other biotic components correspond to variation in the underlying substrate. Most of these studies have been conducted in the broad alluvial piedmonts that flank the mountains of these desert regions. Understanding ecological patterns in these desert landscapes is greatly enhanced through the incorporation of knowledge about processes of landscape evolution and soil development (McAuliffe 1994, in press).

The conceptual approach typically taken in ecological studies has been to consider various landscape and soil features as static conditions to which plants and animals respond. However, organisms can also contribute to changes in soil conditions and landscape features. The potential effects of organisms on changes in the soil environment were explicitly incorporated by Jenny (1941) in his classic formulation of the five factors of soil formation (time, topography, parent materials, climate, and organisms). Although the impacts of organisms on soil characteristics are typically pronounced and well-recognized for more humid areas (e.g., inputs of organic matter in soils of prairies and forests) considerably less attention has been given to influences of organisms on desert soils. The purpose of this paper is to discuss some of the interactions between physical and biological landscape components of the alluvial piedmont flanking the west side of the Providence Mountains in the eastern Mojave Desert.

An Integrated Landscape System Physical and Biotic Components

Any part of the piedmont landscape may be pictured as a system in which component variables are connected by various cause-and-effect relationships (Bull 1991). Climate and lithology are *independent* variables in that they are not influenced by processes occurring in the piedmont environment (Fig. 1). Various *dependent* variables are influenced by the independent variables and other dependent variables. These dependent variables can be further divided into physical components (landforms, soils) and biotic components (plants, animals). Influences of these dependent variables on each other are depicted by the arrows connecting particular pairs of variables (Fig. 1). The interactions of independent variables may produce either direct or indirect feedback loops. For example, in Figure 1, a direct

feedback loop is indicated by a pair of arrows (2 and 3) where soil properties influence animal activity (arrow 2) and the animal activity in turn affects soil conditions (arrow 3). An indirect feedback loop is indicated by the set of arrows 1, 4, and 3, where soil characteristics affect vegetation (arrow 1), vegetation affects animal activity (arrow 4), and animal activity ultimately has an impact on soils (arrow 3).

The Providence Mountains Piedmont Physical Landscape Components

Study of the interaction of dependent variables within a landscape system requires a solid foundation of knowledge about how different independent variables (climate, lithology) have affected landform evolution and soil development. Soil-geomorphic research by McDonald (1994) and McDonald et al. (this volume) on the west-facing piedmont of the Providence Mountains provides this basic foundation to which other components of a landscape system model can be added. Climatic variation has apparently controlled the timing of aggradation of alluvial fans, producing a mosaic of different surface deposits, ranging in age from latest Holocene to middle Pleistocene. The Providence Mountains are the ultimate source for alluvium deposited on the piedmont and variation in rock type within the mountain range has yielded fans with different lithologic compositions. Much of the variation in soil characteristics is explained by ages and lithologic

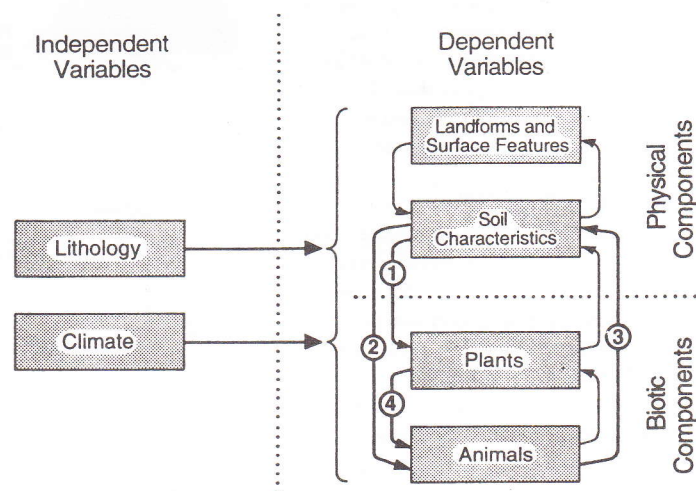


Figure 1. Landscape system model as described in text. Arrows between components indicate direction of cause-and-effect linkages. Only some of the potential linkages are diagrammed. The four numbered linkages (1-4) are the ones specifically discussed in the text.

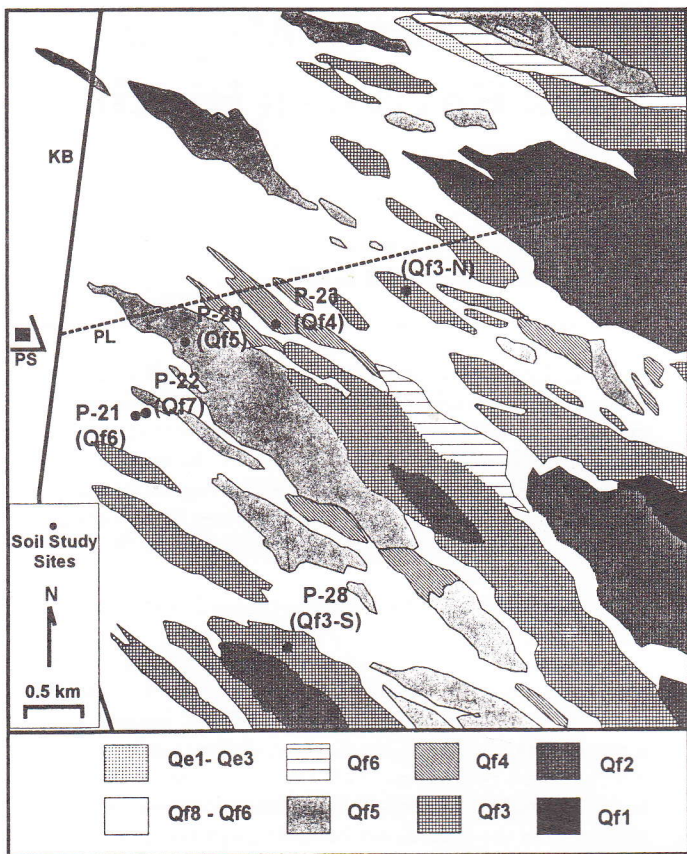


Figure 2. Location of study areas on the mixed plutonic alluvial fan deposits south of the Powerline Road (PL). Shown also are the natural gas pumping station (PS) to the west side of Kelbaker Road (KB). Surface designations Qf7, Qf6, etc. are those referred to in text. The site names P-10, P-21, etc., are the names of soil profiles taken at the same sites by McDonald (1994).

stratigraphic relationships of these deposits]. For this field trip, a series of four surfaces (Qf3-Qf6) on the mixed plutonic parent materials to the south of the Power Line Road (Fig. 2) and one eolian deposit (Qf2) located near field trip stop 2 provide examples of the linkages between biotic and physical landscape elements.

Mixed plutonic fan sequence

Variation in vegetation and faunal activity on fan deposits of different age provides information on the following relationships among various dependent landscape variables (numbers correspond to linkages diagrammed in Figure 1): (1) effects of soil conditions on plant responses, (2) effects of soil conditions on animal (rodent) activity (3) feedback involving ways in which rodent activity further affects soil conditions and (4) how these effects of animal activity on soils influence plant response. These different linkages are referred to by the numbers presented for different specific relationships diagrammed in Figure 1.

compositions of the fan deposits. Increasingly strong soil horizons develop over time; development of some horizons, for example, clay-enriched B horizons and calcic horizons, also varies according to lithology (see McDonald et al., this volume).

Biotic Landscape Components

In 1993-1994, vegetation and animal activity were studied at 22 locations where McDonald (1994) had also conducted detailed studies of soils. These ecological study areas were distributed among fan deposits of three different parent materials (limestone along the road to Cornfield Springs directly east of Kelso, mixed volcanic rocks along Vulcan Mine Road, and mixed plutonic rocks along the access road paralleling the transmission line corridor, referred to as Powerline Road). Within each lithology, an entire sequence of different-aged deposits (Qf2-Qf7) was studied. In addition, data were also collected from three eolian deposits (Qe1-Qe3). [See Fig. 2 in McDonald, et al. (this volume) for ages and

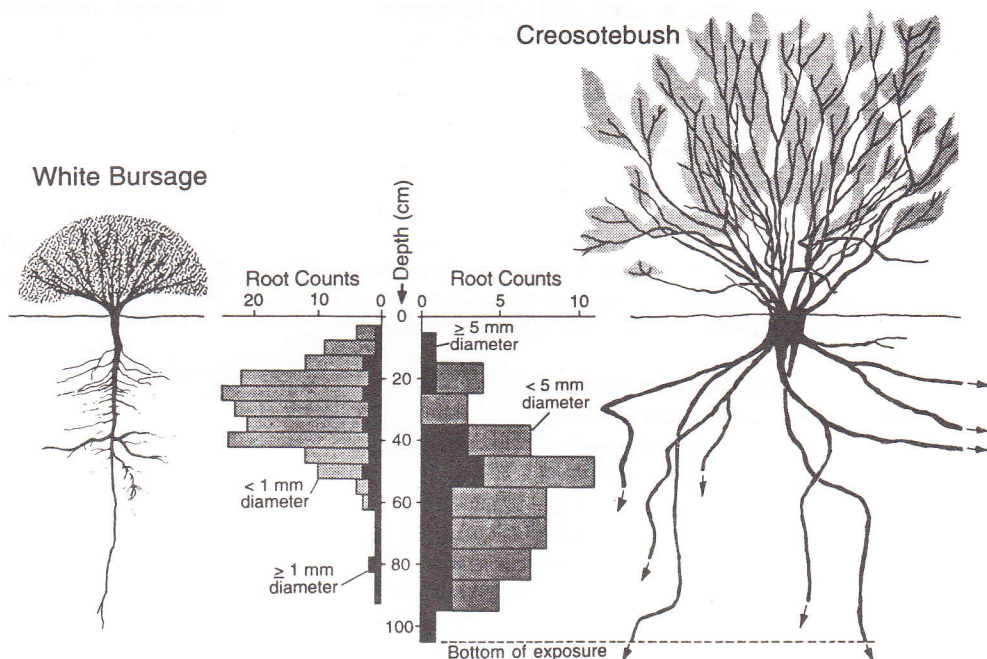


Figure 3. Contrasting vertical root distributions of white bursage and creosotebush. Data were taken from a fresh, vertical exposure of a Qf6 surface. Root counts are the number of roots intersecting horizontal lines spaced at 5 cm intervals for white bursage and 10 cm intervals for creosotebush. The illustration of rooting pattern for creosotebush does not include fine roots. The arrows at the ends of diagrammed roots indicate that these roots were not fully exposed and the lengths of these roots extended an unknown further distance.

Linkage 1: Plant responses to varying soil conditions. In non-saline soil environments in lower elevations of the Mojave Desert, creosotebush (*Larrea tridentata*) and white bursage (*Ambrosia dumosa*) are nearly ubiquitous. However, growth responses and relative abundances of these two shrub species change in response to variations in soil conditions. These two species do not necessarily exhibit the same responses. Soil situations which may be favorable for one species may be unfavorable to the other. The contrasting responses are due to the shrubs' different life forms and modes of water acquisition and use. Creosotebush is a perennial evergreen whereas white bursage is a drought deciduous shrub. White bursage generally has a relatively shallow root system compared to the much more extensive system of both deep and shallow roots of creosotebush (Fig. 3). These differences in above- and below-ground features reflect contrasting ways of surviving extended dry periods. White bursage can rapidly produce new leaves and become photosynthetically active in response to relatively shallow wetting of the soil. However, shallow soil layers are also the most seasonally variable in water content due to rapid use by plants and direct losses by evaporation to the atmosphere. Drought dormancy is an adaptation that enables white bursage to survive during periods when plant-available moisture is completely lacking at relatively shallow soil levels.

Creosotebush, on the other hand, lacks the capacity for drought-dormancy. Extreme drought that leads to loss of leaves typically causes death of whole branches (Runyon 1934). The extensive (deeper and more laterally spreading) root system of creosotebush allows this shrub species to obtain water from a considerably larger and deeper soil volume (Fig. 3). Although precipitation events large enough to wet deeper soil layers occur far less frequently in this desert environment than do smaller events, deeper soil layers exhibit less seasonal fluctuation in the amount of plant-available water (Noy-Meir 1973; Monson and Smith 1982). Creosotebush remains photosynthetically active during dry seasons by extracting the limited, but more seasonally constant supplies of water from deeper soil layers. Only in extremely unusual and long-duration droughts (like the record-breaking 1987-1991 drought period in the Mojave Desert) do creosotebush plants exhibit extreme leaf loss and considerable mortality of entire branches.

Some of the soil horizons discussed in McDonald et al. (this volume) greatly affect the quantity and depth of water

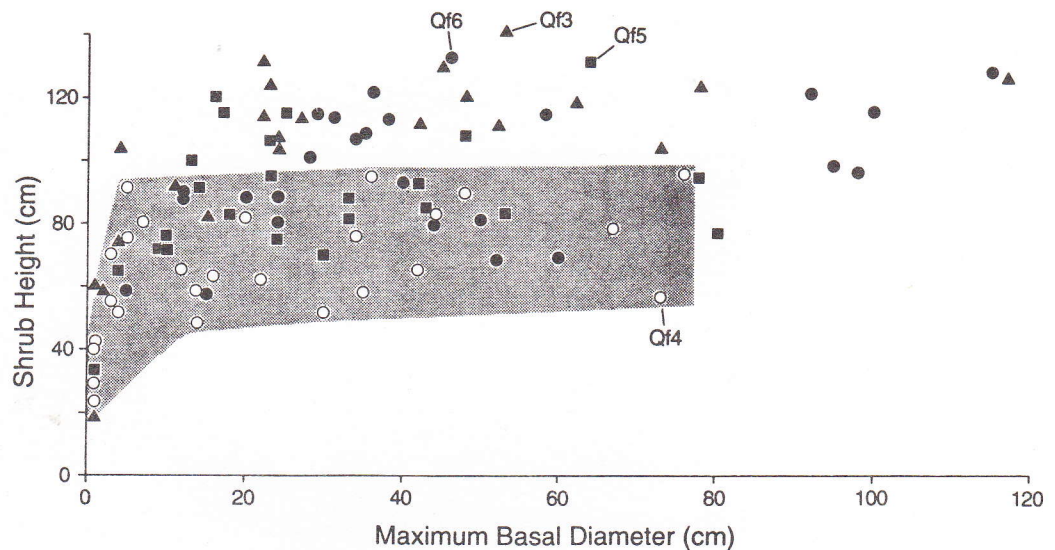


Figure 4. Maximum height as a function of basal diameter of creosotebush from four alluvial surfaces (Qf3, Qf4, Qf5, and Qf6). The shaded area indicates the envelope containing all the creosotebush plants from the Qf4 surface. Symbols refer to individual creosotebush plants with their respective surface designated by symbol type.

infiltration. In fan deposits of mixed plutonic parent materials, fine-textured vesicular (Av) horizons found directly beneath desert pavements have a high water-holding capacity and impede infiltration of precipitation (see Fig. 7 in McDonald et al., this volume). Similarly, clay-rich argillic (Bt) horizons generally found at depths of approximately 20 cm and greater on Pleistocene-aged surfaces in the study area have high water-holding capacities that further inhibit infiltration to greater depths. Both soil horizons generally become progressively more well developed with soil age (see data on Av horizons in Fig. 3B in McDonald, et al., this volume).

Creosotebush exhibits different growth responses on surfaces of different ages that correspond to strength of development of Av and Bt horizons. Creosotebush plants on the Qf4 surface never achieve the height that they do on the Qf6, Qf5, and Qf3 surfaces (Fig. 4). Soils on the Qf4 surface are extensively covered with well-developed desert pavements, underlying Av horizons, and Bt horizons. Pavements, Av, Bt horizons are either lacking or are less well developed on the younger (Qf5, Qf6) surfaces, or have been significantly truncated by surface erosion on the older Qf3 surface. In either case, deeper infiltration of precipitation is possible on the younger soils or on those where Av horizons have been at least partially removed.

Individual creosotebush plants can live to great age as "clones" through a process of outward radial growth (Vasek 1980). A clone typically consists of a circular or oval ring of emergent stems; the basal diameters of these rings may exceed several meters (Fig. 5). At another site in the Mojave Desert, Sternberg (1976) demonstrated that the many clumps comprising such a ring were genetically identical. In other words, a circular clump is derived through outward growth over long periods of time of a creosotebush plant that germinated from a single seed. The outward growth of these rings is extremely slow, on the order of less than an average of 1 mm per year (Vasek 1980).

Creosotebush clones are common only in certain situations. First of all, such clones require long periods to develop and are therefore present only in geologically stable landscape positions. Sites having undergone recent alluvial deposition (such as the Qf8 surface) or rapid erosion lack creosotebush clones (McAuliffe 1991, 1994). The study sites of Qf7-Qf3 alluvial deposits pictured in Fig. 2 range in age from at least 1000-4000 years (Qf7) to over 100,000 years (Qf3), therefore, all of these surfaces satisfy the requirement of stability required for clonal development. However, creosotebush clones are not found on all of these stable surfaces. Large-diameter clones (measured as the maximum basal diameter) are common only on Holocene Qf7 and Qf6 surfaces, typically reaching basal diameters of two meters (Fig. 6). The largest clone found so far on these surfaces is 5.25 m X 4.90 m across at the base (Fig. 5). Creosotebush plants never exceed 80 cm basal diameter on the Pleistocene Qf5 and Qf4 surfaces. However, the Qf3 surface does support occasional clones, though not as frequently as found on the Qf6 and Qf7 surfaces (Fig. 6). The desert pavements and associated Av horizons together with clay-enriched argillic horizons of the Qf4 and Qf5 surfaces may be factors that impede growth and survival of individual creosotebush for extremely long durations, thus preventing clone development. Creosotebush growing on these kinds of soils often show a considerable accumulation of dead, variably

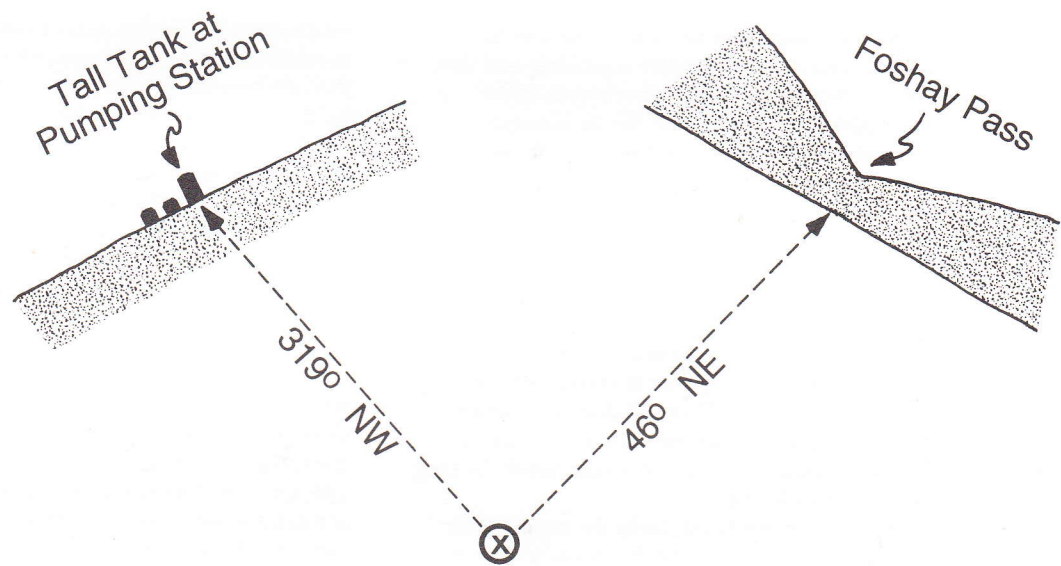
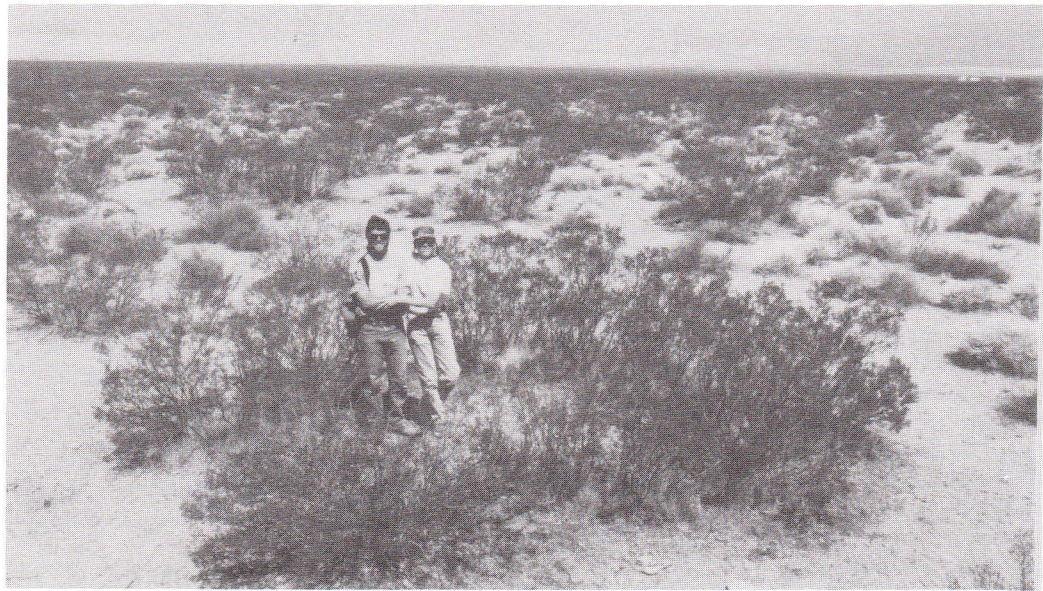


Figure 5. Upper - Claudia Luke gives Jim Andre a hug within the confines of the largest creosotebush clone yet located on the Qf6 and Qf7 surfaces. Claudia and Jim are the co-directors of the Granite Mountains Reserve. The basal diameters of this slightly elliptical clone (measured as diameter between outermost stems at the ground surface) were 5.25 m X 4.90 m in May 1994. This large clone (and dozens of smaller ones) are located about 200 m east of Kelbaker Road, 1.7 miles south of the intersection with the Powerline Road. View to the west; telephone poles along Kelbaker Road visible in background; natural gas pumping station on extreme right. Lower - Magnetic compass bearings to pinpoint the location of the giant clone (marked "X") pictured above. A creosotebush clone with the name "King Clone" already exists, so we affectionately name this pretty clone the "Providence Princess Clone".

weathered branches which indicates repeated bouts of stem mortality.

Strongly developed Av and argillic horizons are not the only soil conditions that impede development of creosotebush clones. For example, although creosotebush clones are common on the Qf6 surfaces of mixed plutonic

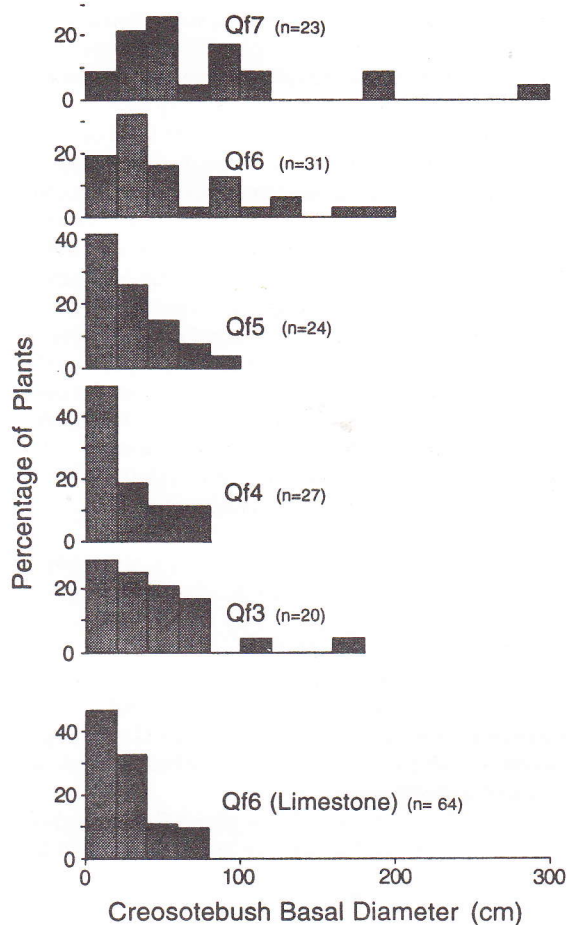


Figure 6. Distributions of creosotebush basal diameters among different alluvial surfaces. The value for "n" indicates the total number of plants sampled in a 500 m² sample area on each surface. Data for the Qf3 surface is from the Qf3-S site.

parent materials, they are absent from deposits of the same age in limestone parent materials (Fig. 6). These mid-late Holocene deposits of limestone alluvium are similar to the mid-late Holocene mixed plutonic deposits in their lack of an argillic horizon and weaker development of an Av horizon. The factor that may inhibit clonal development on the limestone fan deposits is the extreme coarseness of the alluvial parent materials. Marked outward propagation of a creosotebush may be physically possible only in relatively fine-grained (sandy to gravelly) parent materials such as those found in the Qf6 and Qf7 mixed plutonic deposits. The Qf6 limestone deposit, by contrast, is largely composed of coarse cobbly to bouldery materials which may inhibit radial propagation (Figs. 7, 8).

The inhibition of development of large creosotebush clones by extremely coarse deposits may also contribute to the lack of clones on the Qf4 and Qf5 mixed plutonic surfaces (Fig. 8). These Pleistocene surfaces are composed of coarse cobbly to bouldery alluvium. The Qf3 surface also was originally composed of coarse, high-energy deposits, but in this considerably older alluvium, the original clasts on the surface and shallow soil layers have been extremely weathered and fragmented, yielding a general absence of large, intact clasts. The further effects of this physical

weathering of the Qf3 surface on animal activity is discussed in the next section.

The various effects of different soil conditions on growth responses of individual plants translates to differences in vegetation composition among the different surfaces. Creosotebush attains the greatest values for canopy cover on the Holocene Qf6 and Qf7 surfaces and the lowest value on the Pleistocene Qf4 surface (Fig. 9). Since creosotebush inhabiting the Qf4 surface also are significantly shorter than those of the other surfaces (Fig. 4), the difference between the Qf4 and the other surfaces in total canopy volume or biomass of creosotebush is even more pronounced. White bursage, in contrast, does not exhibit the marked reduction in cover on the Qf4 surface (Fig. 9). The ability to rely exclusively on shallow, seasonally variable moisture supplies may allow white bursage to thrive in the soil environment of the Qf4 surface that is marginal for creosotebush.

Linkage 2: Effects of soil conditions on animal activity.

The abundance and activities of pocket mice (*Perognathus* spp. and *Chaetodipus* spp.), kangaroo rats (*Dipodomys* spp.), and other burrowing rodents vary considerably from place to place according to the particle sizes making up the various alluvial deposits. Rodent activity is highest in sites composed of relatively fine particles, including the Qf7, Qf6, and Qf3 surfaces of mixed plutonic parent materials. In contrast, the Qf4 and Qf5 mixed plutonic fan deposits are composed of extremely coarse cobbly to bouldery particles (Fig. 10) in which very limited amounts of rodent activity occur. Pocket mice and kangaroo rats live in underground burrow systems that are usually located beneath canopies of large plants, especially creosotebush. In addition to the burrow systems in which they reside, these seed-eating rodents also make numerous, shallow excavations up to a few centimeters deep in other places, including the open

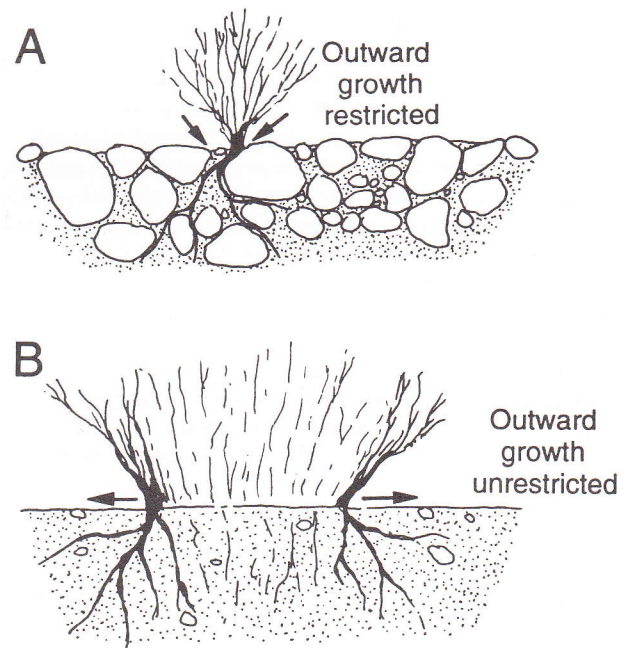


Figure 7. Growth responses of creosotebush in extremely coarse-grained versus fine-grained alluvium. Unrestricted outward growth in fine substrates is apparently crucial for the development of clones.

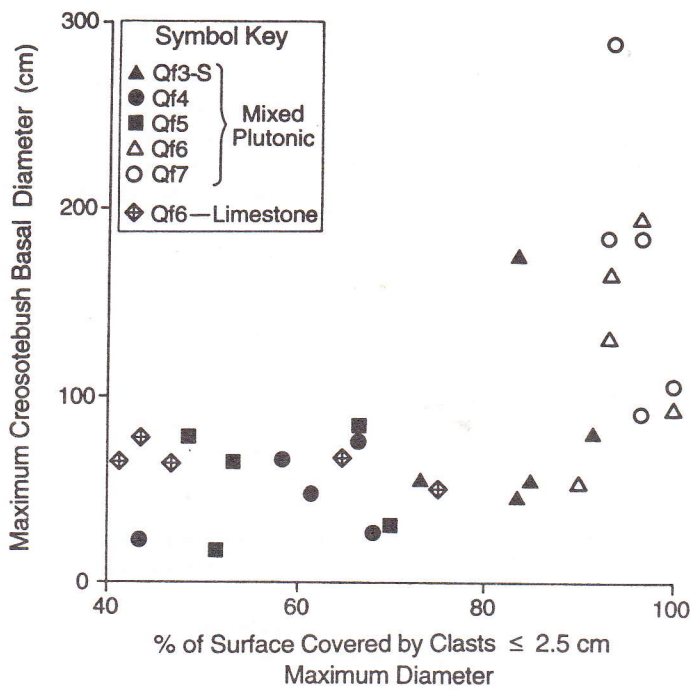


Figure 8. Maximum basal diameters of creosotebush from 6 alluvial surfaces as a function of alluvium texture. Each datum represents the largest creosotebush within a 100 m² sampling plot. Fine, gravelly to sandy deposits are to the right, and coarse, cobbly to bouldery deposits are to the left.

areas between shrub canopies, in their search for and burial of seeds. Both activities are impaired in substrates of extremely coarse, stony materials. The rarity of rodent burrowing and seed excavation in coarse deposits is a pattern that holds throughout the study area in other parent materials, including the mixed volcanic and limestone parent materials.

Along the transmission line road east of Kelbaker Road, the Pleistocene-aged (Qf3-Qf5) deposits of mixed plutonic parent materials were all originally composed of coarse cobbly to bouldery alluvium. However, on the Qf3 surface, enough time has elapsed (estimated time = 130,000 years) to weather the original large plutonic clasts to a degree so that the surface is composed of fine particles and subsurface clasts are considerably decomposed (Fig. 11). Physical weathering of these coarse clasts apparently leads to increased rodent activity. The considerable bioturbation of the Qf3 surface by rodents has further impacts on soils and plant responses. This set of intertwined interactions provides an example of a possible feedback loop between biotic and physical components of the landscape system (Fig. 1).

Feedback loops involving linkages 1, 2, 3, and 4: Soils, plants, and animals. Burrowing and seed excavation activities by rodents on the Qf3 surface provide a significant, continued perturbation of the soil. Scattered, remnant patches of varnished desert pavements are present on this surface but most of these remnants (and especially their margins) exhibit substantial degradation due to bioturbation by rodents. These excavation activities disrupt surface

pavement and underlying Av soil horizon; such disruption enhances infiltration. The greater heights, basal diameters, and cover of creosotebush on the Qf3 surface as compared to the other Pleistocene surfaces (Figs. 4, 6, 9) may be a consequence of this enhanced infiltration. This kind of feedback is patchily distributed over the surface of the Qf3 surface. Shallow excavations in search of buried seeds occur throughout the surface, but since the deeper burrow systems are found exclusively beneath plant canopies, some of the greatest impacts on soil conditions occur directly beneath and near such plants, especially long-lived creosotebush. New burrow systems are rarely constructed on these well-developed soils. Instead, these rodents reoccupy previously constructed burrow systems and these systems may be used for many centuries (Mary Price, Univ. California, Riverside, personal communication). This long-term, localized burrowing activity leads to the disruption of relatively impermeable soil horizons of the soil (such as well-developed argillic and calcic horizons). Soils beneath creosotebush in which rodents are active are considerably looser and more permeable. These kinds of soil alterations directly benefit creosotebush, apparently because of enhanced infiltration (Chew and Whitford 1992). Furthermore, in this environment where rodents rarely excavate new burrow systems but rather re-use and elaborate existing ones, the persistence of individual, long-lived creosotebush plants has a positive feedback to rodent populations and activity.

The presence of larger creosotebush plants (both heights and basal diameters) on the Qf3 surface in comparison to the Qf4 surface may be directly related to the greater amount of

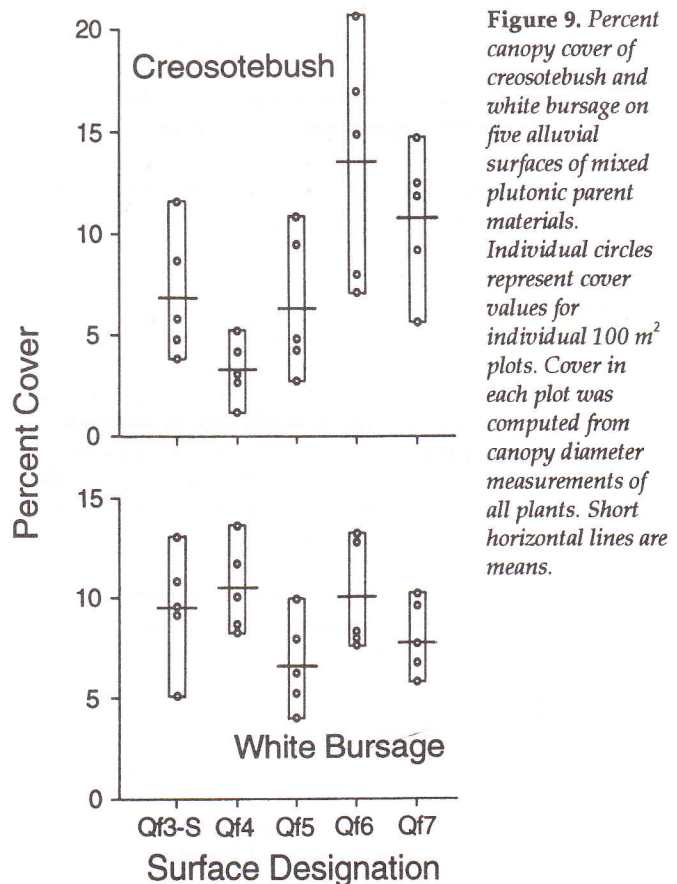


Figure 9. Percent canopy cover of creosotebush and white bursage on five alluvial surfaces of mixed plutonic parent materials. Individual circles represent cover values for individual 100 m² plots. Cover in each plot was computed from canopy diameter measurements of all plants. Short horizontal lines are means.

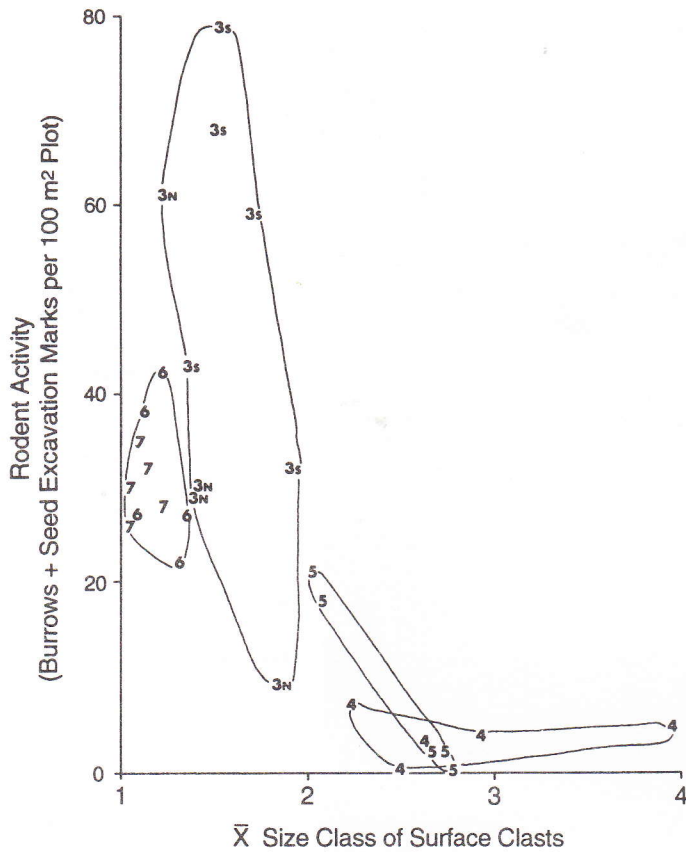


Figure 10. Variation in rodent activity as a function of average clast size within different alluvial surfaces of mixed plutonic parent materials. Boldface numbers indicate individual data points for Qf3, 4, 5, 6, and 7 sites. The values for rodent activity on the Qf6 and Qf7 surfaces underrepresent actual rodent activity because many excavation marks are quickly obscured in the fine-grained alluvium. The cover class designations represent the same logarithmic scale shown in Fig. 11

rodent activity beneath individual shrubs. On the Qf3 surface, rodent burrow systems were present beneath 61% of all creosotebush plants. On the Qf4 surface where clones are absent and creosotebush heights are generally considerably shorter, only 8% of all plants had rodent burrows beneath their canopies.

Such a system in which soil components, plant responses, and animal activity are all dependent variables and simultaneously vary in response to each other is far more difficult to study than a simple relationship between a single dependent variable and one or more independent variables. Despite the complexity of these landscape systems, environments such as the Providence Mountains area provide various comparisons that can potentially lead to a deciphering of some of these relationships.

An example of a contrasting impacts of rodents on soil development and vegetation responses: A desert pavement atop an eolian deposit.

Kelbaker Road crosses several narrow areas of well-developed desert pavements approximately 3.8 mi N of the powerline road and 2.0 miles south of Vulcan Mine Road.

This area is located about 0.2 mi S of Field Trip stop #2. These pavements provide an example of different possible kinds of relationships between soil conditions, vegetation, and rodent activities in a contrasting substrate.

Physical environment. The desert pavements are composed of a single layer of relatively small clasts (coarse gravel and small pebbles). Approximately 90% of pavement clasts are between 4 and 16 mm in greatest diameter (Fig. 12A, B). The pebble pavement is formed on top of a deposit of eolian sand (S. G. Well, personal communication, May 1994) that is generally 25-40 cm deep. This deposit of eolian sand directly overlies a coarse, gravelly alluvial deposit; the contact between these deposits is abrupt (Fig. 12C). The eolian deposit is probably similar in age to the Qe2 surfaces dated to the earliest Holocene (8.4K-10.4 K ybp), and the underlying alluvium is probably of late Pleistocene age (Qf5 surface, 8.6K-16.8K ybp)(McDonald 1994). The thin varnish coatings on upper sides of pavement clasts and level of soil development in the eolian deposit are consistent with an age of earliest Holocene. The soil consists of a thin Av horizon approximately 1.5 cm thick located directly beneath the pavement clasts and an underlying, slightly reddened (7.5YR-10YR 6/4 dry Munsell color) sand-textured, B horizon that exhibits weak structural development (Fig. 12D). This pavement is referred to as the Qe2 pavement.

Studies of these areas of pavements focused on two principal questions: (1) soil conditions that limit plant colonization and growth and (2) factors responsible for the

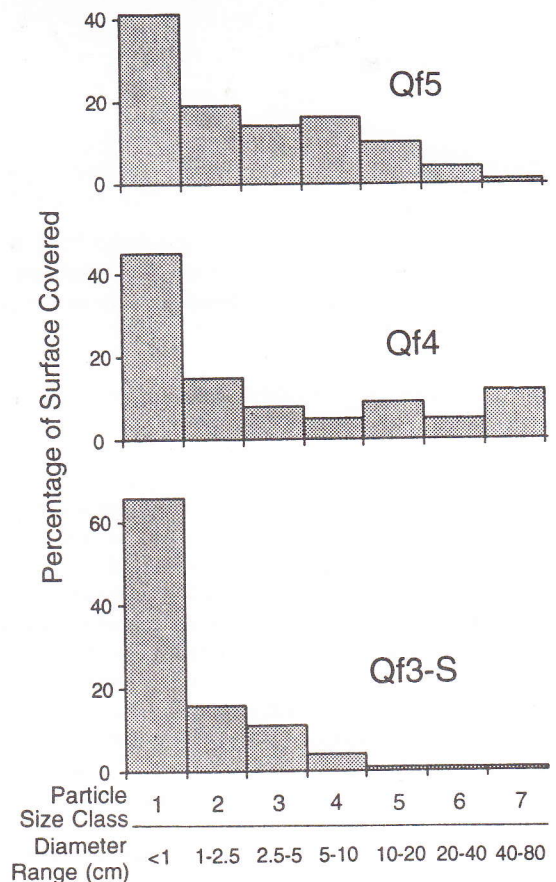


Figure 11. Size distributions of surface clasts on Pleistocene surfaces of mixed plutonic parent materials.

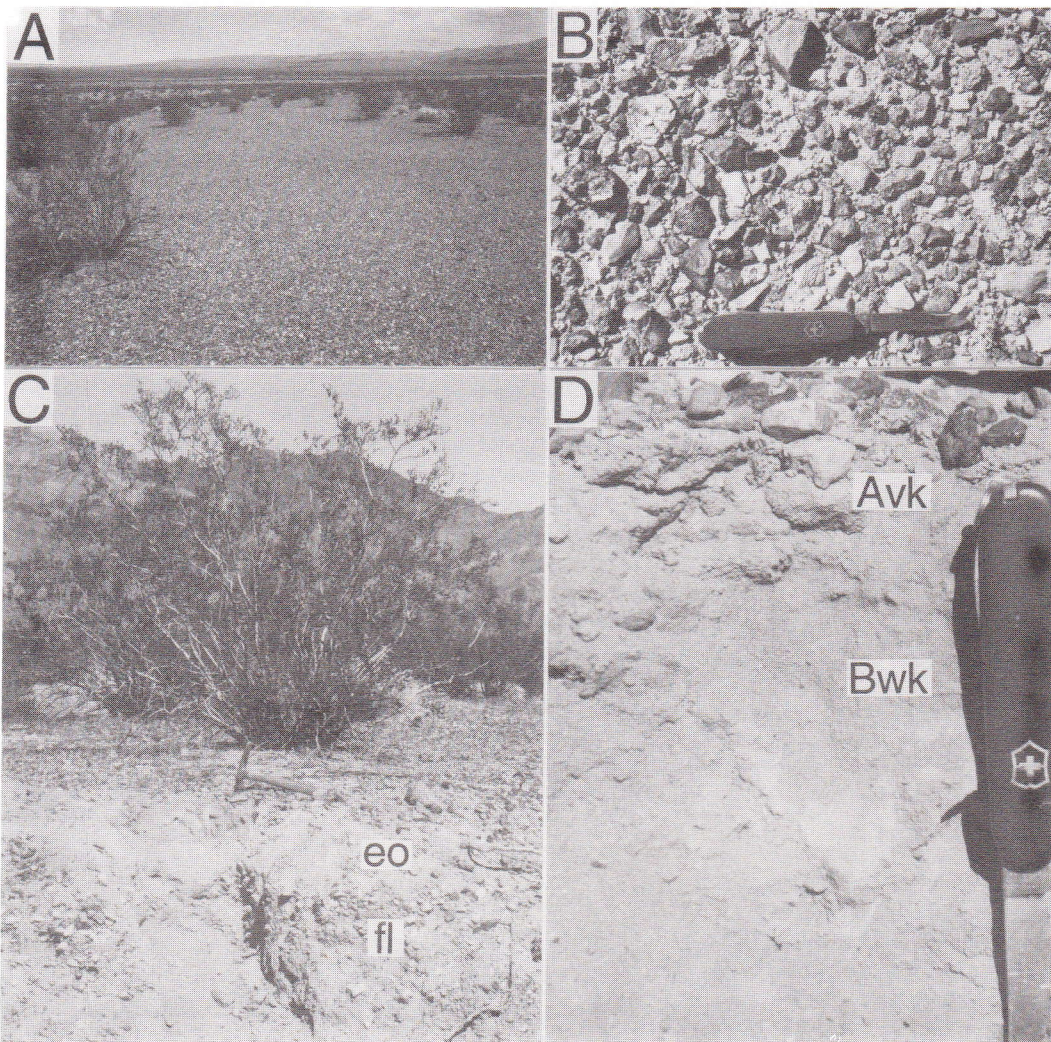


Figure 12. A - Qe2 pavement directly west of Kelbaker Road. B - Close-up of surface pavement. C - Excavation at edge of pavement remnant showing thin eolian unit (eo) capping the coarse, gravelly fluvial deposit (fl). D - Close-up of Avk and Bwk horizons developed below the surface pavement. Note prominent vesicles, 1-2 mm diameter in the Avk horizon.

origin of a stone pavement on top of a deposit of eolian sand. These areas are discussed separately below:

(1) Soil conditions limiting plants. The Qe2 pavement has less perennial plant cover than any other site within the study area. The sparse cover is due to a few scattered creosotebush. White bursage is conspicuously absent from the pavements. Several conditions may be responsible for limiting the development of vegetation on desert pavement surfaces. A tightly packed stone pavement may directly shed considerable water and limit infiltration. Fine-textured Av horizons typically found beneath pavement surfaces also impede infiltration. On extremely old pavement surfaces, a shallow depth of infiltration may lead to substantial salt accumulation at relatively shallow soil levels (Musick 1975, McAuliffe *in press*).

The sparseness of vegetation on the Qe2 pavement cannot be attributed to accumulated salts since this soil exhibits low values for total soluble salts comparable to those found in soils of the Qf6 mixed plutonic surface (where vegetation is

well-developed) and only slightly above those of the Qf7 surface (Fig. 13). On this relatively young pavement, the inhibition of infiltration by the stone pavement surface and the underlying Av horizon is probably responsible for the extremely low density and relatively small sizes of creosotebush.

The reason for the absence of white bursage on this surface is enigmatic, though, since this shrub is abundant on alluvial surfaces with extremely well-developed pavements (e.g., the Qf4 and Qf5 mixed plutonic surfaces; see Fig. 9). However, in comparison to the Qe2 pavement, the Qf4 and Qf5 alluvial surfaces probably respond differently in terms of infiltration of small precipitation events due to great differences in the distribution of sizes of surface clasts. The pavement of the Qe2 surface is extremely homogeneous; ninety-nine percent of pavement clasts are less

than 5 cm in diameter (Fig. 12B). Although water enters soil at contacts and slight gaps between pavement clasts, the extremely low hydraulic conductivity of the underlying Av horizon leads to lack of infiltration and most water is probably lost as runoff from the pavement (Fig. 14). Only where shallow flutes are cut into the pavement and the Av horizon is truncated does white bursage appear. The lack of infiltration is probably most pronounced in more frequently occurring, small precipitation events. The only significant soil moisture recharge at rooting depths for any plants may occur only after extremely infrequent, long-duration, low intensity storms. This extremely infrequent recharge of deeper soil layers may be sufficient for limited numbers of creosotebush with extensive root systems. Although drought-deciduousness allows white bursage to persist for periods when soil water is unavailable, this shrub species may require more a more frequent wetting of the root zone by small precipitation events than likely occurs beneath the Qe2 pavement.

A completely different pattern of infiltration during small

precipitation events probably occurs on relatively coarse-textured alluvial surfaces. A large fraction of the Qf4 and Qf5 surfaces consists of cobble and even small boulder-sized clasts (Figs. 11, 14). Runoff locally generated by large clasts could contribute to locally high receipts of water at the soil surface and infiltration to considerable depth, even in small precipitation events where incident precipitation alone does not infiltrate below the Av horizon. In this way, small precipitation events that are incapable of wetting the rooting zone on the fine-grained, homogeneous Qe2 pavement, are able to provide more frequent, but patchy soil water recharge on the coarse-grained Qf4 and Qf5 surfaces.

(2) Origin of the Qe2 pavement: hypothesized role of past rodent activity. The pavement that covers the surface of the fine-grained Qe2 deposit probably has a different origin than pavements formed on coarse gravelly to stony alluvial deposits. Pavement formation on such coarse parent materials has been attributed to separation of a layer of surface clasts from other underlying coarse material by the accumulation of fine-grained eolian materials (primarily silt- and clay-sized particles) in the underlying Av horizon. The continued accumulation of eolian fines in the Av horizon leads to uplift of and increased separation between surface clasts and rocky parent materials (McFadden et al. 1987, McDonald et al., this volume). However, on the Qe2 surface and other similar, fine-grained parent materials that initially lacked coarse materials on the surface, deflation of fines leading to an accumulation of an erosion-resistant surface lag may be the process responsible for initial pavement formation (McFadden et al. 1992).

The sandy eolian deposit beneath the pavement surface contains a small fraction of coarse clasts and the size distribution of these subsurface clasts is nearly identical to that of the overlying pavement (Fig. 15). Two 1000 cc samples (10 x 10 x 10 cm cubes) of the eolian sand taken at 5-15 cm depths yielded 72 g and 104 g of coarse clasts exceeding 5.5 mm diameter. Three samples of 10 x 10 cm areas of the Qe2 surface pavement yielded 90-111 g of clasts exceeding 5.5 mm diameter. Given the quantity of coarse clasts contained within the eolian deposit compared, a little more than a 10 cm depth of fines would have had to be deflated to yield a lag as dense as the present pavement.

Before such a deflation process could yield a stabilizing surface lag, the eolian sand deposit had to have contained sufficient coarse clasts. These coarse clasts were certainly not wind-transported, nor is there any evidence of intermixed fluvial activity (e.g., stratified, fluvially deposited layers) within the upper eolian unit. The presence of the coarse clasts within the eolian sand deposit may be due to an episode of bioturbation by rodents before the time when the surface lag was created. Burrowing activity could have easily mixed coarse gravel to pebble-sized clasts from the underlying alluvium into a relatively thin, overlying eolian deposit. The following scenario is hypothesized for the formation of the Qe2 pavement (Fig. 16):

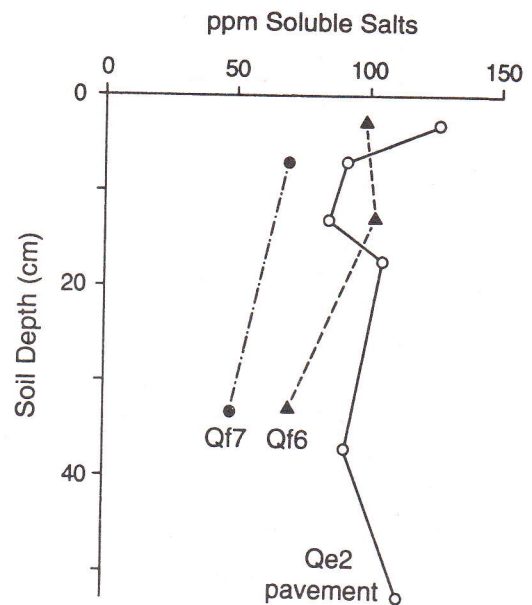


Figure 13. Soluble salts as a function of depth in the Qe2 pavement soil and two mixed plutonic alluvial surfaces.

(A) Deposition of alluvium in latest Pleistocene.
 (B) Burial of alluvium by thin deposit of eolian sand in earliest Holocene.

(C) Eolian sand deposits provide a productive environment due their high permeability and rapid infiltration of precipitation. Some of the most productive environments in the Providence Mountains piedmont are eolian sands upon which pavements have not formed either because of their young age (e.g., Qe3 deposits overlying Qf6 deposits approximately 2.5 km to the northwest) or greater thickness (e.g., a Qe1 surface remnant located 6 km S-SW). Burrowing activity by rodents is extremely high on such eolian surfaces. If the eolian deposit is relatively thin, as is the case of the Qe2 surface, such burrowing mixes coarse alluvial clasts throughout the upper eolian deposit. The present Qe3 surface, about 40 cm thick, and also positioned

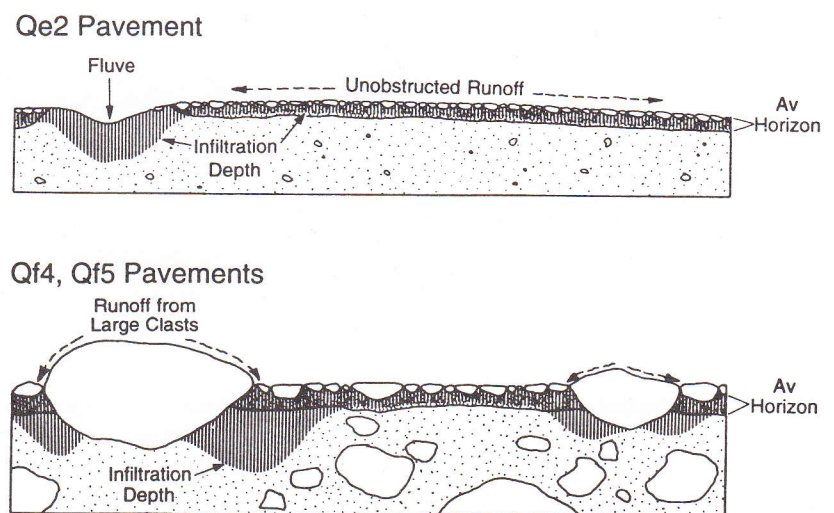


Figure 14. Effects of variation in surface clasts sizes in runoff and patterns of infiltration on the Qe2 pavement versus Pleistocene-aged alluvial surfaces

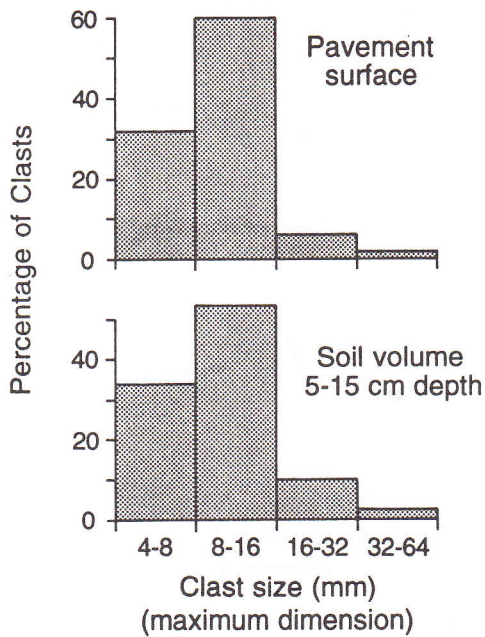


Figure 15. Size distributions of clasts found on surface of Qe2 pavement (top) versus in the soil volume (bottom). Pavement surface data is from three- 100 cm² sample areas and the soil volume data is from two- 1000cc cubic sample volumes.

on top of a coarse, gravelly alluvial deposit, exhibits extremely high rodent activity and the hypothesized mixing of coarse clasts throughout the upper eolian unit.

(D) Deflation of the surface in the early Holocene removes fines, leaving a surface lag of coarse clasts. The accumulated lag protects the surface from further deflation.

(5) Stable surface conditions lead to development of the Av horizon which further impedes infiltration and plant

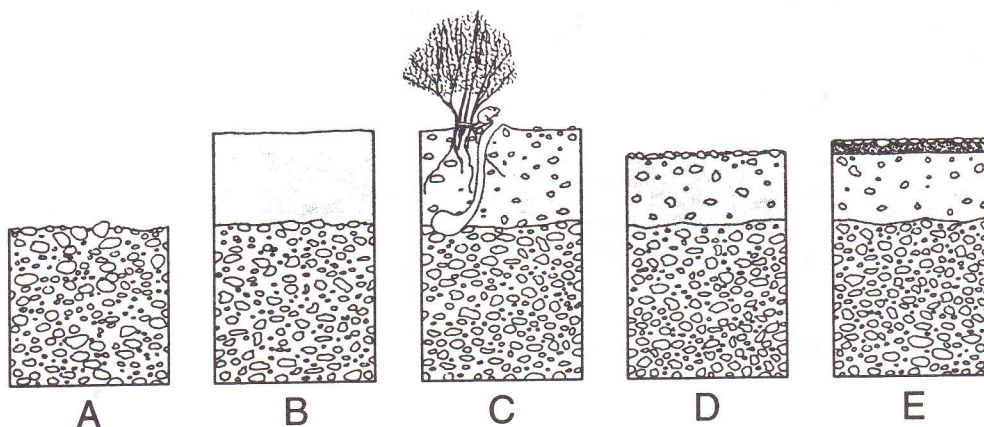


Figure 16. Hypothesized scenario of the formation of the Qe2 pavement. A - Deposition of coarse gravelly alluvium in late Pleistocene. B - Alluvium is capped by thin eolian deposit in earliest Holocene. C - Sandy substrate promotes high infiltration and corresponding high plant productivity and rodent burrowing. Rodent burrowing is responsible for mixing coarse clasts from underlying alluvium throughout the eolian deposit. D - Deflation of surface fines produces a lag of coarse clasts. E - Under stable surface conditions, influx and accretion of loess beneath surface clasts leads to development of the cumulic Av horizon. With increasing development of the Av horizon, the potential for infiltration to lower levels in the soil is reduced.

development. Such conditions could only arise with a marked reduction in rodent burrowing activity. The pebble lag overlying the eolian sand would probably not have been sufficient in itself to mechanically restrict excavation by rodents. Drier climate conditions that apparently occurred in the Mojave Desert region between approximately 6000 and 9000 years ago (McFadden et al. 1992) may have contributed to substantially diminished rodent activity due to the lack of food production in the form of ephemeral plant seeds. Rodent populations respond directly to such climate signals. For example, results of a 23-year long monitoring project in the Sonoran Desert have revealed 10-to 15-fold differences in rodent biomass in years immediately following moist periods (El Nino events) compared with extensive dry intervals, such as the mid-1970's drought (personal communication, Y. Petryszyn, University of Arizona, Tucson). The period between 6000 and 9000 years ago was also a time when considerable deposition of desert loess occurred in the immediate region (McFadden et al. 1992). Enhanced input of loess during this period may have rapidly generated an Av horizon which created an effective barrier to infiltration of moisture.

Conclusion

An understanding of ecological patterns and processes in desert piedmonts has benefitted greatly by a foundation of knowledge about the physical environment gained by geomorphologists and soil scientists. Ecologists, in turn, can reciprocate through investigations of how biotic components affect soil and landscape processes in these desert environments. This chapter has presented a few preliminary hypotheses and results of research in progress on the Providence Mountains piedmont which combine perspectives of ecology, geomorphology, and soil science in order to better understand potentially complex landscape systems. Describing many of the specific relationships within these systems and testing hypotheses regarding various cause-and-effect linkages requires much more study and will provide continuing challenges for interdisciplinary research. The west slopes of the Providence Mountains, with contrasting parent materials and ages of alluvial and eolian deposits, will continue to be an important natural laboratory for these studies.

Acknowledgements

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The Joshua Tree

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Certainly the most spectacular plant in the Mojave Desert is the Joshua tree, *Yucca brevifolia*. Its large mass and endlessly-varying appearance make it scenically striking and its very existence in a hot and arid land is curious. In all of the California deserts it is the only tree that survives on the open flatlands. What's more, it can withstand fire, often depends upon "nurse" plants when young and in death provides critically-important nest sites for birds.

To many, the Joshua tree is the symbol of the deserts of California. From the center of the Mojave Desert, it ranges beyond all of the Mojave's boundaries—west to Gorman in the Antelope Valley, north to Goldfield in Nevada, east to Wickenburg, Arizona, and south to the Cottonwood Mountains in Joshua Tree National Park. It is unquestionably the most visually-dominant lifeform in the region.

Contrary to what many early naturalists believed, however, *Yucca brevifolia* is not confined to the Mojave Desert. The Joshua tree can be found growing next to saguaro cactus in the Sonoran Desert of west-central Arizona and grows alongside pinyon pines in the conifer forests of the San Bernardino Mountains. It freely mixes with scrub oak and other chaparral plants in the Cajon Pass and occurs with manzanita just east of Morongo Valley. It is an adaptable plant and although the vast majority of individual Joshua trees do occur within the Mojave, the species does not depend upon this desert's unique climate for survival (Cornett, 1989).

Perhaps what is most odd about the story of *Yucca brevifolia* is how little is known of its ecology. That such an unusual and well-known desert plant has been studied so little reflects the remoteness of these trees to major population centers and universities. In fact only one recent paper of an ecological orientation has been written and, although the study was an excellent one, it was never published (Peter Rowland, 1978. Ph.D. dissertation, Univ. of Calif., Riverside). This paucity of information was what led the Palm Springs Desert Museum's Natural Science Department to begin a twenty-year study of these curious plants.

The first person to ever write about Joshua trees was John C. Fremont, an

early explorer and the man who became one of California's first senators. On his initial trip across the Mojave Desert in April of 1844, Fremont wrote that he was "struck by the sudden appearance of tree yuccas which gave a strange and southern character to the country." He also noted that they were "stiff and ungraceful . . . the most repulsive tree in the vegetable Kingdom" (Rowland, 1978).

What is one man's repulsion is another's attraction. To the early Mormon pioneers who trekked across the Mojave Desert from Salt Lake City, the Joshua tree was a sign that their journey was half over. Legend has it that they named the tree in the late 1800s on one of their first migrations to

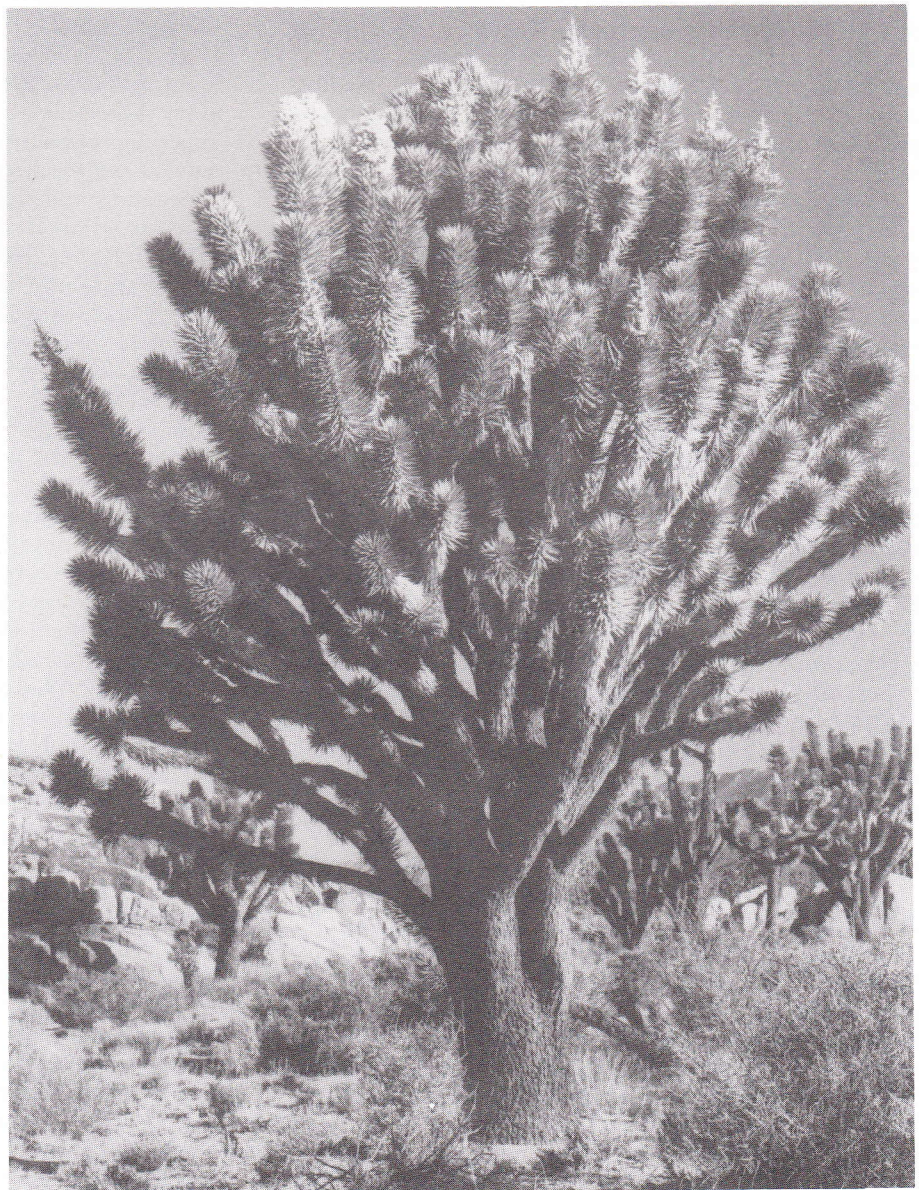


Figure 1. Joshua tree on Cima Dome.

San Bernardino. To the Mormon pioneers, the limbs of the giant tree yucca resembled the up-stretched arms of Joshua beckoning them to the promised land.

Webster's dictionary (1981) defines a tree as "a woody perennial plant having a single elongate stem generally with few or no branches on its lower part." By that definition, *Yucca brevifolia* usually grows in the form of a tree. But some are clearly more tree-like than others. In the western and southern Mojave Desert, Joshua trees have tall trunks that often extend over ten feet before branching. Technically speaking, this variety is known as *Yucca brevifolia brevifolia*. In addition to its tall trunk and relatively long leaves, it is characterized by "pseudodichotomous branching." This means that it only branches after the growing tip has been destroyed during the flowering process.

Of the two varieties of Joshua tree, it is *Y. b. brevifolia* that reaches the greatest height. The tallest specimen is located in Joshua Tree National Park and was named "Emily's Tree." As of July 1, 1992, it stood 36.4 feet in height. The second tallest Joshua tree is located on lands managed by the United States Forest Service within the San Bernardino Mountains. It has been named "Champion" and as of May 30, 1988, measurements show this tree to be 32.3 feet in height with a crown diameter of 32.7 feet. Unfortunately, it receives no protection from vandals and is riddled with gunshot injuries.

The other variety of Joshua tree, *Yucca brevifolia jaegeriana*, is found in the eastern Mojave Desert and is generally much smaller. It is characterized by somewhat shorter leaves, shorter trunk and true dichotomous branching. In practical terms this means that this variety usually (but not always) branches prior to its first flowering. Typically, *Y. b. jaegeriana* is much more densely branched giving it a compact and symmetrical appearance.

Joshua tree seeds germinate readily in a laboratory environment but under natural conditions the resultant seedlings are quickly devoured by herbivores. Robert Moon of Joshua Tree National Park relates how a park ranger in the park once placed numerous seedlings out on his porch for sun one morning. By late afternoon, ground squirrels had eaten every one of the sprouts. Crown and rhizome sprouts can also become victims. I recall searching for desert night lizards (*Xantusia vigilis*) under a branch that had broken off a large, mature Joshua tree. When I turned the branch over not only did I find a night lizard, I also revealed a four-inch rhizome sprout. The next day I returned to the site and found the exposed sprout had been devoured by a black-tailed jackrabbit (*Lepus californicus*).

The elimination of sprouts by herbivores favors those young Joshua trees that can somehow avoid being devoured. One habit found among Joshua tree populations on Cima Dome (in San Bernardino County, California) is the utilization of a "nurse" plant during the early years of the tree's growth. By germinating underneath another perennial plant species, particularly one that is not relished by herbivores, a Joshua tree can spend the first few vulnerable years of its life protected by the dense stems of a nurse plant. Then, after it is at least three years old and protected by dozens of spine-tipped leaves and tough fibrous tissues of its own, it emerges above its host. Eventually, the nurse plant dies as it is overwhelmed by the ever-enlarging Joshua tree.

The requirement of a nurse plant seems dependent upon

the intensity of foraging. In Joshua Tree National Park it appears that about half of the young trees are associated with nurse plants. Potential Joshua tree foragers include mule deer, rabbits, woodrats and ground squirrels. In the Antelope Valley, near Palmdale and Lancaster, deer are nonexistent and in the vicinity of paved roads other herbivores seem uncommon. In such situations less than one percent of the young trees are found associated with nurse plants. However, on the Cima Dome in the eastern Mojave Desert, a young Joshua tree cannot survive unless the seed lands in the middle of a shrub. On the dome, cattle grazing is severe, deer and other native herbivores are common, and apparently every seedling is eaten unless it is protected.

We are continuing to gather data on the relationship between Joshua tree populations and nurse plants. Interestingly, our early suspicions are that cattle grazing may inadvertently result in an increase in Joshua tree densities. An ungrazed Joshua tree habitat is likely to harbor more bunch grass than shrubs. When herbivores are present they consume the bunch grass and anything else growing within it. Intense grazing by cattle reduces or nearly eliminates the bunch grass. Eventually it is replaced by shrub species such as desert tea (*Ephedra nevadensis*) and spiny menodora (*Menodora spinescens*). Since shrubs provide superior protection for Yucca seedlings, more young trees survive and the density of Joshua trees increases.

Seedling predation is not the only hazard a Joshua tree may face within its lifetime. Because this species typically grows at higher elevations, from about 2,000 to over 6,600 feet, temperatures are cooler and rainfall somewhat higher than at lower-altitude desert locations. The result is more abundant vegetation with a denser cover of perennials and sufficient fuel to support occasional wildfires. Joshua trees burn readily since their trunks are either covered with expired dry leaves or coarse bark. With these characteristics the above-ground portion of the trunk is usually destroyed in a wildfire.

That does not mean, however, that fire kills Joshua trees—at least not all Joshua trees. In the Keys Fire of 1981, near the community of Pioneertown, San Bernardino County, California, approximately 100 acres burned to the ground apparently killing most of the plants including about six hundred Joshua trees (Cornett, 1994). However, when we returned in 1989 many of the trees were still alive. Nearly all of the original trunks were killed but 40% had grown one or more crown and/or rhizome sprouts at the base of their old burned trunk. In some cases eight or nine new sprouts had emerged, some reaching four feet in height in just eight years (quite rapid growth for a notoriously slow-growing plant).

Not surprisingly, no seedling sprouts were found—only crown and rhizome sprouts. Seedlings are easily killed when eaten since they do not have the nutritional and moisture reserves of an established parent plant. Even if browsed severely, a crown or rhizome sprout can generally continue to produce new leaves. Sprouts face the greatest threat from herbivore attacks after fires, when Joshua tree leaves may be one of the few food resources available.

There are many areas where Joshua trees have multiple trunks as a result of the development of several crown sprouts after a wildfire. In light of our research on the effect of fire on Joshua trees, it is tempting to speculate that such

areas reflect the occurrence of a past wildfire, perhaps decades earlier. However, there are several *Yucca brevifolia* populations where multiple-trunked trees predominate yet no records of fire exist. Hopefully, further study will explain the cause or causes of multiple-trunked individuals.

Although fire may temporarily eliminate a majority of the Joshua trees in a given area, there is at least one positive effect. Dead Joshua trees provide critically-important nesting sites for birds. Since the wood becomes relatively soft and dry after the fire, and the trunks are typically left erect, a vast number of nest sites become available to woodpeckers—in a region where none may have existed before. Rarely can the ladder-backed woodpecker (*Picoides scalaris*) or the common flicker (*Colaptes auratus*) drill nest cavities in living branches or trunks of Joshua trees. With one exception in my experience, only dead wood is used. Once a fire sweeps through an area, dead wood is easy to find and nest cavities can be easily constructed. I have seen ladder-backed woodpeckers take advantage of this but only on the periphery of the burned area. Fire removes everything and woodpecker food may be in short supply. I suspect that nesting on the interface between the burned and unburned areas enables woodpeckers to utilize the best of both worlds: the good nesting sites of the burned area and the abundant food resources of the unburned area.

As a species, Joshua trees are slowly declining in number. This is a result of the ever-increasing human population and the thousands of humans who wish to escape crowded urban areas to live within the Mojave Desert. Around the communities of Victorville, Palmdale, Lancaster and Yucca Valley, thousands of these trees have been uprooted by bulldozers to make way for new commercial and residential developments. Although stands of *Yucca brevifolia brevifolia* are protected in Death Valley and Joshua Tree National Parks, Saddleback Butte State Park and Red Rock Canyon State Recreation Area, up until 1993 there were no preserves that protect the subspecies *Yucca brevifolia jaegeriana*. This has now changed with the establishment of the Mojave National Preserve in 1995 and the preservation of the vast stand of Joshua trees on Cima Dome.

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Characteristics, Age, and Tectonic Implications of the Mid Hills Pediment

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Abstract

Erosional surfaces exposed over an area of the northeastern Mojave Desert from the New York Mountains to the Fenner Hills describe a single gently south-sloping compound surface of at least 1000 km². The compound surface is outlined by pediments and by planar unconformities beneath early Miocene tuffs. The pediments and unconformities are similar in elevation and slope, and in places grade to one another, suggesting that early Miocene or older erosional surfaces have been modified only slightly. Deeply weathered nearly planar surfaces may have formed in the early Tertiary and were stripped to form modern pediments. The early Miocene Wild Horse Mesa Tuff thins southward in the present downslope direction, suggesting that the early Miocene surface subsequently was tilted to the south. Tilting probably took place before or as Pliocene(?) playa sediment was deposited near the Piute Range, so was during the late early Miocene to Pliocene. Faults bounding the tilt block lie on the west side of the Providence Mountains and Mid Hills, east side of the Piute Range, and north of the southeastern Granite Mountains.

Introduction

Pediments are low-relief rock-floored erosion surfaces common in desert areas underlain by granitoids. Oberlander (1974) distinguished *pediments*, in which erosion surfaces are cut into the same rock underlain by adjacent high-relief regions, and *planation surfaces*, which are bounded by high-relief areas underlain by more resistant rock. Pediment-forming processes are still enigmatic, and Oberlander (1974) considered that pediments in the Mojave Desert represent stripped relic Tertiary erosional surfaces and do not form under current arid conditions. Pediments are widespread in the northeastern Mojave Desert (Fig. 1), and were used explicitly by Hewett (1956) to analyze the extent of onlap of Miocene volcanic rocks in determining Miocene paleotopography. Pediments in the Mid Hills area of the northeastern Mojave Desert can be contrasted to areas of thicker alluvium where no rock outcrops occur.

This paper describes pediments and other planar erosional surfaces in the Mid Hills area, making a case for near-equivalency of several adjacent surfaces that together may compose a single compound surface. Where parts of this surface are dated by overlying sediment and volcanic rock, information on the time needed for its development can be estimated. The compound surface may delineate a nearly unbroken tectonic block.

Mid Hills Pediment

Pediments cut on granite are exposed in parts of the Mid Hills and nearby areas, and many gently sloping erosional surfaces of the region are overlain by early Miocene volcanic rocks (Fig. 2). They pediment exposures and pre-volcanic unconformities appear to represent parts of the same extensive surface, here called the Mid Hills pediment. Pediment surfaces in the Mid Hills area primarily expose the Cretaceous Mid Hills Adamellite and Rock Springs Monzodiorite (Beckerman and others, 1982); other substrates include Cretaceous granite in the Vontrigger Hills and Fenner Hills (Miller and others 1991), Early and Middle(?) Proterozoic granite in the Vontrigger Hills, and Jurassic granite in the Colton Hills (Goldfarb and others, 1988). The

pediments are younger than the youngest (~75 Ma) granites they bevel.

Description

Some of the best pediment examples in the region occupy Round and Gold Valleys and lie near the southern slopes of the New York Mountains (Fig. 2). A pediment underlies all of Round Valley, where it exposes the Mid Hills Adamellite and Rock Springs Monzodiorite and slopes gently east. Ringing highlands are primarily underlain by the same granites, but gabbro forms hills on part of the south side and thick dikes contribute some resistant exposures on the north side. Relief across the 4-km-wide pediment is 100 m, whereas relief on adjacent highlands on the south, west, and north is 50 to 150 m, across lateral distances of 100 to 300 m. An equivalent surface lies north of Cedar Canyon, indicating that the canyon was cut subsequent to the development of the pediment. The Gold Valley pediment lies south of Round Valley across a narrow neck of high-standing Mid Hills Adamellite and Rock Springs Monzodiorite. The Gold Valley pediment is partly cut on the Mid Hills Adamellite and Rock Springs Monzodiorite, and a central part is overlain by alluvium. Part of the bordering highlands are underlain by Early Proterozoic gneiss, Cretaceous gabbro, and Miocene volcanic rocks, but over half is bordered by rocks like those under the pediment. The Gold Valley pediment slopes south about 150 m over a distance of about 3 km. Its eastern side consists of a low divide across which lies more pediment. Pediments lie south of the New York Mountains from the mouth of Caruthers Canyon west to Pinto Mountain, as indicated by fragmentary exposures of granite among thin, discontinuous patches of alluvium. A dissected higher pediment, 40 to 50 m above the current pediment, is visible near Caruthers Canyon.

The Gold Valley pediment is adjacent to a wide tract of ground underlain by gently dipping tuffs associated with the Wild Horse Mesa Tuff, which is 17.8 Ma (McCurry, this volume). The base of the tuff nearly grades to the exposed pediment in several parts of Gold Valley, suggesting that the Miocene erosion surface was similar to the modern

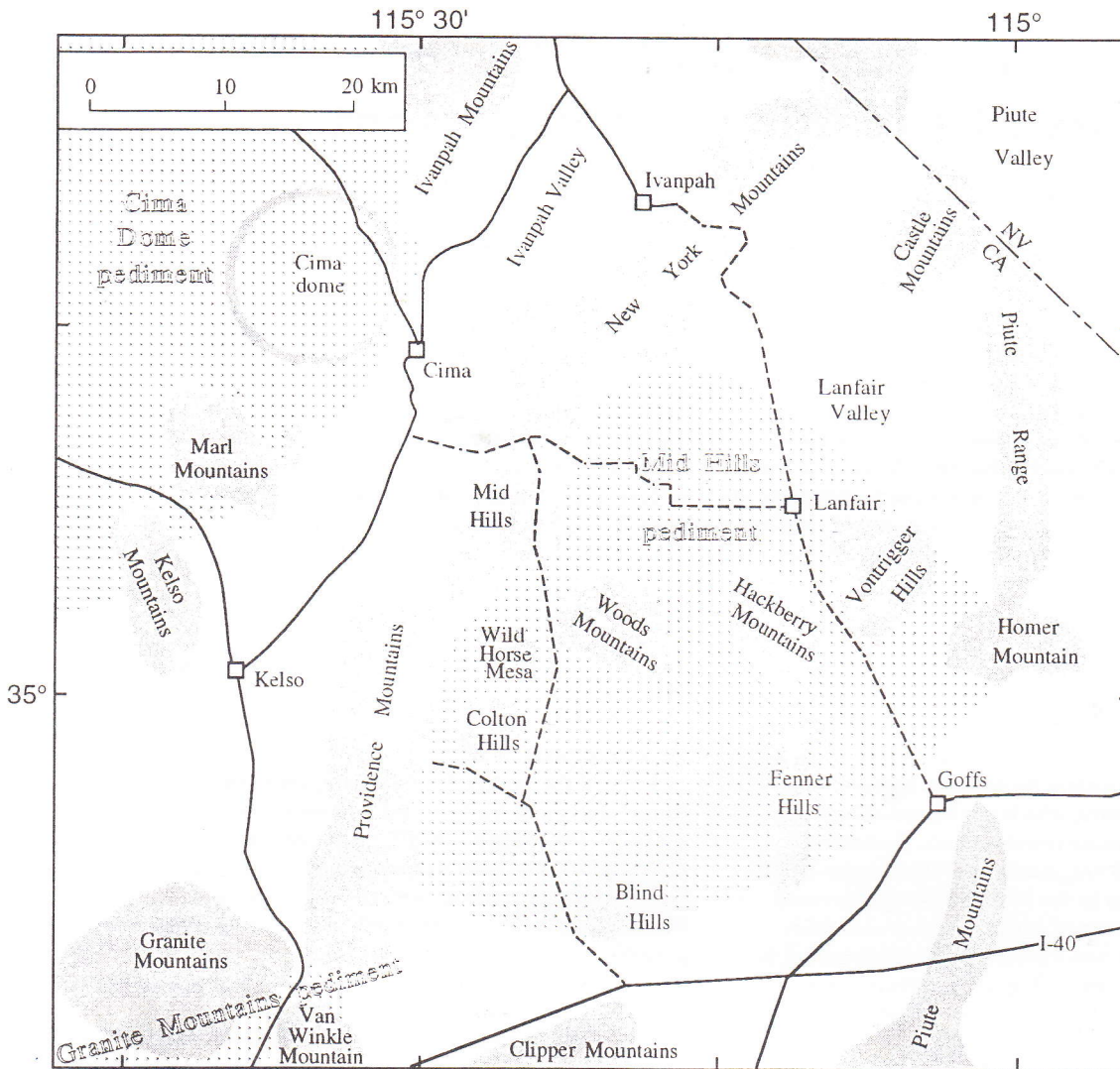


Figure 1. Location map for the Mid Hills pediment. High-relief bedrock areas are shaded; pediments are shown by dot pattern.

pediment but more extensive. Similar Miocene erosion surfaces are indicated by flat bases on mesas of Wild Horse Mesa Tuff south of the Cedar Canyon fault, but the Miocene erosion surfaces are tilted to the north in fault blocks near the mouth of Caruthers Canyon.

Other pediments are present in western Lanfair Valley, the Colton Hills, the Blind Hills, the Vontrigger Hills, and the Fenner Hills (Fig. 1). Much of the land south of the Woods and Hackberry Mountains is underlain by alluvium. Scattered pediments at the Colton Hills, the Blind Hills, and the Fenner Hills indicate a possibility that a pediment is shallowly buried through that area. Lanfair Valley is mostly underlain by alluvium (Miller and others, 1991), but gravity studies (U.S. Geological Survey, 1991) indicate that the alluvial cover is thin an extensive pediment may underlie the area. Western Lanfair Valley is a pediment cut on the Rock Springs Monzodiorite, east of which scattered outcrops of granitoids, carbonate strata, and conglomerate may indicate that a pediment is within a few meters of the surface

(“probable pediment”, Fig. 2).

Miocene volcanic rocks appear to outline old erosion surfaces in addition to those near Gold Valley. Lava and tuff broadly associated with the Wild Horse Mesa Tuff underlie the Woods Mountains, Hackberry Mountains, and part of the Vontrigger Hills (McCurry, 1988). Volcanic rocks locally lie on flat surfaces at each of these locations, and may throughout these ranges.

A pediment dome and ridge lie among the nearly planar pediments described above. The Vontrigger Hills represent a northeast-elongate dome cut on Proterozoic granitoids. The dome is asymmetric, with less than 50 m of relief on the northwest side and about 150 m on the southeast side, and is about 6

km long and 4 km wide. Miocene volcanic rocks lie on a gently-dipping surface on the northwest side of the dome, and suggest the possibility that the dome is an eroded fault block that tilted the volcanic rocks. The pediment appears to merge with the erosional surface beneath the volcanic rocks. A pediment cut on Cretaceous granite west of Homer Mountain forms an elongate north-trending whaleback-like shape whose northern crest merges with the southern Piute Range. The pediment is 3 km wide and 6 km long, and has about 50 m of relief east to west.

The group of pediments and Miocene erosion surfaces described above appear to form an extensive residual low-relief compound surface covering 1000 to 1600 km², depending on how much of Lanfair Valley and the area between the Blind and Fenner Hills is included. No sharp deflections in topographic gradient on the compound surface are visible. This surface has 1000 m of relief, and presently slopes 25 meters per kilometer to the south-southeast on average. This compound surface is here referred to as the *Mid Hills pediment*.

Age

The Mid Hills pediment cuts across early Miocene lacustrine deposits and is locally buried by younger early Miocene volcanic rocks. Lacustrine deposits in the Hackberry Mountains (McCurry, 1985) and along the western part of Wild Horse Mesa close to the northern Providence Mountains (McCurry, 1985; Goldfarb and others, 1988) contain vertebrate fauna assigned to the early to middle Miocene. These lacustrine deposits indicate formation of local sags or faulted margins on subsiding blocks during the early Miocene and postdating pedimentation. The lacustrine sediments are thin in places where they dip gently, and are overlain by volcanic rock, making it hard to determine whether pediments developed above them as well as below them. Laterally, these erosional surfaces match pediments cut into granite. The 18.5-Ma

(Nielson and others, 1990) Peach Springs Tuff (Young and Brennan, 1974) partly buries the pediment, and was possibly deposited widely on it. The Wild Horse Mesa Tuff (McCurry, 1988; this volume) also was widely deposited on a gentle erosional surface. The Wild Horse Mesa Tuff was assigned an age of 15.8 ± 0.4 Ma by McCurry (1988) on the basis of a conventional K-Ar age on sanidine from a pumice fragment. I have obtained a conventional K-Ar age for sanidine from tuff high in the section, where it is presumably unlikely to have acquired detrital feldspar, of 16.8 ± 0.4 Ma. This age discrepancy can be due to detrital contamination in my sample, and/or incomplete fusion of Ar in one or both samples. McCurry and others (this volume) have now revised the age of the tuff to 17.8 to 17.7 Ma on the basis of $^{40}\text{Ar}/^{39}\text{Ar}$ dating; errors and analytical data are not given. This age and older ages for underlying

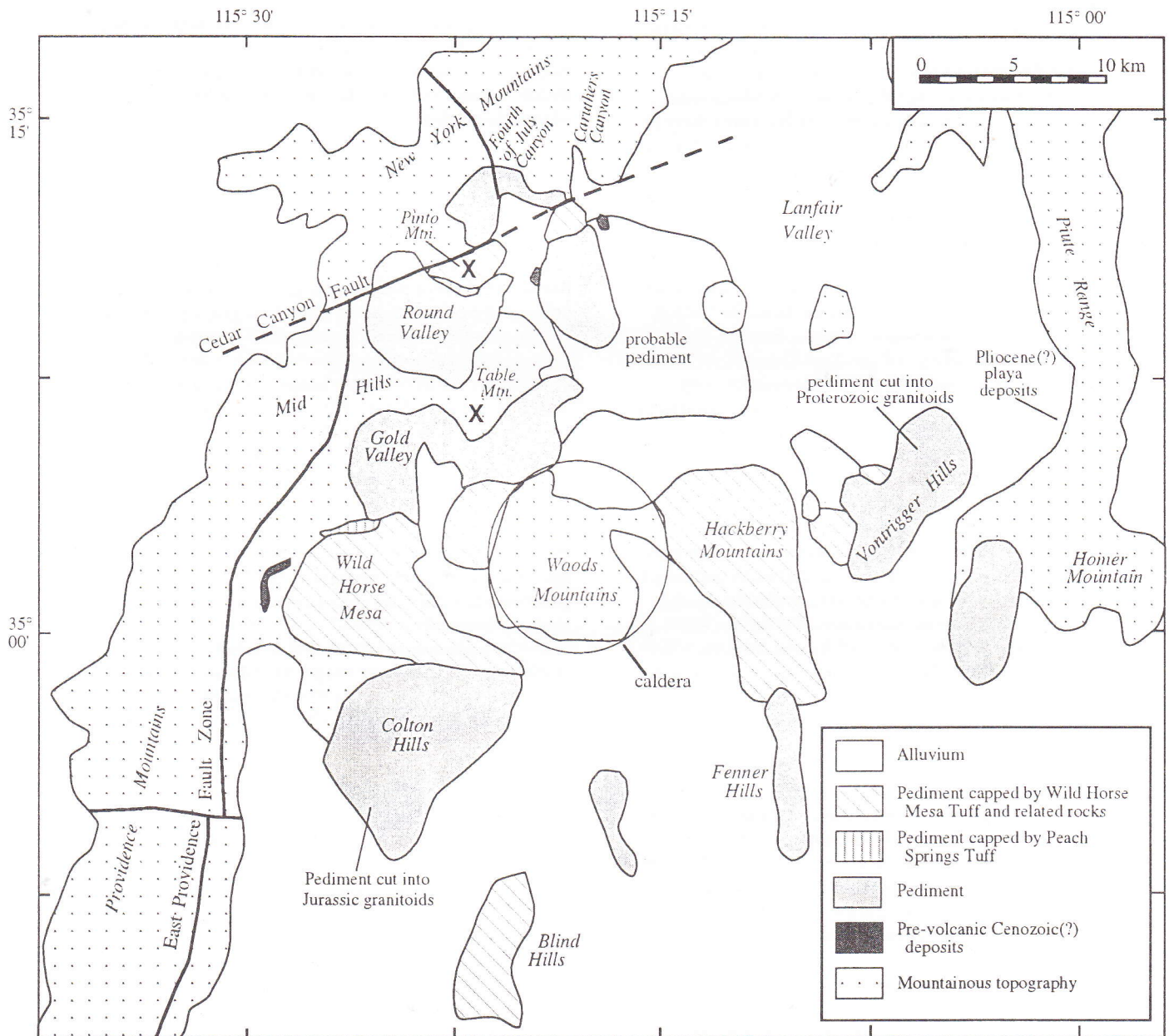


Figure 2. Map of the Mid Hills pediment, showing distribution of pediments, Miocene volcanic rocks on pediments, and features described in the text. Pediment areas enclosed by dashed lines are extrapolated based on sparse exposures of bedrock in largely alluviated areas.

rocks in the Hackberry Mountains are used herein, although some of the conclusions depend on the precise age of the tuff, and will need modification if the assumed age is modified. Volcanic rocks of the Hackberry Mountains, considered part of eruption sequence of the Wild Horse Mesa Tuff (McCurry, 1988), rest on part of the pediment dome in the Vontrigger Hills (McCurry, 1985). The pediment west of Homer Mountain beveled Miocene (17 to 19 Ma) dikes and their Cretaceous granite host (Spencer and Turner, 1985). These relations indicate that at Homer Mountain the pediment formed during the early Miocene—after dikes were emplaced and before 14-Ma basalt was deposited. Elsewhere, the broad erosional surface may have formed prior to the early Miocene before the Peach Springs Tuff and lacustrine rock were deposited.

Paleotopography

Clues to topography preceding the early Miocene suggest that it differed greatly from the Miocene and present topography. The Providence Mountains shed debris eastward, much as today, but the New York Mountains probably were low relief (as described below) and clues in the form of scattered conglomerate deposits south of the New York Mountains indicate that drainage was to the north. Conglomerate east of Pinto Mountain (mismapped by Miller and others, 1991, as Mid Hills Adamellite) is fluvial in origin with clast imbrication and cross-bedding indicating transport to the north. Clasts consist mostly of limestone and Late Proterozoic quartzose rocks, and deformed granitoids of uncertain origin (but possibly sourced from the Colton Hills). Notably lacking are volcanic rocks. Similar, but poorly exposed, deposits lie south of Caruthers Canyon. At the time of deposition, which could be Late Cretaceous to early Miocene, the drainage divide in the Providence Mountains was west of its current position (to provide a source for Proterozoic clasts) and a highland may have been present in the Colton Hills area. Drainage may have proceeded north of the present New York Mountains.

Miocene topography was similar to present-day topography. The widespread distribution of early Miocene tuffs suggests that much of the area of Mid Hills pediment exposure was low relief and parts were probably nearly planar in the early Miocene. Exceptions to the low relief include the ridge between Gold and Round Valleys, where tuffs lapped onto paleohills that partly remain. For instance, only the upper part of the Wild Horse Mesa Tuff lapped onto granitoids at Table Mountain (Fig. 2), whereas the full thickness of the Wild Horse Mesa Tuff was deposited across pediments to the north and south. The east side of the Providence Mountains was probably a high during the early Miocene, since conglomerate was shed eastward (Hazzard, 1954) and overlapped by the Peach Springs Tuff. The Peach Springs Tuff crops out in several parts of the Mid Hills area, suggesting once-widespread exposures, but it apparently was significantly eroded before deposition of the Wild Horse Mesa Tuff because the latter tuff rests on granite in most places.

The Wild Horse Mesa Tuff was erupted locally from the Woods Mountains caldera (Fig. 2) and formed sequences of welded and nonwelded tuff averaging 275 meters in thickness near the caldera (McCurry, 1985). However, rather than thinning uniformly away from the caldera, the tuff is

thicker to the north than the south. The tuff is thin on paleohills south of Round Valley but is about 150 meters thick farther north at Pinto Mountain, suggesting that the deposit at Pinto Mountain represents ignimbrite that was diverted around or over paleohills. The tuff is also thick at Caruthers Canyon; although faulted there, the tuff is greater than 135 m thick. Southward along the flank of the Providence Mountains the tuff thins (Goldfarb and others, 1988) and in the vicinity of the Blind Hills it is only a few meters thick. These data suggest that the topography on this pediment was generally north-sloping at the time of the Wild Horse Mesa Tuff eruption, about 17.8 Ma.

The New York Mountains probably were low at the time of the eruption of the Peach Springs Tuff, for a boulder conglomerate perched on ridges near Fourth of July Canyon (Fig. 2) at about 1760 m (5,800 ft) elevation carries huge clasts of the Peach Springs Tuff in a matrix of fragments of altered granites that are typical of the New York Mountains just to the north. The deposit appears to have been derived from higher in the New York Mountains, suggesting that the Peach Springs Tuff once lay in what is now the highest part of the mountains.

Pediment-forming rates

Erosion between the eruptions of the Peach Springs Tuff and Wild Horse Mesa Tuff provides insight into the rapidity with which pediment surfaces were formed. Lack of Peach Springs Tuff beneath much of the flat-lying Wild Horse Mesa Tuff, except near the fringes of the Mid Hills pediment, suggests two scenarios. (1) High topography in much of the region precluded Peach Springs Tuff deposition, but that topography was reduced by erosion to a flat surface before eruption of the Wild Horse Mesa Tuff. (2) The Peach Springs Tuff was deposited across the region on a flat surface, but central uplift (presumably due to intrusion) and erosion preceding the eruption of the Wild Horse Mesa Tuff removed the earlier tuff. The base of the Wild Horse Mesa Tuff is nearly planar, showing no evidence of deposition on a domed surface. Therefore, less than approximately 1 million years (between the eruption of the Peach Springs and the Wild Horse Mesa tuffs) was needed for erosional processes to either develop a pediment across high-relief areas or to restore a domed pediment mantled by resistant tuff. The pediment was restored nearly to its original position and to a planar shape, as indicated by the lack of alluvial deposits under the Wild Horse Mesa Tuff and the local coincidence of surfaces beneath the Peach Springs Tuff and the Wild Horse Mesa Tuff.

Tectonic blocks

The evidence cited above indicates that the compound Mid Hills pediment is a partially stripped early Miocene erosional surface that has been restored after tectonic and volcanic events. The lateral extent of the compound surface outlines a broad tectonic block, extending from the Blind Hills to the New York Mountains, and from the Providence Mountains to the Piute Range, that has undergone tilts of a few degrees from the Miocene to the present. The compound pediment is bounded by the Providence Mountains on the west, but the tectonic block probably includes the Providence Mountains, since most activity on the East Providence Mountains fault zone (Hazzard, 1954) took place during the

latest Cretaceous (Miller and others, 1994; D.M. Miller, unpubl. mapping, 1995). Inferred faults along the west side of the Providence Mountains and the Mid Hills probably acted as one boundary of the tectonic block (Miller and Jachens, 1995). Along the south margin of the Mid Hills pediment, the tectonic block boundary must lie between the Blind Hills and Clipper Mountains, and the Fenner Hills and Old Woman Mountains, as indicated by tilt directions for the Peach Springs Tuff. Faults bounding the Piute Range and the New York Mountains form the eastern and northern boundaries of the tectonic block, respectively. These latter two boundaries are complex and are explained in more detail below. A down-to-the-south tilt of the Mid Hills tectonic block took place after the Wild Horse Mesa Tuff eruption at 17.8 Ma.

During the middle Miocene to Pliocene, the New York Mountains were uplifted with respect to Round and Lanfair Valleys. A deposit containing boulders of Peach Springs Tuff that lies west of Fourth of July Canyon appears to have been derived from northward and higher in the mountains. These relations suggest that Peach Springs Tuff was deposited on part of what is now the New York Mountains, which later was uplifted with respect to Pinto Mountain and Lanfair Valley. The Cedar Canyon fault that lies at the south margin of the New York Mountains accomplished some of this relative uplift. The Cedar Canyon fault juxtaposes the Wild Horse Mesa Tuff to the south against granite to the north, and moved last in the early Quaternary or latest Pliocene, the age assigned to faulted gravel near the mouth of Caruthers Canyon. The New York Mountains had been uplifted with respect to Lanfair Valley by the time that playa deposits were accumulating on the west side of the Piute Range (Fig. 2). These deposits carry clasts typical of the New York Mountains, including the granites and metamorphic rocks of the Caruthers and Sagamore Canyons area (Burchfiel and Davis, 1977). Strong soil development in alluvium above the playa deposits is thought to indicate the playa deposits are older than early or middle Quaternary (Katzenstein and others, this volume). The New York Mountains therefore were uplifted with respect to Lanfair Valley during the middle Miocene to Pliocene.

The Piute Range has acted in concert with the Mid Hills tectonic block. It was uplifted with respect to Piute Valley to the east and, to a lesser extent, to Lanfair Valley to the west (Nielson and others, 1987). Drainages west of the Piute Range were dammed by relative uplift of the Piute Range to form the Pliocene(?) playa sequence (Nielson, this volume; Katzenstein and others, this volume), suggesting that the fault on the west side of the Piute Range was active in the latest Miocene and/or Pliocene. Early and middle Miocene volcanic rocks in the Piute Range are very thin in the south in the vicinity of Homer Mountain (Spencer and Turner, 1985) and thicken northward (Nielson and others, 1987). This relation suggests that the Homer Mountain area was higher than the rest of the range during Miocene volcanism, and that an overall tilt of down to the south took place after volcanism. It is probable that the Piute Range was tilted down to the south at the same time that the entire block outlined by the Mid Hills pediment underwent the same sense of tilting. This relation suggests that the Piute Range acted in concert with the Mid Hills block, and the fault on the west side of the Piute Mountains represents minor block

segmentation. The timing of this tilt event is similar to that for the greater Mid Hills pediment region to the west, constrained as being Pliocene or late Miocene.

Comparison with Pediments in the Region

Several large upland pediments near the Mid Hills, such as Cima Dome and a pediment south of the Granite Mountains (Fig. 1), seem to be inherited from Miocene or older erosional surfaces. Oberlander (1974) noted that many pediments across the Mojave Desert represent stripped Tertiary surfaces that he suggested were surfaces of deeply weathered materials. The information for the eastern Mojave Desert indicates that older, deeply weathered surfaces were early Miocene or older.

The Cima Dome area is a composite upland feature characterized by many interconnecting pediments that form several gentle topographic domes. The Peach Springs Tuff (rhyolite of Sharp, 1957) was deposited on the southeastern side of Cima Dome along a surface that may have been similar to, but inclined slightly steeper than, the present pediment. This old surface may represent an early stage of an early Miocene Cima Dome, or could be the floor of a preceding basin. Pediments and erosional surfaces on other domes predate basalts that are 7 Ma and younger, and many therefore are of Miocene age (Sharp, 1957; Dohrenwend and others, 1984). The primary difference between the Cima Dome area and Mid Hills pediments is topographic shape. The Cima Dome area is a broad upland dome punctuated by smaller domes, whereas the Mid Hills pediment is close to planar. The pediment domes of the Cima Dome region are variously thought to represent end-stage desert erosion as desert domes (e.g., Dohrenwend and others, 1984; Wilshire and others, 1987), features eroded on tectonic warps (Sharp, 1957), or features eroded on tilted fault blocks (R.E. Reynolds, 1995, personal commun.).

A south-sloping planar pediment lies south of the Granite Mountains and bridges the area between the Bristol and Marble Mountains, and Van Winkle Mountain (Fig. 1). The pediment primarily is cut into Late Cretaceous granite of the eastern Granite Mountains (Howard and others, 1987). At Van Winkle Mountain, a flat erosional surface beneath volcanic rocks that predate the Peach Springs Tuff and are stratigraphically equivalent to nearby rocks dated by Miller (1993) as 22 to 19 Ma is moderately tilted to the south in a series of fault-bounded blocks (Miller and others, 1985). The modern pediment adjacent to Van Winkle Mountain averages the elevation of the Miocene erosional surfaces in the tilt blocks. Despite tectonism, the pediment has reestablished itself at about the same position as the early Miocene erosional surface.

Conclusions

Several modern pediments on granite and erosional surfaces beneath volcanic rocks in the Mid Hills region describe a compound, nearly planar surface tens of kilometers in width. The erosional surfaces were first established during or before the early Miocene, before being buried by lacustrine deposits and tuffs during the 19- to 17-Ma interval. These Miocene surfaces grade laterally to modern pediments, suggesting that the pediments are inherited from the older surfaces. Repeated volcanism in the area demonstrates that the pediment was restored after

disturbances, and required less than 1 million years for the process. The compound surfaces define a large tectonic block that, between about 17 and 5 Ma, tilted a few degrees down to the south. Faults bordering this block lie along the west side of the Providence Mountains and Mid Hills, along the east side of the Piute Range, and on one or both sides of the southwestern New York Mountains. The present high elevations of the Mid Hills, New York Mountains, and Lanfair Valley may have resulted from uplift, relative to the Fenner Valley, as part of the tilt block.

Adjacent pedimented regions in the Granite Mountains and Cima Dome areas were also inherited from early Miocene or older surfaces, suggesting that much of the topography of the northeastern Mojave Desert is of that antiquity. Tectonic stability, and possibly climate, in the early Tertiary apparently promoted the widespread development of low-relief surfaces mantled by thick intervals of deeply weathered materials. Many of these surfaces have been stripped to form modern pediments.

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Neogene Structural Evolution of the Woods Mountains Volcanic Center, East Mojave National Scenic Area

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Abstract

Woods Mountains volcanic center (WMVC) evolved through pre-, syn-, and post-caldera phases of activity between 18.5 and 17.6 Ma. Initial eruptions (18.5-17.8 Ma) were of relatively small volumes and from widely distributed vents. They produced a complex volcanic field of high-K trachyte, trachydacite and rhyolite domes, flows, pyroclastic deposits and hypabyssal intrusions. Volcanism culminated with eruptions of the Wild Horse Mesa Tuff ignimbrite, a sequence of metaluminous to mildly peralkaline, compositionally zoned high silica rhyolite ash-flow tuffs. The ignimbrite was emplaced in three major eruptions between 17.7 and 17.8 Ma, producing a large ignimbrite plateau, remnants of which now occur over an area of >600 km². At least two ignimbrite eruptions produced overlapping calderas. Interpretations of gravity and magnetic anomalies and geologic data suggest the system of calderas is ~10 km across and has a funnel like shape ~4 km deep. Subsequent eruptions filled in the caldera, overflowing the rim to the west and southwest. WMVC activity ended with resurgence-like uplift of the center of the caldera at ~17.6 Ma.

Relatively weak Neogene deformation overlaps with and post-dates the formation of the WMVC. Most prominent is a 50-km-long curvilinear belt of faults and open folds extending from the Providence Mountains on the west to Hackberry Mountain on the east. Radiometric dates of differentially tilted volcanic rocks near the Providence Mountains document 9° of east to southeast tilting between 18.5 and 17.8 Ma; an additional 6° tilting occurred after 17.8 Ma. Faulting, folding and tilting of rocks produced 300-500 meters of uplift between 18.5 and ~17(?) Ma of the northern Providence Mountains relative to the Colten Hill - Black Canyon area. In addition, a consistent pattern of caldera-boundary-fault formation suggests that stresses responsible for the curvilinear belt of deformation also influenced caldera formation.

Introduction

Woods Mountains volcanic center is located between the northern and southern Basin and Range provinces in a region of widespread and locally intense Miocene volcanism and tectonism (inset to Figure 1). It is bounded by belts of intense coeval Neogene volcanism and extensional tectonism to the west (Mojave Extensional Belt; Dokka, 1986, 1989; Dokka *et al.*, 1991), to the east (Colorado River Extensional Corridor; e.g., Anderson, 1971; Howard and John, 1987), and to the north (Kingston Range-Halloran Hills Extensional Detachment System; Davis *et al.*, 1993). The WMVC is tectonically and geochemically anomalous in comparison to these surrounding regions in its paucity and low intensity of Neogene faulting, and by the occurrence of large volumes of peralkaline volcanic rocks (McCurry, 1988).

Features of the geology, petrology and isotope geochemistry of the WMVC have been described by McCurry (McCurry, 1982, 1985, 1988), and Musselwhite *et al.* (1989). This paper revises the evolution of the center based upon more recent high-resolution ⁴⁰Ar/³⁹Ar radiometric dating and geophysical work, and elaborates from previous publications on its Neogene structural evolution. Additional work on the geochemical and geophysical features of the center are in progress (McCurry, Mickus and Lux, in preparation).

Woods Mountains Volcanic Center

A simplified geologic map and cross-sections illustrates major features of the center (Figures 1, 2). Generalized aspects of the stratigraphy are illustrated in Figure 3; descriptions of spatially overlapping, older Neogene volcanic and sedimentary rocks can be found elsewhere in this volume (Stop 8 of fieldguide). Volcanism evolved through three principal phases (Figure 4): 1. formation of a precaldera volcanic field, the Hackberry Spring Volcanics (HSV); 2. eruption of relatively voluminous compositionally zoned, dominantly rhyolitic ignimbrites, the Wild Horse Mesa Tuff (WHMT), accompanied by a least two phases of caldera collapse, and 3. extrusion of numerous, relatively small volume rhyolite flows and eruption of tephra from vents within the caldera, the Tortoise Shell Mountain Rhyolite (TSMR).

High-resolution ⁴⁰Ar/³⁹Ar dating of samples from pre-, syn-, and postcaldera components of the volcanic center have recently been completed and are a subject of a manuscript in preparation by McCurry, Mickus and Lux. The new dates indicate a significantly older age for the center than was previously determined by K/Ar dating methods (McCurry, 1988). To briefly summarize the revised ages: 1. Hackberry Spring volcanics ~18.5 to 17.8 Ma; Wild Horse Mesa Tuff, 17.8 to 17.7 Ma; Tortoise Shell Mountain Rhyolite, 17.7 to 17.6 Ma.

Precaldera volcanics accum

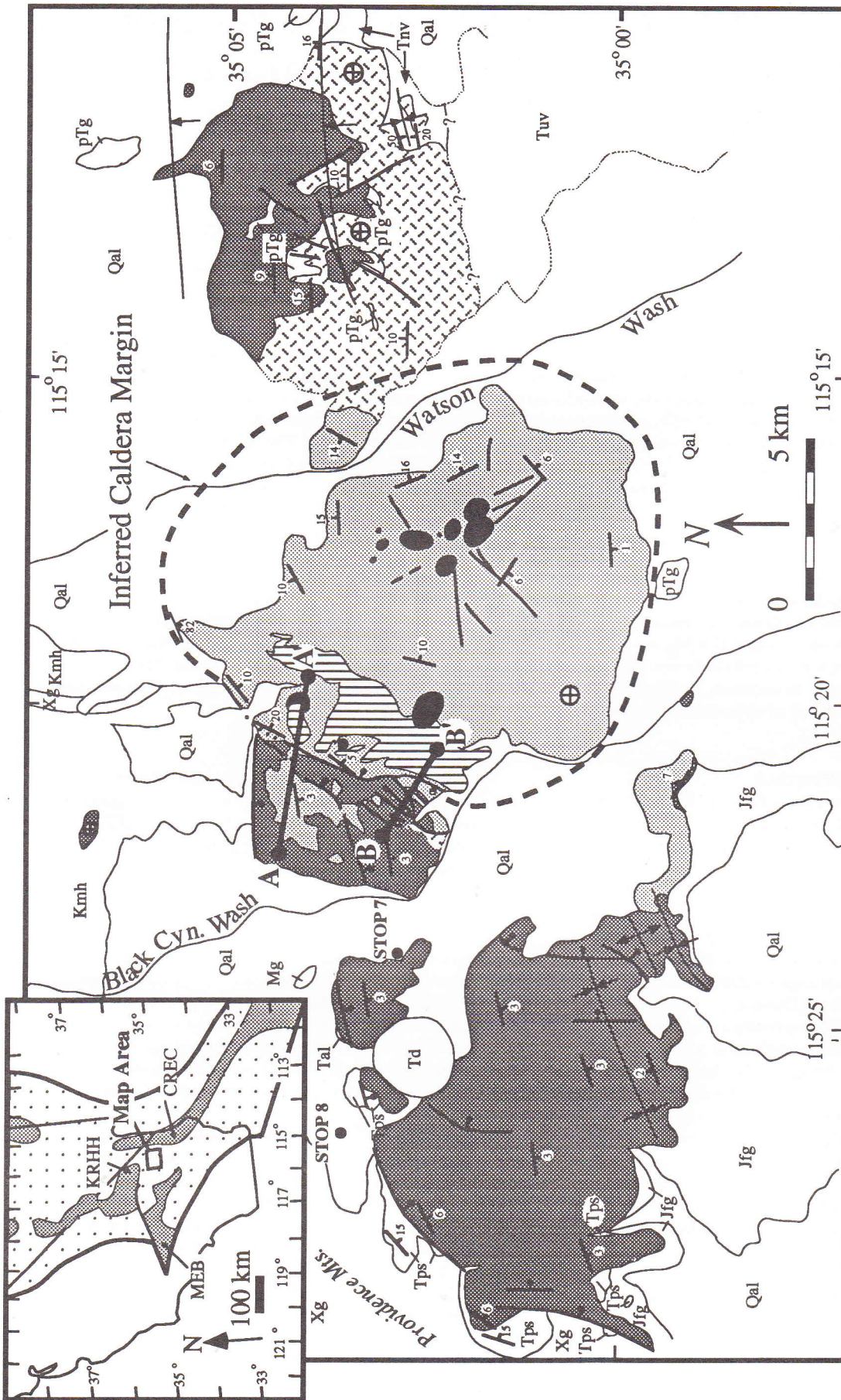


Figure 1. Simplified geologic map of the Woods Mountains Volcanic Center (after McCurry, 1985). Fieldguide stops (Stops 7 and 8) are shown with bold dots. Some units are too small to illustrate on this figure, particularly near Stop 8. Refer to Figure 3, and to the fieldguide for a more complete representation of units at that location. Cross-section locations for Figure 2 are shown with lines AA' and BB'. Cross-hatched pattern - Hackberry Spring Volcanics; dark stipple - Wild Horse Mesa Tuff; light stipple - Tortoise Shell Mountain Rhyolite (TSMR); black - hypabyssal intrusions of the TSMR; vertical lines - basaltic lava flows; Xg - Precambrian metamorphic rocks; Jfg - Jurassic granitic rocks; Kmh - Cretaceous Mid Hills Adamellite; pTg - undifferentiated granitic rocks; Tal - Neogene sedimentary rocks; Tps - Peach Spring Tuff; Tuv - early Neogene volcanic and sedimentary rocks; Tuv - undifferentiated volcanic and sedimentary rocks; Qal - Quaternary alluvium. Inset at top left illustrates the location and geologic setting of the center. State outlines are shown, and regions of Neogene extension are illustrated in a light stipple pattern. Regions of particularly strong extension are shown in a dark stipple pattern (MEB - Mojave Extensional Belt, Dokka, 1989; Colorado River Extensional Corridor, Howard and John, 1987; KRHH - Kingston Range - Halloran Hills detachment zone, Davis, et al., 1993).

Cross-Sections of the Western Margin of the Woods Mountains Caldera

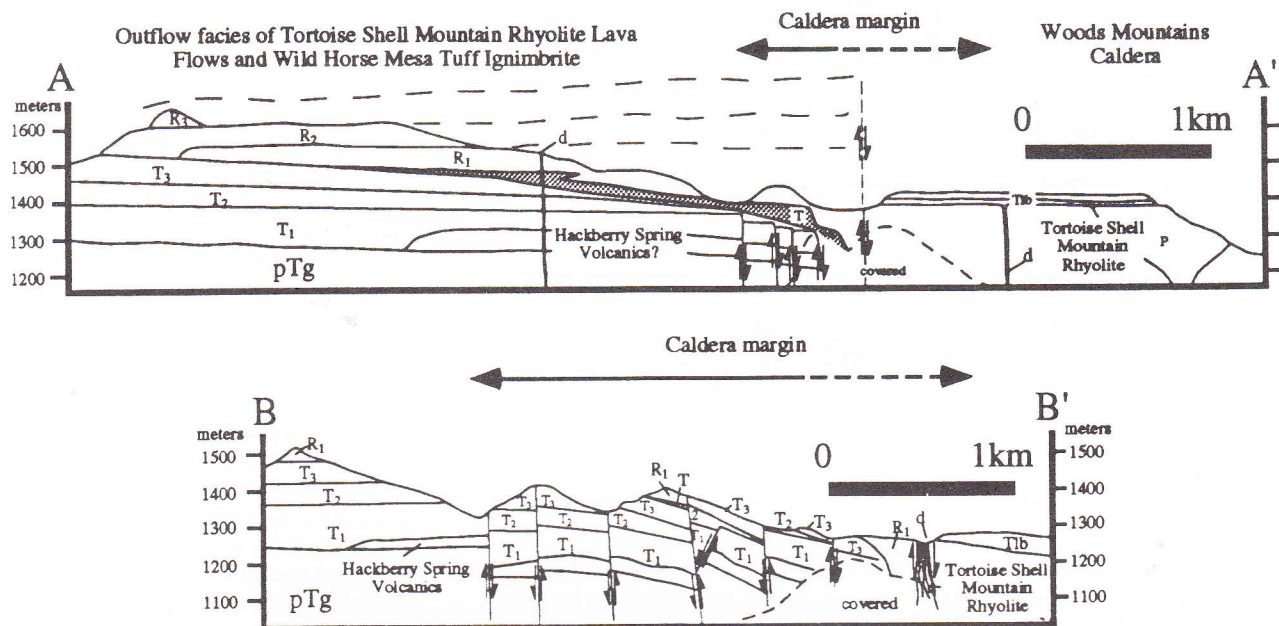


Figure 2. Cross-sections across the western margin of the Woods Mountains caldera. Locations are shown on Figure 1. R_1 , R_2 , R_3 - outflow facies of rhyolite lava flows of Tortoise Shell Mountain Rhyolite; P - a plug-like rhyolite intrusion; T - postcaldera tephra; T_1 , T_2 , T_3 - outflow facies of lower, middle and upper members of the Wild Horse Mesa Tuff (McCurry, 1988); T_{lb} - sequence of basalt lava flows (McCurry, 1985, 1988; Musselwhite et al., 1989); d - feeder dikes for T_{lb} ; pTg - pre-Tertiary rocks, mostly Cretaceous and Jurassic granitoid rocks in this area.

ulated to a thickness of up to 200 meters from vents distributed over a roughly E-W trending area of $\sim 200 \text{ km}^2$. Volcanics consist of interlayered high-K trachyandesite to rhyolite flows, domes, and relatively small-volume ignimbrites. They are intruded in some areas by compositionally similar hypabyssal plugs and dikes, some of which merge upward into flows and domes.

WHMT was emplaced in three major eruptions, producing high-K, dominantly rhyolitic ignimbrites labeled I_1 through I_3 in Figure 2 (cross-sections AA' and BB'). Original volumes of the ignimbrites are estimated at ~ 40 , 20 and 20 km^3 , respectively—corresponding to 23, 14 and 20 km^3 of magma, when density differences are taken into account. Erosional remnants are well exposed over an area of $\sim 600 \text{ km}^2$. Their distribution was strongly controlled by a pre-existing topography which channeled the ignimbrites into four distinct lobes (Figure 4C). The three are distinguished by cooling breaks and distinctive changes in phenocryst assemblage and welding and crystallization zonal features (McCurry, 1985, 1988). Only minor erosion occurred between emplacement of the units, and there are no intervening sediments or lava flows aside from small amounts of localized reworking of the top of I_1 . A reddish soil horizon, and the occurrence of a thin, discontinuous layer of charcoal at the top of I_2 indicates enough time passed between eruptions of I_2 and I_3 for the formation of significant vegetation.

Proximal facies of the WHMT are identified by coarse lag deposits at the base of many flow units, and by an eastward

exponential increase in the maximum size of dense lithics within individual flow units toward the inferred caldera margin and vent zones (McCurry, 1989). Vents are not directly exposed because they are covered by later intracaldera volcanic rocks. However, outflow facies of ignimbrites I_2 and I_3 are apparently exposed to within a few hundred meters of their source vents because of the extremely large sizes of some of their intrained lithic fragments. The lithics consist mostly Mid-Hills Adamellite (Cretaceous) and accessory rhyolite; detailed geologic mapping of the area suggests they must have been derived by ejection from the subsurface (McCurry, 1985, 1989). They include some of the largest lithics ever documented—14-20 meters across!—in the outflow facies of an ash-flow tuff (McCurry, 1989). Proximal ash flow tuff sheets overlap a system of caldera-related faults and buttress-unconformities formed by faulting of ignimbrite members I_2 , I_3 and older rocks along the western caldera border zone (Figure 2, cross-sections AA' and BB').

Rhyolite flows and tephra deposits of the TSMR were extruded in rapid succession after unit I_3 of the WHMT, overflowing from within the caldera to cover outflow sheets of the WHMT for $\sim 3 \text{ km}$ to the west (Figure 2, cross-section AA'). Delicate features at the top of ignimbrite I_3 are well preserved suggesting the overlying flow, labeled R_1 in cross-section AA' of Figure 2, was emplaced almost immediately after the ignimbrite. This figure illustrates that the flow can be traced laterally from west to east across near vertical buttress unconformities formed by earlier caldera collapse.

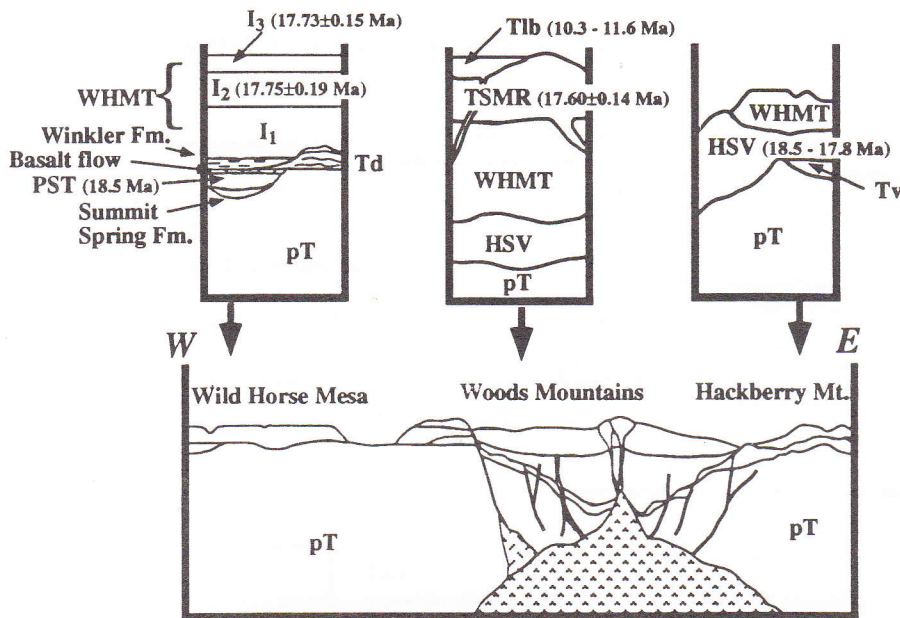


Figure 3. Diagrammatic cross-section and stratigraphy of the Woods Mountains Volcanic Center. pT - undifferentiated pre-Tertiary rocks; Tv - undifferentiated pre-WMVC, early Neogene volcanic and sedimentary rocks; SSF - "Summit Spring Formation" (see Stop 8 of fieldguide, this volume); PST - Peach Spring Tuff (date from Nielsen, et al., 1990); WF - "Winkler Formation" (see Stop 8 of fieldguide, this volume); Tb - basaltic lava flow; Td - dacite dome; HSV - Hackberry Spring Volcanics (⁴⁰Ar/³⁹Ar dates from McCurry et al., in preparation); WHMT - Wild Horse Mesa Tuff, subdivided into members 1 - 3 on left column (⁴⁰Ar/³⁹Ar dates from McCurry et al., in preparation); TSMR - Tortoise Shell Mountain Rhyolite; Tlb - basaltic lava flows (dates from Musselwhite, et al, 1989).

These flows were subsequently down-faulted to the east in another major phase of caldera collapse. Extrapolation of flows R₁ to R₃ from west to east requires at least 500 m of down-drop to the east, since none of these are exposed east of that fault. The caldera subsequently was infilled by a combination of numerous lava flows, pyroclastic deposits and, near the western caldera border zone, by caldera-scarp breccia. Some of these flows overflowed the caldera boundaries again, this time to the north and southwest.

Final phase of WMVC activity consisted of emplacement of rhyolite plugs and domes near the center, and near the western border zone of the caldera. The plugs are identical in composition and mineralogy to the rest of the nearly chemically homogeneous caldera fill (McCurry, 1985). Intrusion and extrusion of plugs and domes near the center of the caldera structurally uplifted the central part of the caldera (McCurry, 1988), tilting older flows outward and producing a series of radiating faults (Figure 1).

Three mafic lava flows disconformably overlie the TSMR within the western part of the caldera but are apparently not directly genetically connected to the WMVC (McCurry, 1988).

Neogene Structural Evolution

Neogene structures in the WMVC area are divided into three temporal and genetic groups: 1. precaldera (> 18.5 Ma); 2. WMVC-related (18.5 - 17.6 Ma); 3. post-WMVC (<17.6 Ma). Precaldera structures are exposed in a small erosional window through the HSV in the eastern part of the

WMVC volcanic field (T_{nv} in Figure 1). Here older epiclastic rocks are folded into a relatively tight syncline with a near vertical axial plane, and a horizontal fold axis which trends roughly east to east-northeast. Moderately to steeply dipping interbedded layers of multilithologic breccia, and arkosic and volcanoclastic sandstone are unconformably overlain by nearly flat-lying HSV. Relatively strong folding and deep erosional dissection of these rocks contrasts sharply with the conformable and weakly deformed nature of the overlying HSV, and younger volcanic rocks. However, the roughly E-W trend of the fold axis parallels later structural features of the WMVC.

WMVC-related structures are dominated by a one- to two-kilometer-wide zone of north to northeast to east trending high-angle normal and reverse faults and associated buttress unconformities in the western Woods Mountains (Figure 2). Many of these faults are remarkably well exposed; related buttress unconformities are mantled by ignimbrite members in a manner which clearly records multiple episodes of caldera collapse (McCurry, 1982, 1985, 1988). At least three diachronous periods of caldera-related faulting occurred within the zone. They correlate with eruptions of ignimbrites I₂, I₃ as well as with post-ignimbrite lava flow extrusion (McCurry, 1985, 1988). Caldera collapse is also inferred to have coincided with eruption of I₁, but these structures are covered by younger rocks. Many of the faults and scarps exhibit a "typical" pattern of stair-step-like downdrop towards the caldera (e.g., Figure 2, section AA'). However, others exhibit the opposite sense of offset (i.e. up-to-the-east). A well-exposed system of such reverse faults along the western caldera border zone produced a net eastward down-drop of ~100 meters over a horizontal distance of ~1 km by incremental eastward tilting of fault blocks (Figure 2, section BB').

The wide zone of faulting on the west, and westward tilting rather than faulting on the east side of the caldera (McCurry, 1985; 1988) suggests the caldera formed primarily by a sagging mechanism (e.g., Walker, 1984), which was accompanied and enhanced by faulting along the west and northwestern border zones. This interpretation is supported by three-dimensional geophysical modeling of gravity and magnetic anomalies associated with the center (McCurry, Mickus and Lux, 1995, in progress). Faulting along the west and north sides of the caldera produced a distribution of structures whose overall geometry parallels that of later, spatially more extensive faults and folds in the area, suggesting a possible influence of a regional stress field in the development of the caldera (e.g., McCurry, 1988).

Postcaldera structures are dominated by high-angle

normal faults, open folds and a large monocline. Many of these structures vary systematically in their orientations. On the west (northern Providence Mts. and Wild Horse Mesa) faults and fold axes trend to the north and northeast; a system of en echelon faults bounding the north end of Wild Horse Mesa (Barber Canyon Fault Zone, McCurry, 1985) appears to merge to the southwest with the north trending East Providence Fault, a major range bounding fault (Goldfarb, *et al.*, 1988). Trends of faults in the Barber Mesa and western Woods Mountains is east-northeast. Further to the east, Hackberry Mountain is dominated by a east trending monoclinical fold. The fold is overprinted by numerous relatively small normal faults with no obvious systematic orientation (McCurry, 1985).

Northeast to east trending faults and folds are anomalous in this region. Strongly extended regions to the west, north and east are dominated by north-northwest to north trending faults. However, Spencer (1985) documents a system of east trending dikes in the Homer Mountains area adjacent to the east margin of WMVC, of about the same age. Spencer proposes mechanical models for producing localized changes in stress fields. These include crustal flexing during isostatic rebound, and secondary divergence associated with either diachronous extension or variations in extensional strain rates along the Colorado River Extensional Corridor. Is it possible that the pattern of structures observed across the WMVC reflects a transition in the orientation of the local stress field?

Uplift of the Northern Providence Mountains - Mid Hills area

Although there is no evidence of Quaternary faulting along the eastern flank of the Providence Mountains, exposures of Neogene epiclastic and lacustrine sedimentary rocks ("Summit Spring" and "Winkler Formations"; STOP 8 of field guide, this volume) and volcanic rocks suggest active uplift occurred there during the early Miocene. The oldest exposed volcanic rocks are correlated with the ~18.5 Ma Peach Spring Tuff (D. Miller and K. Howard, USGS, personal communication 1985; Nielsen, *et al.*, 1990). This is a very widespread ignimbrite probably derived from a source in the southern tip of Nevada (Hillhouse and Wells, 1991; Buesch, 1992). Near the crest of the Providence Mountains the tuff is up to 70 meters thick, dips about 15° to the southeast, and overlies a thin bed of fanglomerate consisting of coarse carbonate and granitic lithics (McCurry, 1985; Stop 8 of fieldguide, this volume). The thin fanglomerate and absence of evidence that the Peach Spring Tuff pinches out across the crest of the range, suggests the northern Providence Mountains were much more topographically subdued at 17.8 Ma than today.

Up to 60 meters of lacustrine sediments, dominantly limestone and partially dolomitized limestone, conformably overlie the Peach Springs Tuff. These in turn are overlain by the WHMT which dips 6° to the southeast. The angular

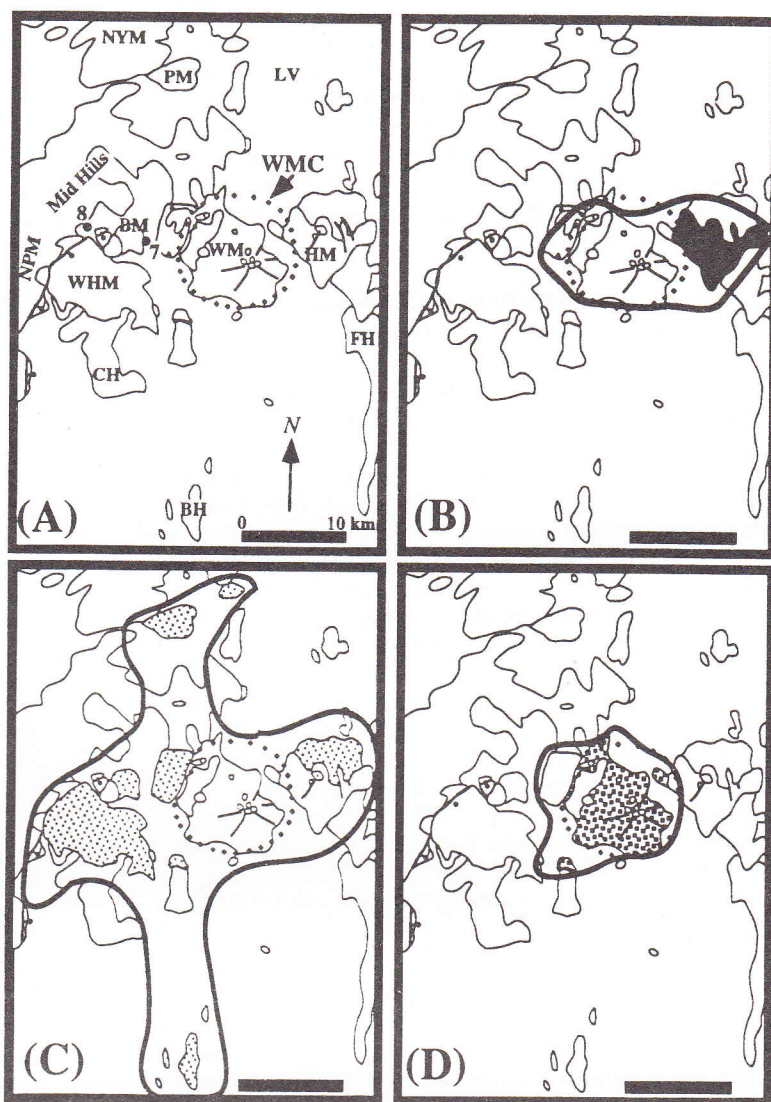


Figure 4. Evolution of the Woods Mountains Volcanic Center. Frame (A) - lines bound major ranges and mesas; NPM - northern Providence Mountains; NYM - New York Mountains; PM - Pinto Mountain; TM - Table Mountain; LV - Lanfair Valley; HM - Hackberry Mountain; FH - Fenner Hills; BH - Blind Hills; CH - Colten Hills; WHM - Wild Horse Mesa; BM - Barber Mesa; Stops 7 and 8 of the fieldguide (this volume) are illustrated with large filled circles; the outline of the Woods Mountains caldera (WMC) is dotted. Frame (B) - precaldera evolution of the WMVC, black color - outcrop areas of the Hackberry Spring Volcanics. Frame (C) - caldera phase of the WMVC and emplacement of the Wild Horse Mesa Tuff; outcrop areas of the ignimbrite are stippled. Frame (D) - postcaldera evolution; gravel pattern represents outcrop areas of the Tortoise Shell Mountain Rhyolite. In frames (B) through (C) the inferred original extents of the respective units are shown with bold lines.

discordances suggests that the older Neogene rocks were tilted 9° to the southeast between 18.5 and 17.8 Ma. The remaining 6° of tilt occurred after 17.8 Ma. Additional faulting, folding and tilting between the northern Providence Mountains and Black Canyon Wash, ~15 km to the southeast, produced a cumulative post-WHMT uplift of the northern Providence Mountains of between 300 and 500 meters, relative to Black Canyon Wash. Although these tectonic features are of relatively minor magnitude, they

demonstrate that the region was undergoing active deformation during the early to mid Miocene. They are also geometrically consistent with the type of uplift required to produce the current topographic expression of the northern Providence Mountains and Mid Hills areas. Lacking evidence for significant recent faulting, this may indicate that at least some of the current physiographic features took their form at that time.

Summary and Conclusions

Miocene rocks of the Woods Mountains Volcanic Center underlie much of the distinctive landscape across the southeast part of the East Mojave National Scenic Area. Relatively undeformed and well exposed, these rocks record a unique part of the structural and volcanic evolution of the eastern Mojave Desert. Unique features include the best example of a caldera system in the Mojave Desert, and one of the best exposed examples of a caldera margin and proximal outflow facies of genetically related ignimbrites observed anywhere. The center produced large volumes of silicic, mildly peralkaline rhyolite magmas not observed elsewhere in the region (McCurry, 1988; McCurry, et al., in progress). The center is also characterized by a distinctive style of tectonism suggesting the existence of an anomalous extensional stress field both during and shortly after evolution of the center. Related deformation resulted in at least some component of uplift of the Providence Mountains—Mid Hills region. Therefore many of the present day physiographic features of this region appear to have been produced in a relatively brief but intense episode of tectonic and volcanic activity between ~18.5 and ~17.5 Ma.

Acknowledgments

This paper is dedicated to the memory of Jane and Fleet Southcott, past owners of the Gold Valley Ranch. Jane and Fleet had an intense interest and love for all aspects of the Woods Mountains. Their cheerful enthusiasm, interest and support for our work were greatly appreciated. Thoughtful reviews by Paul K. Link and Scott S. Hughes greatly improved the manuscript.

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The Rustler Rockshelter Site

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In the summer of 1992, an all-volunteer crew from CSU Bakersfield (CSUB) and other institutions undertook test-level excavations at Rustler Rockshelter. The purpose of the work was to obtain additional and more detailed information to augment the 1958 excavations by UC Berkeley. The Rustler Rockshelter site (CA-SBR-288) is located on an alluvial terrace between a large cliff and an intermittent wash at an elevation of 1,250 meters (4,125 feet). The entire site measures about 80 meters long (E/W) and 30 meters wide (N/S). A very dark midden (about 50 by 20 m) is visible over much of the terrace and a large number of artifacts and fire-affected rock lie on the surface. The site is not a true rockshelter but is an open midden located against a south-facing cliff. The cliff is composed of a volcanic ash or mudflow containing a variety of other materials including small obsidian nodules (Apache tears).

The site originally was recorded by the University of California, Berkeley (UCB) early in 1958 during a "reconnaissance" inventory of the region. Upon its discovery, two small units were excavated, the results of which led the investigators to believe there was a "culturally stratified" deposit present. In the fall of 1958, additional excavations (a total of seven 5 x 5 ft units) were undertaken to obtain more data. Considerable cultural material was recovered, including ceramics, ground stone, 34 projectile points, bifaces, cores, hammerstones, and faunal remains. The UCB work at the site was reported by Davis (1962) and it quickly became the "type" site for the eastern Mojave Desert.

The Late Culture History of the Providence Mountains

Based on the excavations at Rustler Rockshelter (Davis 1962) and Southcott Cave (Donan 1964; also see Sutton et al. 1987), Donnan (1964) formulated a basic culture history (chronology) for the Providence Mountains. The latter part of this chronology was called the Providence Mountains Complex (see Davis 1962) and included three phases: I) a pre-ceramic Yuman (to A.D. 800); II) a ceramic Yuman (ca. A.D. 800-1400); and III) a Shoshonean Horizon (A.D. 1400-1850). Warren (1984:395) noted that the cultural sequence of the Providence Mountain area diverged "from that of the northeastern Mojave Desert at the end of Amargosa I" (early Rose Spring Period). Warren (1984:395) believed that the "Providence Complex, possibly preceded by a 'nonceramic Yuman' assemblage, appears to represent the Hkataya influence in the southeast Mojave Desert."

Research Design

The CSUB work at Rustler Rockshelter was undertaken with a variety of objectives in mind. The first goal was to obtain a greater and more detailed sample of cultural materials. Improved field methods and recently developed

special studies enable us to gain a greater understanding of the site. More specifically, we wanted to obtain materials suitable to: 1) better delineate the various culture history proposals; 2) consider questions of ecology; 3) address the question of the Desert Mohave and the Numic Expansion; 4) the classification, distribution, and dating of ceramics; 5) environmental change; and 6) settlement/subsistence patterns.

Results

Soils and Stratigraphy

Davis (1962:28) described two basic soil strata from the UCB excavations: 1) a "dark greasy" midden and 2) sterile gravels. We recorded four relatively clear soil strata during the 1992 excavations. The upper three contain cultural remains while the fourth is a culturally sterile layer. The "dark greasy" midden of UCB includes our upper three strata (1-3) while our Stratum 4 corresponds to the sterile gravels described by Davis (1962:28).

Features

Five features (four distinct, one possible) were identified in the deposit: three hearths, one possible hearth, and one concentration of rock, mostly fire-affected.

Material Culture

The ground stone collection from CA-SBR-288 is relatively small and consists of grinding, crushing, pulverizing, and rubbing or polishing tools. Two complete metates, 35 metate fragments, two complete manos, eight mano fragments, one pestle, and five stone beads have been recovered.

Considerable flaked stone was recovered, including points, bifaces, cores, and large quantities of debitage. Sixty-one projectile points have been recovered, including Pinto, Elko, Humboldt, Gypsum, Rose Spring, and Desert series forms. A large number of bifaces and cores were found, along with smaller number of unifaces, drills, and hammerstones. A very large number of flakes was found in the excavations, we believe about 250,000 from the 10 units.

A total of 609 ceramic pieces was recovered, tentatively classified into two major wares, Lower Colorado Buff Ware and Tizon Brown Ware (Intermountain Ware). Other artifacts from the site include a number of shell beads and a cane fragment.

Faunal and Floral Remains

A large quantity of faunal remains was found. The formal analysis is not yet complete but known categories include artiodactyl, lagomorph, rodent, lizard, tortoise, and bird. In addition, several shell fragments of *Anodonta* sp. were found. These pieces may represent food refuse, the shellfish being obtained along the Colorado River to the east or perhaps the Mojave River to the northwest. While we are sure that considerable floral materials will be recovered from the various samples, none is so far known.

Obsidian Studies

Of the 31 samples, 17 were sourced directly to the Rustler Rockshelter area (named the Providence Mountains source, the type specimen being removed from the wall of the rockshelter itself), one to Mono Glass Mountain, one to the Coso Volcanic Field, nine to "Unknown 1," two to "Unknown 2," and one to "Unknown 3." Assuming that the unknowns are relatively local float sources, only two of the 31 samples come from outside the eastern Mojave Desert region. The hydration rims of 29 of the specimens could be measured. The hydration values from the obsidian range from 1.0 to 9.8 microns.

Dating

The dating of the occupation of the site currently rests on marker artifacts and obsidian hydration analyses; the radiocarbon results will be available soon. Based on the artifacts, occupation of the site may have begun during the Pinto Period, perhaps as early as 7,000 BP. Gypsum Period and Rose Spring Period points also are present. Desert series points are the most abundant and that, coupled with the plentiful ceramics, suggests an intensive occupation during the Late Period. The obsidian hydration values are in general accordance with the artifact data, except that there is no strong indication of a late component.

Discussion

The UCB excavations resulted in the delineation of "ceramic" and "pre-ceramic" components at the site, leading to the formulation of the culture history still in use. From the CSUB work, it is clear that there is such an organization to the site but it is horizontal as well as vertical. The internal organization of the site is quite complex.

One of the surprises from the 1992 excavations was the quantity of materials related to lithic reduction activities; the large number of bifaces, cores, hammerstones, and debitage. Lithic workshop activities were a major function of the site, people obtained their stone from the nearby slopes and then worked by the cliff, protected from wind, rain, and temperature.

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A Brief Summary of the History of Mining in the East Mojave Desert, 1863-1947

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The 1860s

By late 1859 the Mojave Road was established as a viable wagon road between Los Angeles and the Colorado River. As early as 1861 miners working El Dorado Canyon, on the Colorado River, used the route to a point about six miles east of Rock Spring where the trail to El Dorado Canyon took off. Early in 1863 copper ore was discovered, probably by soldiers from Fort Mojave, in the Dead Mountains 25 miles southwest of Fort Mojave. And the Irataba Mining District was established - named after the Mohave Chief.

In light of the presence of prospectors passing through the east Mojave, the discovery of silver ten miles west of Rock Spring, nearly astride the Mojave Road in 1863 is not surprising. With the March 12 discovery by Charles Hamilton and Francis Austin, the Rock Spring Mining District was established the following month. The district embraced Macedonia Mountain. The townsite of Providence was laid out, and consisted of a string of stone cabins and tents. By March 1864 five mining companies with interest in the district were listed in the San Francisco business directory, but little true work was conducted.

Indians killed a prospector, Moses Little, June 12, 1866 creating another kind of rush of prospectors - out. Camp Rock Spring was established December 30, 1866 by the U.S. Army to protect mail carriers on the Government Road, and slowly prospectors returned.

The 1870s

Just to the north in the New York Mountains, the New York Mining District was organized April, 1870. In June 1872 Matt Palen erected a smelter and in August 1873 the Elgin Mining Company of Elgin, Illinois hauled a mill to the New York Mine. By December the mill was running but was shut by May 1874.

The remainder of the 1870s were quiet in the East Mojave Desert, that is until the

discovery of silver at the Bonanza King Mine set off a flood of prospectors who in turn made new discoveries.

The 1880s

George Goreman and Pat Dwyer, prospectors from Ivanpah in the Clark Mountains, discovered rock that assayed phenomenal silver values in the spring of 1880. By spring 1881 J. D. Boyer and H. L. Drew, San Bernardino Business men acquired the mine, and by December J. B. Osborne and N. Hasson had joined them. However, the investment needed to turn the ore deposit into a mine was still a daunting obstacle.

According to one account, Thomas Ewing (a mining scout for George Hearst) asked if it was for sale, and at what price. "There are four of us," was the reply, "and we want \$50,000 each." "It's a sale," said Ewing, as quick as a flash, and securing a piece of rough wrapping paper, he immediately drew a draft for the amount. It read: "George Hearst, San Francisco; pay to bearer \$200,000." Within forty-eight hours the draft was paid through Wells Fargo.

A 10-stamp mill was hauled from Mojave costing in all over \$50,000 in July 1882, and a hoist was shipped via Colton. Up to 150 were men employed sinking the shaft and preparing for mill. The mill commenced operations January 1883. At the mill a town named Crow Town grew



Figure 1. Manvel (later Barnwell) before the construction of the railroad to Searchlight in 1907. San Bernardino County Museums collections.

up at what is the site of Domingo spring. But there also was Providence near the mine, whose structures were constructed of quarried volcanic tuff. On March 11, 1885 the mine and mill were closed down by the owners. When they reopened a week later the labor force was paid \$3 a day, fifty cents less. The mill burned July 21 1885. The mine had produced \$1.8 million in silver.

At the nearby Kerr Mine, experienced mining man Godfrey Bahten constructed a five stamp mill which started in January 1887.

A short distance south of Providence at Hidden Hill, gold was discovered as early as 1882, the Arrow Mining district established, and ore milled in an arrastra.

The 1890s

The early 1890s saw substantial mining activity in the East Mojave desert.

In the early 1890s Isaac Blake purchased the New York Mine, constructed the Needles Reduction Works and by 1893 had 80 men developing the mine. But, the panic of 1893 silenced this silver mine. However the one remnant of Blake's empire was the Nevada Southern Railway. By August 1893 the railroad reached Manvel. Manvel (later renamed Barnwell) served as the distributing point for a vast area.

At the same time that all of Blakes's energy was being poured into the East Mojave, the ephemeral but thriving town of Vanderbilt literally sprang up overnight. Gold was discovered by Paiute Indian Bob Black in January 1891. In 1892 M.M. Beatty, a relative of Black, staked the first claims. Beatty and Allen Green Campbell began developing the Boomerang Mine. Simultaneously, two miners from Providence, Richard C. Hall and Samuel King, filed claims which became the Gold Bronze mine. Two other Providence miners, Joseph P. Taggart and James H. Patton, joined them in June 1892. By the time of a significant strike by Taggart that fall, the camp consisted of perhaps 300 men and a camp consisting of a store, boarding house and several saloons. In March 1893 Campbell's mill started, and the Gold Bronze mill was operating by May 1894.

In April 1894 water was struck in the Gold Bronze shaft, and in June the Boomerang mine hit water. Upon hitting the water, the character of the ore changed and the mines were unable to recover the gold. The mines and town began their decline and by 1897 were essentially finished.

At Hidden Hill in the south portion of the Providence Mountains, some rich highgrade gold was discovered February 1894, and a 2-stamp mill was erected 1895 by Monaghan and Murphy, two Needles merchants.

The 1900s

The Nevada silver discovery at Tonopah in 1900, and the gold discoveries at Goldfield (1903) and Rhyolite (1904) stimulated renewed prospecting throughout Nevada and Eastern California. Within the east Mojave numerous older mines were reactivated in the wake of these discoveries, and some new discoveries were made.

Gold discovered in the western Castle Mountains on December 19, 1907, transformed the area into the thriving town of Hart. In January there was a stampede to Hart with people leaving Needles and Searchlight "in

automobiles, buggies, wagons and on bicycles and burro." Many came from Goldfield. By the end of the month there was telephone service, and an estimated population of 200. In May 1908 a small mill capable of processing 8 tons (a day?) was installed by the Big Chief mine. The Oro Belle Mine with a main shaft eventually 860 feet deep, produced very little gold, and never constructed a mill. Much of the town of Hart burned December 1910.

In late summer 1908, high grade gold was found 28 miles southwest of the new boom town of Hart. This discovery, the Lost Burro, was made by D. G. Warfield and Mark Neumayer. The townsite of Gold Valley was laid out. At the mine a 100 foot deep shaft was sunk, ore was worked in an arrastra. By 1910 a small stamp mill was operational.

On the north end of Gold Valley the mine camp of Out West drew some attention. In 1909 the camp consisted of a stone house and 3 frame-tent houses.

The New York Mine was revitalized in 1907 as the Sagamore Mine by N. P. Sloan. A mill with the capacity of 50 tons per day was erected in 1908 - but shut down after 6 weeks.

In 1906 there was a serious, yet failed attempt to reactivate the Bonanza King Mine. A ten stamp mill was erected. The mine operated one year. There were additional attempts in the late 1910s and in 1924.

The California Gold and Copper Company led by the able hand of Albert Cram of Riverside, appears to have been little more than a stock scam. Cram filed the first mining claims southeast of Vontrigger Spring in 1902 at the height of a copper rush that lasted until 1907. A huge mill was constructed as was a company town with homes for the miners. Investors continued to be milked until 1915 when operations were suspended.

In the old Rock Spring District, the Macedonia Mine (one of the original locations) was renamed the Columbia. In December 1910 a 5-stamp mill was installed

In 1913 there was yet another attempt to reactivate the gold mines at Hidden Hill. A substantial camp was erected and some beautiful pockets of free gold were found. All told up to 1920 perhaps \$100,000 had been recovered from the mines here.

In 1925 a camp named Vontrigger sprang up at the Getchell mine, owned by J. L. Workman and Senator Getchell of Nevada. This camp located in the Hackberry Mountains consisted of 30 tents. The mines were worked intermittently in the 1930s and early 1940s.

The biggest thing to happen in the east Mojave was the Vulcan iron mine in Foshay Pass. The claims here were patented as early as 1902 with only minor development. There was a small camp here in the 1920s. However between December 1942 and July 1947 over 2.6 million tons of iron ore was mined and shipped to the Fontana steel mills. A camp was constructed to house 65 men near the mine and another 35 men lived with their families in trailers at Kelso.

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Wild Horse Mesa Mule Trails

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Introduction

Wild Horse Mesa is the largest volcanic mesa in Southern California. Its table land can be seen east of the Providence Mountains from which it is separated by Barber and Beecher canyons. It lies south of Macedonia Canyon and Columbia Mountain and west of Wild Horse Canyon, Hole in the Wall, and Black Canyon Road. Deep canyons cut the volcanic rocks of the mesa, which covers about 14 square miles. McCurry (this volume) describes the Peach Springs Tuff and the Wild Horse Mesa Tuff that form the flat surface and resistant cliffs of the mesa.

Wild Horse Mesa ranges in elevation from 4800' on the south to 5610' at its north end. The south end holds a playa surrounded by creosote and Mojave yucca. The middle of the mesa contains Joshua trees and cat claw. The north half of the mesa is covered by piñon, juniper, and agave. It was this woodland that was harvested before the turn of the century to fuel mining operations in the Providence Mountains.

Mining north of the mesa started with the formation of the Rock Springs mining district in April, 1863. Work progressed slowly, in part because of a lack of workers, but shafts and tunnels were dug at several mines. By November, three separate districts were recognized: Rock Spring, Macedonian, and Silver Hill. The district was abandoned



Figure 1. Wild Horse Mesa and Beecher Canyon (left). R.E. Reynolds photograph.

when a miner was killed by an Indian at Rock Spring in June, 1866. The district remained idle until 1871, after army redoubts had been established along the Government Road and calm restored (Casebier, 1987; Vredenburg and others, 1981). In 1872 a smelting works was established in the Macedonian district and supplies were hauled by wagon from San Bernardino. Fifteen tons of ore were shipped in September of that year. About the turn of the century, the Macedonia mine was renamed the Columbia mine. The

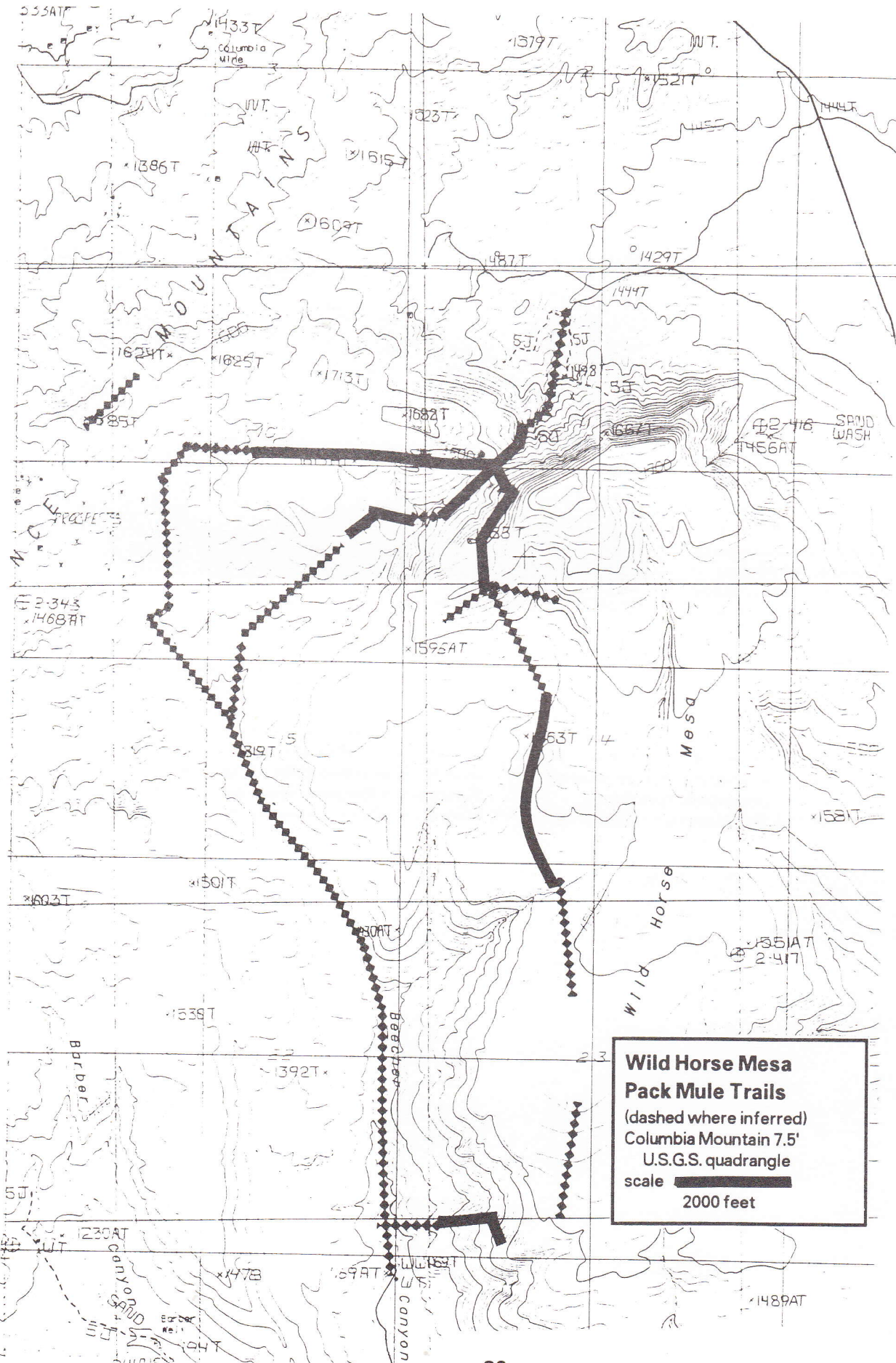
Columbia, with a five stamp mill, operated in 1910 - 1911 and again in 1935- 1936. Nearby, the Francis Copper mine was open in 1917- 1918 and in the 1930s (Vredenburg and others, 1981).

The Columbia mine is just one mile north-northwest of the north end of Wild Horse Mesa; Macedonia Canyon and Columbia Mountain are 1.5 miles and two miles north, respectively. The Globe and Providence mines are 1.5 miles west of Wild Horse Mesa and approximately one mile north of Summit Spring (Wright and others, 1953). And the most productive mine in the area, the Bonanza King is 15 miles south of the old Macedonia district and only about three miles west of the south end of Wild Horse Mesa.

The silver lodes that were to become the Bonanza King were located in 1880. There was a flurry of claims, mining, recapitalization, and partnerships through 1881 and into 1882. In July, a ten-stamp mill was freighted to the Bonanza



Figure 2. Switchbacks on the Wild Horse Mesa Trail leading up to the mesa top. R.E. Reynolds photograph.




**Wild Horse Mesa
Pack Mule Trails**
(dashed where inferred)
Columbia Mountain 7.5'
U.S.G.S. quadrangle
scale 
2000 feet



Figure 3. Summit Trail, view toward Providence Mountains. R.E. Reynolds photograph.

King from Mojave, and between 100 and 150 men were actively employed (Vredenburg and others, 1981). Called in 1884 the "richest silver mine in California" (Myrick, 1963), the mill shipped bullion that in a total of 28 months was valued at about \$1,700,000.00 (Vredenburg and others, 1981). The mill burned to the ground in July 1885, but the nearby Kerr mine operated a five-stamp mill from 1885 to at least 1890. The Bonanza King was reactivated in 1906 and 1907 with a gasoline-powered ten-stamp mill. The mine and its dumps continued to yield silver in 1915-20 and 1923-24 (Vredenburg and others, 1981).

The productivity of the Macedonian district and the Bonanza King mines in the 1870s and 1880s demanded wood for mining timbers and charcoal for ore reduction. These materials could readily be obtained by mule trains from the flat surface of Wild Horse Mesa reached by trails through Beecher Canyon. A network of trails for wood and charcoal hauling was developed across the northern and western portions of Wild Horse Mesa. A supply cabin or overnight waystation was built near the north end of the mesa.

The mule trains could find water at Rouse Well, one-half mile north of the mesa. Beecher Well, two miles up Beecher Canyon is of uncertain date, but was probably an historic source of water. Rouse Well, actually a pair of wells, is 1.25 miles southwest of the Wild Horse Canyon Road which leads to the Macedonia Canyon road. A road runs west-northwest from Fenner to Bonanza Well. What is now the Black Canyon Road is shown in 1908 (Mendenhall, 1909) as a branch road running north from the Fenner Road through Wild Horse Canyon, continuing a short distance down Macedonia Canyon, then to the Government Road west of Cedar Canyon.

Mendenhall's description, however, indicates that this route is not accurately drawn. A more straightforward road was in use by 1917 (Thompson, 1921) which allowed passage through Wild Horse Canyon, Macedonia Canyon, and then to the railroad sidings of Ames or Elora on the Los Angeles and Salt Lake railroad.

There were also undeveloped water sources for the mules. Geologic mapping (McCurry thesis) shows major landslide structures in the canyon wall west of Beecher Well and these structures may have caused a spring to flow near the canyon bottom. Personal observations over the last 20 years indicate that Wild Horse Mesa itself is very wet in the spring of most years and after monsoonal summer rains. After spring snows and rains, the creeks flow, ponds fill, and soil is hard to walk on because it is so muddy. Summer rains fill natural tanks and eroded volcanic basins to provide water

for deer, big horn sheep, porcupine, coyote, and mountain lions.

The Trails

The Wild Horse Mesa mule trails are described using informal geographic names used by campers, hunters, and from discussion with local residents including T. More and the late James Winkler of Needles.

Wild Horse Mesa Trail

The Wild Horse Mesa trail probably started at Rouse Well at the northern foot of Wild Horse Mesa. It is partially obscured by bladed roads up the north slope of the mesa. The trail runs to the mesa saddle where it branches. It switchbacks steeply upslope to the crest of the mesa and then runs 1/3 mile southwest to reach its maximum elevation at 5420'. Mule traffic has cut down a path through the desert varnish and white tuffs about an inch deep. From



Figure 4. Juniper wood pile along the Summit Trail. R.E. Reynolds photograph.

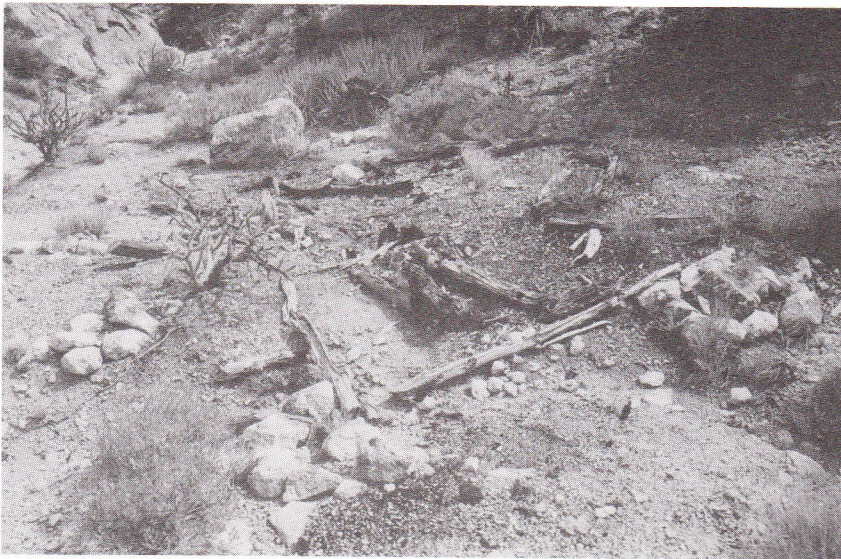


Figure 5. The charcoal burning area on Summit Trail, showing four rock piles. R. E. Reynolds photograph.

this high point, you can look over the flat terrain of the mesa and west where the Bonanza King mine and Mitchell Caverns are visible at the southeast base of the Providence Mountains. After this high ridge, the trail becomes difficult to follow, probably because timber was everywhere and mule trains chose various routes to stands of trees. Panniers ("J" shaped hooks forged from one inch steel rod and mounted on a mule's pack frame to support a balanced load) were found in this area (SBCM collections). There are very few piles of charcoal from here south on the mesa. However, there are numerous stumps and stacks of juniper cordwood which generally run along the west edge of the mesa. Although the trail can not be followed, it must have reached a point on the west edge of the mesa that is approximately east of Beecher Well. Remains of a rock buttressed trail run down a slope so steep that the route was probably only used to descend from the mesa. In Beecher Canyon, existing roads obscure the trail, but the route was probably down canyon, past Domingo Spring and then past the 7IL Ranch to Bonanza King Well and north to the Bonanza King mine. This trail gains 600 feet in its first mile and drops 1800 feet by the time the 7IL Ranch is reached; it is about 7.5 miles long.

Beecher Trail

At Wild Horse Mesa saddle, Beecher Trail runs southwest, then crosses the wash to a low plateau of tuff on the north wall of the canyon. It drops down cactus-covered slopes of metamorphic rock to the main channel of Beecher Canyon. There is evidence of wood cutting and charcoal making at the saddle and along the upper portions of the trail. Although the trail can not be followed to the canyon bottom, it probably connected with the Wild Horse Mesa trail at Beecher Well. It drops 1400 feet in four miles.

Summit Trail

The Summit Trail runs west-northwest from the mesa saddle in gullies cut in soft sediments of the Summit Spring Formation that underlies the Peach Springs Tuff (McCurry, this volume). The 3/4 mile trail runs past numerous piles of

juniper logs three to four feet long and up to 18 inches in diameter. Associated with these piles are concentrations of charcoal. In these charcoal areas there are often four large but moveable blocks of rock or rock piles. Perhaps a "floor" of logs was laid out on the rocks to increase ventilation and assure good initial draft when a stack of wood on the "floor" was ignited. The charcoal, once the coals were cool, would then have been loaded onto mules and hauled to the reducing furnaces of the mines. A shovel blade found near the saddle may reflect efforts to control the spread of flames from the charcoal operations. The Summit trail runs westerly, then south through Beecher Canyon to connect with Beecher Trail. This trail drops 900 feet along its three mile length.

Ridgeline Trail

A trail has been noticed west of Hill 5593 on the ridge that runs westerly to the Globe mine. My inspection has not yet located a trail connecting the Summit Trail with this ridgeline trail or with the Columbia mine to the north.

Summary

Wood is a precious commodity in the Mojave Desert. Before the turn of the century, when mining operations depended upon wood for timber and shoring and charcoal for ore reduction and when supplies were difficult and expensive to obtain, the piñon-juniper woodlands of Wild Horse Mesa were as much a bonanza to miners as was the ore they sought to extract from the nearby mines.

Acknowledgements

Larry Vredenburgh kindly reviewed a draft of this manuscript, and his insights are appreciated. I would also like to recognize the generosity and kindness of the late Jim Winkler, who not only shared his knowledge about the Mid Hills area but his cabin near the foot of the mesa.

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Architecture of the Grasses of the Eastern Mojave Desert

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Form and function are one.

— Frank Lloyd Wright

Survival in order to reproduce is basic in all life forms. Grasses have evolved many strategies to accomplish this. Ten thousand or more grasses grow world-wide — in Antarctica, the high Andes, coastal salt marshes, vernal pools, meadows, deserts, and sidewalk cracks. In the Eastern Mojave several dozen species thrive including some that grow nowhere else in California. A few are rare and endangered.

The non-professional botanist will find an exciting project in observing the survival techniques of these grasses in a harsh environment.

A ten-power magnifier, a sharp eye, and a delight in discovery are recommended.

The Grass Plant

Roots stabilize the plant in wind-swept loose soils. Deep rooted grasses seek subterranean water, the shorter

spreading roots take advantage of surface water from intermittent desert rains. Nutrients are carried to the plant by roots.

Annual grasses shoot up after early rains and some have the ability to produce seeds in a very few weeks that are scattered by winds, birds, and other mechanical means. Perennial grasses spread by stolons and rhizomes and seeds.

The **blades** provide photosynthesis and transpiration. In the dark tropic forests grasses may develop almost round leaves to capture more light and provide greater transpiration. Opposite needs exist in the desert. Blades are often grey, narrow and inrolled to reduce the reaction to heat, intense light and water loss. Prickles, hairs, sharp points and silica discourage browsers.

The **grass flower** is small and inconspicuous. It has no need to attract pollinators because grasses are wind pollinated. In the tropics there are a very few species insect-pollinated and these exhibit showy amounts of pollen as an attractant.

The **lemma** is the principal organ of the grass flower. It is protected by empty bracts and it, in turn, protects the seed. The center vein is often extended to provide the awn.

Spreading seed is the main job of the **awn**. It can also discourage grazers. Some awns plant the grass seed.

Changing the grasses' location is often needed to locate more fertile soil or moisture. Above-ground **stolons** can sprout new plants at the nodes. **Rhizomes** are below ground. **Bunch grasses** grow out from the center of the plant and often form rings. Wind, flash floods, fur, feathers, tires, and socks also disperse seeds.

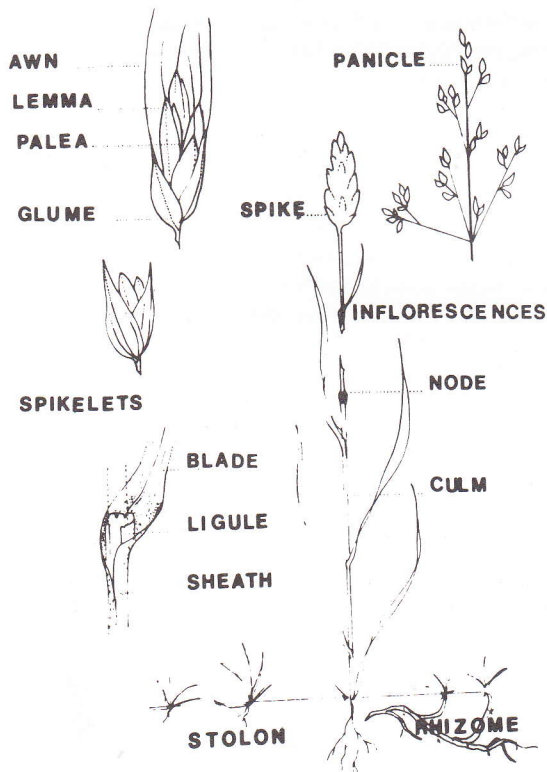


Figure 1. Parts of a grass plant.



Figure 2. Side oats gramma.



Figure 3. *Blue gramma*.

In sexual reproduction pollen is produced by **anthers** and wind-spread to the **stigma**.

Monoecious plants carry both male and female flowers. Dioecious plants carry only male or female flowers. *Scleropogon brevifolius*, a recent arrival to the Mojave, has distinctive male and female inflorescences and if in bloom exhibits a noticeable soft pink color over the colony. Self pollination exists in a few grasses, the anther and stigma contained in some florets.

Asexual reproduction includes vivipary. The grass converts parts of the flower into bulblets that often fall from the plant complete with root. Cleistogamous spikelets can occasionally be found on underground stems.

Polyploidy is common in grasses. Chromosomes will double in the cell, producing more than two of the basic chromosome sets and doubling the genes of the parent plant. You can often find variations within the same species as a result. This is an aid to evolution and adaptation to changing habitats — often brought about by roads, campgrounds, and ORVs.

Now, a chance for you to explore form and function in a few widespread Eastern Mojave grasses.

Eastern Mojave Grasses

Bouteloua curtipendula (Sideoats Grama) is easy to recognize in the field. The florets twist to one side of the flat axis. The florets fall in a package, leaving a bare zig-zag axis.

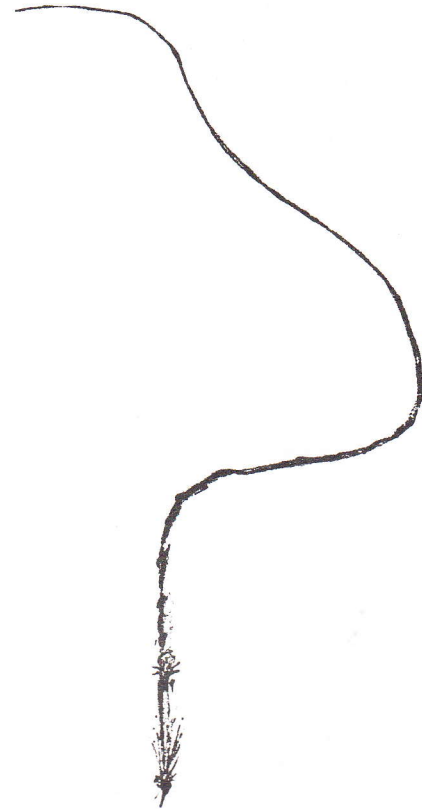


Figure 4. *Needle-thread grass*.

B. curtipendula: excellent opportunities to observe results of polyploidy and hybridization, using magnification.

B. gracilis bears an inflorescence with tightly packed florets that are awned and hairy. As it matures the inflorescence curves to an obvious button hook shape.

Stipas. New floras have regrouped the *Stipas* into *Achnatherum*, *Nassella*, and *Hesperostipa*. *Oryzopsis* (Indian rice grass) is also included. The field appearance of the *Stipas* is similar. Most species have strong awns that twist in the heat. When the desert cools the awns untwist and, using the sharp callus at the base of the seed, plant the seed by twisting it into the ground. It can also produce sock-ouch!

Stipa comata (Needle and Thread Grass) has flexuous awns that can reach 18 cm in length (two Russian species reach up to 50 cm!). Awns and lemmas can have long, showy hairs that aid in protection and dispersal.

Aristida (Three Awn) also makes prominent use of its awns. A short column at the top of the lemma supports three very visible awns.



Figure 5. *Three awn*.

These awns can be from 0.5 mm to 10 cm in length and are a good field mark. The awns are less tightly twisted than the *Stipas*, but they do disperse the seeds. The three awns tumble well in the wind.

Muhlenbergia (Muhly). This grass grows in tight, narrow spikes, low mats, or airy, wide-spreading panicles. Lemmas are awned or not. In the high Sierras, *M. minutissima* can be found growing out of the edges of snow banks, 2 cm tall and in full flower. *M. porteri*, a more robust desert form, has an open inflorescence with awned purple single spikelets growing at the tips of the branches. It seeks support in high winds by growing up and out of lower plants and using residual shade and moisture from its host.

M. rigens (Indian Basket Grass) grows near springs or other moist settings so the basal clumps of leaves are green and wider. The florets very closely circle the long spike giving a furry appearance to the bloom.

Botanists spend a great deal of time studying grasses, and that is good for knowledge. Amateurs can observe, discern, and enjoy, and that is good for the soul.

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Margaret W. Flesher has been studying grasses and habitats for the past 20 years in the western United States, Central America, and northern South America.

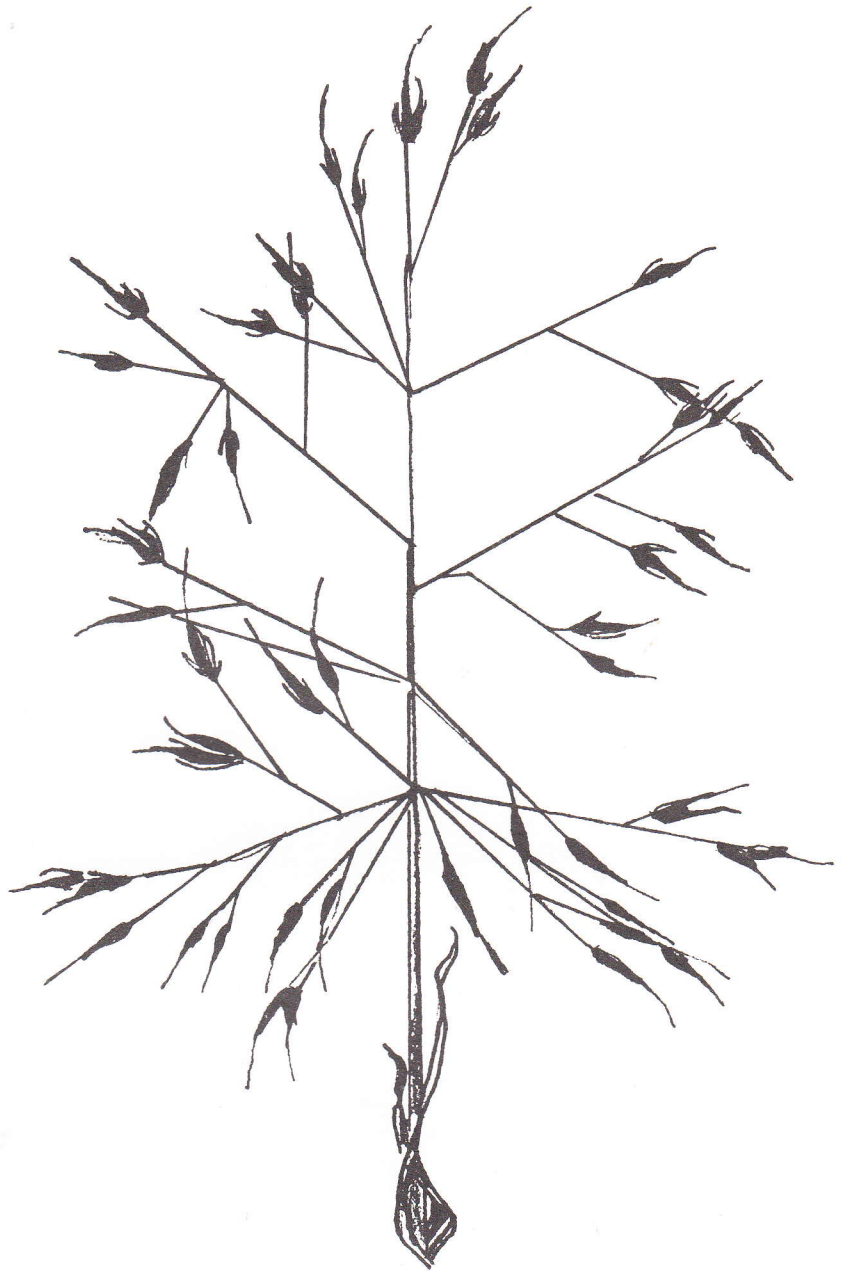


Figure 6. Bush muhly.

The Vegetation of Lobo Point and North Wild Horse Mesa, Eastern Mojave Desert, San Bernadino County, California

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A small valley near a prominent point of rocks called Lobo Point by local residents contains a pinyon-juniper woodland of about six hectares. Lobo Point is approximately three km northwest of BLM's Hole-in-the-Wall Campground, see Figure 1, and is located in the Columbia Mountain, California 7½" Quadrangle. At 1400 m elevation, the site is the lowest pinyon-juniper grove in the Mid Hills. A larger woodland is found in a valley north of Wild Horse Mesa, and about 3 km west of Lobo Point. The purpose of this study is to develop a description of the vegetation at these two sites, and relate it to local and regional vegetation. Botanical nomenclature follows Hickman (1993).

Geology and Climatic Setting

Rocks at Lobo Point and North Wild Horse Mesa are Miocene rhyolitic pyroclastic flows interbedded with lacustrine sedimentary rocks (McCurry 1985). Washes at Lobo Point contain grus (sand derived from granite) from the Little Thorn Mountains to the north. Precambrian metamorphic gneiss and schist crop out just north of the North Wild Horse Mesa and Lobo Point. Lobo Point is also well known to rock hounds for its "opalite" silicified lake bed deposits (Perry, 1977).

Climatic data is not available for either Lobo Point or North Wild Horse Mesa. The nearest weather station is Mitchell Caverns, CA, which is 15 km southwest of, and 50 m lower than, Lobo Point. Nearly 40 years of consistent weather data are available for Mitchell Caverns. Mountain Pass, CA, is 46 km north of and about the same elevation as Lobo Point. Unfortunately, only 10 years of discontinuous data are available.

Walter (1963) suggested a climatic diagram to relate precipitation to temperature as a surrogate for water balance. Climatic diagrams for Mitchell Caverns and Mountain Pass are shown in Figure 2. The climate at Mountain Pass is very similar to Mitchell Caverns, except slightly colder winters. The diagrams show a humid period from November through March, and an arid period from April through October. The favorable growing season, defined by absence of frost, is limited to June through September. May, the best month for flowering displays, may have frost and is arid. The most striking feature of the diagram is the summer precipitation peak. August is the wettest month in the eastern Mojave. However, the climate in August is still classified as arid.

West, Rea and Tausch (1975) published climatic diagrams for 16 stations where pinyon-juniper woodlands are located, ranging from California to Texas, and north to Idaho. Visually, the diagram for Sedona, AZ is the most similar to the eastern Mojave stations. However, the climatic diagrams for Mitchell Caverns and Mountain Pass are more

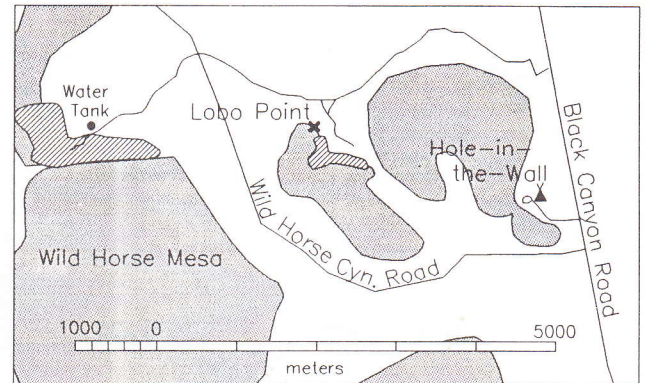


Figure 1. Location map.

similar to each other than they are to any of West *et al.*'s (1975) diagrams.

Methods and Results

I have measured four transects at Lobo Point and North Wild Horse Mesa. Transects from which absolute data can be calculated show the typical density of woody plants and shrubs in this area to be 1.3 plants per meter² and coverage about 25%. In addition, this data set includes seven published transects of others (Garcia-Moya and McKell, 1970; Cody, M. L. 1986a; Vasek and Barbour, 1988; Vasek and Thorne, 1988) and 46 transect sampling locations in the Columbia Mountain Quadrangle. I estimate the sampling in the quadrangle to be about ⅓ complete.

The vegetation at Lobo Point is a desert scrub, consisting of BLACKBUSH (*Coleogyne ramosissima*), WILD BUCKWHEAT (*Eriogonum fasciculatum*), BROOM SNAKEWEED (*Gutierrezia sarothrae*), *Ericameria* (*E. linearifolia* and *E. cooperi*), BUCKHORN CHOLLA (*Opuntia acanthocarpa*), NEVADA EPHEDRA (*E. nevadensis*), TURPENTINE BROOM (*Thamnosma montana*), and BLUE YUCCA (*Yucca baccata*). The scrub is similar to blackbush scrub, which can be problematic in definition, but is often found in the Mojave Desert above the creosote bush scrub and below the pinyon-juniper woodland. Several species are consistently associated with a wash habitat, particularly DESERT ALMOND (*Prunus fasciculatum*), WOOLLY BURSAGE (*Ambrosia eriocentra*), PURPLE SAGE (*Salvia dorrii*), BLADDERSAGE (*Salazaria mexicana*), and CATCLAW (*Acacia greggii*).

A pinyon-juniper woodland, with PINYON (*Pinus monophylla*) and UTAH JUNIPER (*Juniperus osteosperma*), is found in the sheltered north-facing valleys at Lobo Point and North Wild Horse Mesa. WRIGHT'S BUCKWHEAT (*Eriogonum wrightii*), GREEN EPHEDRA (*E. viridis*) and DESERT BITTERBRUSH (*Purshia glandulosa*) are often found in association with pinyons and junipers.

A small area on the north slope of Wild Horse Mesa has

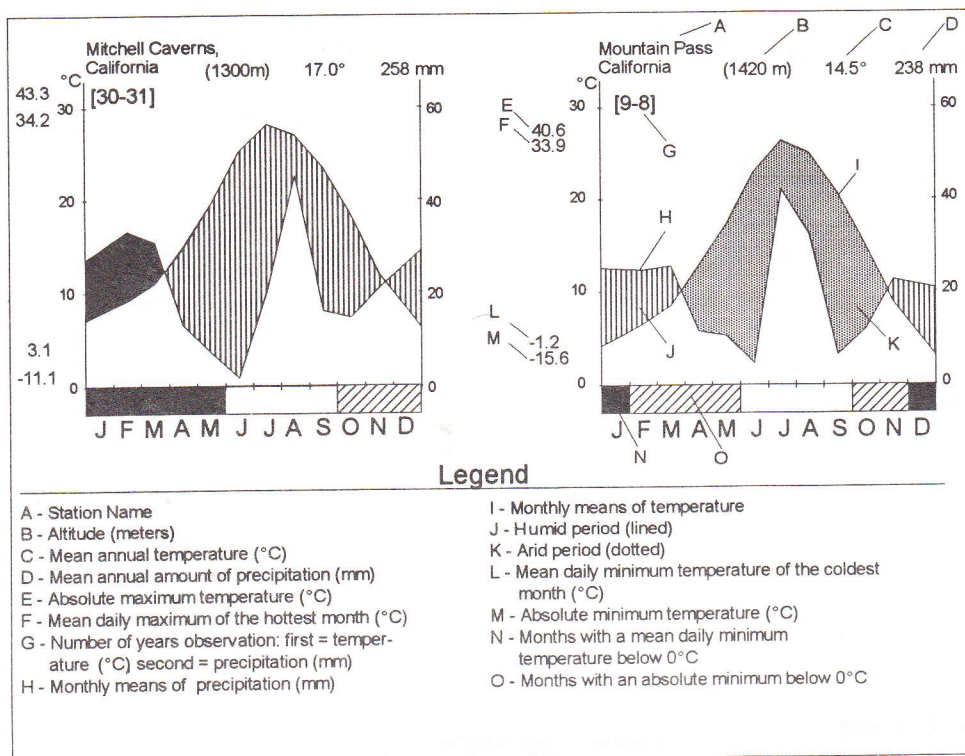


Figure 2. Climatic diagram for Mitchell Cavens and Mountain Pass, CA, after Walter (1963).

several species reported previously for the New York Mountains, but not the Providence Mountains or the Mid Hills. They are: CANYON LIVE OAK (*Quercus chrysolepis*), PALE SILK-TASSEL (*Garrya flavescens*), SINGLE-LEAVED ASH (*Fraxinus anomala*), WESTERN SERVICE-BERRY (*Amelanchier utahensis*), HOLLY-LEAF REDBERRY (*Rhamnus ilicifolia*) and *Swertia albomarginata*.

MOHAVE YUCCA (*Yucca schidigera*) is found at Lobo Point but is apparently at the upper elevational limit of its range. Many other species are found at Lobo Point. A complete plant list is available from the author.

Discussion

Yeaton *et al.* (1985) studied the three species association of JOSHUA TREE (*Yucca brevifolia*), MOHAVE YUCCA (*Y. schidigera*), and BLUE YUCCA (*Y. baccata*) occupying a 700 m elevational gradient along Cima-Ivanpah Road and Cedar

Table 1. Representative conclusions from Cody (1986a).
Species which tend to be uniformly distributed, especially at higher densities.

- Yucca brevifolia*
- Opuntia echinocarpa*
- Opuntia ramosissima*
- Yucca schidigera*

Species which tend toward a clumped distribution, especially at higher densities.

- Hymenoclea salsola*
- Acamptopappus sphaerocephalus*
- Ericameria cooperi* [*Haplopappus c.*]
- Salazaria mexicana*
- Saloia dorrii*

Canyon Road, 13 to 36 km north of Lobo Point. Their descriptions of the lower elevation transects correspond with the author's knowledge of the region. However, once the transect enters Cedar Canyon, factors other than elevation may cause changes in species density. Throughout Cedar Canyon a patchwork of Joshua tree woodland, pinyon-juniper woodland, and sagebrush scrub is observed. I believe this results from geomorphic expression of underlying geologic structure, and soil chemistry from three different rock types in the canyon.

Yeaton *et al.* (1985) found that vegetation increases in density and complexity from lower elevations to higher. Species composition also changes constantly with increasing elevation from creosote bush scrub to pinyon-juniper woodland. This indicates that temperature decreases and rainfall

increases with increasing elevation over this transect. *Y. schidigera* and *Y. brevifolia* are found together in the lower portion of the elevational gradient. *Y. baccata* replaces *Y. schidigera* at higher elevations. The maximum density and biomass for all three *Yucca* species totaled occurs at 1375 m. *Y. schidigera* and *Y. brevifolia*, which are associated extensively over the gradient, reach peaks of abundance toward the upper edges of their range but *Y. baccata* shows no change in its density or biomass from the upper portions of its range down to 1450 m where it drops out suddenly, to be replaced down slope by *Y. schidigera*.

Y. schidigera and *Y. baccata* are so similar in appearance that for many years *Y. schidigera* was considered a tree-like form of *Y. baccata*. Yeaton *et al.* (1985) proposed that the similarity of *Y. schidigera* and *Y. baccata* causes interspecific competition that limits the lower edge of the distribution of *Y. baccata*. They further proposed that the mode of competition between the two species is water utilization, in particular which subsurface zones might be used by the various species. Cody (1986a, 1986b), reviewed below, also discusses interspecific competition for water.

Applying the data of Yeaton *et al.* (1985) to Lobo Point (1400 m), I would predict dominance of *Y. baccata* over *Y. schidigera*. That is indeed the case. I would also predict presence of *Y. brevifolia* at Lobo Point. However, the nearest occurrence is along the Mid Hills ridge line, 6 km northwest at 1650 m elevation. The absence of *Y. brevifolia* on the southeast side of the Mid Hills ridge line remains a puzzle.

Comstock, Cooper, and Ehleringer (1988) studied seasonal patterns of canopy development and carbon gain in nineteen warm desert shrub species. This study was initiated to collect baseline data on the phenology of carbon gain in warm desert species. A special emphasis was placed on

Table 2. Classification of desert plants by photosynthetic strategy.

1. Non-photosynthetic stems
 - Group 1. Much leaf die-back during drought.
None identified at Lobo Point.
 - Group 2. Little leaf die-back during drought.
Ambrosia eriocentra
Salvia dorrii
2. Photosynthetic stems
 - a. Canopies die back to ground during drought
 - Group 3. High photosynthetic rate in leaf and stem.
None identified at Lobo Point.
 - Group 4. Photosynthetic rate higher in leaf than stem
None identified at Lobo Point
 - b. Maintains green canopy during drought, leaves deciduous
 - Group 5. Leaves make major contribution when present.
Gutierrezia sarothrae
Hymenoclea salsola
 - Group 6. Leaves make minor contribution when present.
Salazaria mexicana
Thamnosma montana

contrasting the performance of photosynthetic leaves and twigs. The 19 species studied could be subdivided into six useful groups based on 1) the percent contribution of green twigs to whole plant carbon gain, and 2) the percent reduction of photosynthetic area during drought periods. The groupings are shown in Table 1. The species names are shown only for those species that also occur at Lobo Point. In those species that maintained a substantial canopy area through the drought period, previously stressed tissues showed substantial recovery after fall rains.

Bradley and Deacon (1967) described blackbush scrub as an ecotonal vegetation type in southern Nevada, at the transition from the Mojave Desert to the Great Basin Desert. There, blackbush scrub forms the ecotone between creosote bush scrub and basin sagebrush scrub (Trimble 1989). Creosote bush scrub is found just to the south of and below Lobo Point. Basin sagebrush is found at 1600 m, just 3 km north of Lobo Point on the slopes of the Little Thorn Mountains. It appears that something like the ill-defined blackbush scrub also forms an ecotone between creosote bush and sagebrush at Lobo Point. This is in reverse order to that found in the White-Inyo Mountains, where Blackbush is found above the Sagebrush (Schoenherr, 1992).

Blackbush does not dominate at Lobo Point, occurring there with relative densities of 0.1 to 0.3. Blackbush does dominate at some sites in this area (e.g., along Black Canyon Road north of Hole-in-the-Wall, where relative densities of 0.5 to 0.6 are common). The species associated with blackbush in the eastern Mojave are different from those reported by Bradley and Deacon (1967). Vasek and Barbour (1988) point out that the definition of blackbush scrub by species is problematic. However, a scrub of low dark shrubs that lies above the creosote bush and extends up into Joshua trees, junipers, and pinyons, occurs over much of the Mojave Desert. This scrub often has the appearance of blackbush scrub even when BLACKBUSH (*Coleogyne ramosissima*) is not present.

Cody (1986a) studied shrub spacing patterns at the Granite Mountains and in the Mid Hills. Five species show uniform distributions, 30 species show a clumped distribution, and 32 species are distributed randomly. Top r

anking species in volume tend toward uniform spacing. The larger species trend toward more uniform spacing at sites where they reach higher densities, see Table 2. Some smaller species tend toward increasing aggregation as species density increases. Other smaller species are randomly scattered when rare and clumped when common.

Cody's (1986a) data also shows that some species have certain other species as neighbors far more often than would be expected by chance, and others occur less often than expected by chance. Cody discusses two possible causes: germination opportunity, and root interaction. Cody (1986b) reports direct observations of desert shrub root system morphology. A great deal of interspecific variability exists, along with a great deal of intraspecific constancy. He concludes that space preemption occurs among species that compete for a single limiting resource, such as water. Shrub species interact over space and presumably over the water rights included within this space. Therefore, a substantial component of the shrub diversity in this vegetation may be owed to a diversity of water uptake strategies, from strictly soil surface root systems, e.g., HEDGEHOG CACTUS *Echinocereus engelmannii* to strictly deep tap-rooted system, e.g., BURROBRUSH (*Hymenoclea salsola*), and including all kinds of intermediate morphologies.

As reported in Vasek and Barbour (1988), Vasek and Rowlands (not published) measured vegetational characteristics of a Joshua tree woodland north of Cima. This would be on the south slope of Cima Dome. Elevation is not reported. However, the elevation of Cima is 1273 m and the top of Cima Dome is 1751 m. The dominant species by far is *Ericameria [Happlopappus] cooperi*. Other common species are WINTER FAT (*Krascheninnikovia [Eurotia] lanata*), *Ephedra nevadensis*, and BLADDERSAGE (*Salazaria mexicana*). JOSHUA TREE (*Yucca brevifolia*) is visually dominant, resulting in the vegetation type being called Joshua tree woodland, but the species scores low on quantitative measures of dominance.

I found no published studies of the Sagebrush Scrub in

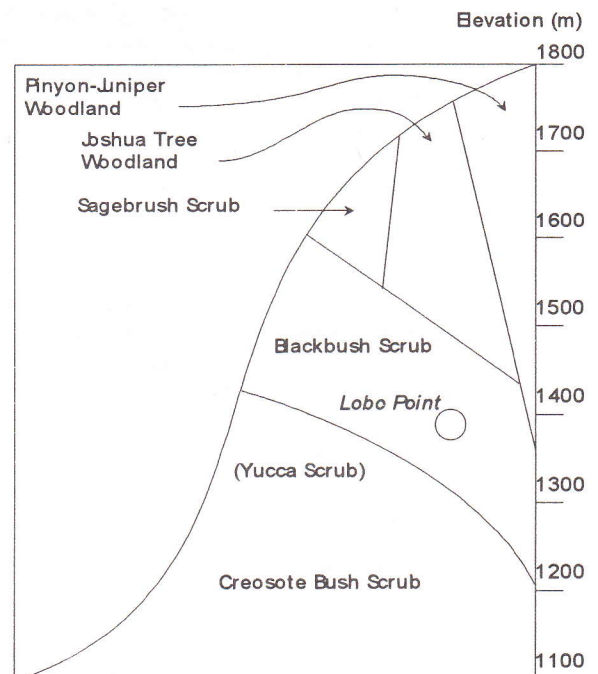


Figure 3. Vegetation zones in the vicinity of Lobo Point.

the Lobo Point area, only occasional references to the Round Valley Sagebrush assemblage, just north of the study area, in the Mid Hills, California 7½" Quadrangle and near the BLM's Mid Hills Campground. This assemblage extends south into the Columbia Mountain quadrangle between 1525 m to 1650 m elevation. My sampling transects, not reported here, show relative densities for Sagebrush commonly in the 0.70 to 0.80 range.

The distribution, autecology, synecology and economic use of pinyon pine have been summarized in many recent references; see, for example, Everett (1987). Much of the literature focuses on economic use and "management." This generally means improvement as cattle range, deer habitat for hunting, or increased rainfall runoff for storage in down stream reservoirs. Hurst (1977) points out "... the pinyon-juniper type has been a very hospitable living area since the arrival of man. Because of this, pinyon-juniper stands are rich in archeological treasures -- a value not adequately recognized." Recreation other than hunting is rarely mentioned in the literature. This is important because the increasing population of the metropolitan Los Angeles area is causing increasing pressure on the eastern Mojave Desert for recreation.

Bailey (1988) reported on his extensive investigation of pinyons with single-needle fascicles whose natural ranges lie mainly in the Californias and the Great Basin. Characters used by Bailey are number of resin ducts, number of stomatal lines, fascicle sheath curl back, and endosperm composition. He recognized three taxa with limited geographical overlap. Bailey's Collection No. 241 was made at the north end of Wild Horse Mesa. His sample showed no individuals of *P. monophylla*, and four individuals each of *P. californiarum* and *P. californiarum* subsp. *fallax*. Along the axis of the Granite Mountains, Providence Mountains and New York Mountains, few individuals of *P. monophylla* are found. Most common is *P. californiarum*, with *P. c.* subsp. *fallax* in smaller numbers and found mainly in the central and northern sections of the mountain range axis.

In another study about a close-by area, F. C. Vasek and H. B. Johnson (not published, reported in Vasek and Thorne, 1988) measured the relative densities of vegetation at sites in Caruthers Canyon, New York Mountains and on Clark Mountain. Caruthers Canyon ranges from 1650 to 2100 m in altitude and is 22 km northeast of Lobo Point. Vasek and Johnson's data showed presence of COFFEEBERRY (*Rhamnus californica*), REDBERRY (*R. crocea*), DESERT CEANOTHUS (*Ceanothus greggii vestitus*), YERBA SANTA (*Eriodictyon augustifolia*) and DESERT OLIVE (*Forestiera neomexicana*). The vegetation found in Caruthers Canyon suggests that it is unique among reported eastern Mojave sites. This is the only other study that reported WRIGHT'S BUCKWHEAT (*Eriogonum wrightii*) found among the pinyons at Lobo Point. Perhaps the *R. crocea* reported by Vasek and Johnson was *R. crocea* ssp. *ilicifolia*, now renamed *R. ilicifolia* and found at North Wild Horse Mesa.

At even higher elevations, the ROCKY MOUNTAIN WHITE FIR (*Abies concolor* var. *concolor*) grove in the New York Mountains has achieved legendary status. It contains about 30 trees in 0.8 ha at an elevation of 2100 m (Hendrickson and Prigge, 1975). White fir is also found in the Kingston and Clark Ranges of California and the Charleston Mountains of Nevada.

Summary

A diagram of biotic zonation in the Mid Hills is shown in Figure 3. Perhaps the unusual form of this diagram is a reflection of vegetation zonation resulting from topography of the Mid Hills. However, it is also interesting to speculate that it may result from the Mid Hills' position in the eastern Mojave mountain ranges containing a jumble of Californian, Sonoran, and Great Basin floristic regions with Rocky Mountain influences (Raven and Axelrod, 1978). My continuing work of systematically sampling vegetation in the Columbia Mountain quadrangle, and studying the distribution of species found in the Mid Hills is aimed at exploring these issues.

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Wildland Fire and Early Postfire Succession in Joshua Tree Woodland and Blackbrush Scrub of the Mojave Desert of California

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Although fire occurs regularly in the Kalahari semidesert of savanna woodlands in South Africa and in the central desert of Australia (Short et al 1976; Catling and Newsome 1981; Gill 1975; Gill et al. 1981; Griffin et al. 1983), fires are nearly absent from the Sonoran and Mojave Deserts of North America owing to limited biomass, wide spacing between shrubs, sparse ground cover, and insufficient accumulation of flammable material (Humphrey 1962, 1974; Brown and Minnich 1986). Recent studies in the Mojave Desert have focussed on the increasing threat of fire in creosote bush scrub due the combined effects of abnormally heavy precipitation between 1978-1985 and increases in flashy fuels caused by the invasion of exotic annuals from the Mediterranean basin and Middle East including *Bromus rubens*, *Schismus barbatus*, and *Brassica tournefortii* (Rogers and Steele, 1980; O'Leary and Minnich 1981; McLaughlin and Bowers, 1982; Brown and Minnich, 1986; Rogers and Vint 1987). These studies indicate that members of the creosote bush scrub ecosystem have a low tolerance to burning, have weakly developed adaptations to fire, and are subject to longstanding changes in the structure and species composition.

It has been speculated that fire has always occurred regularly, if infrequently, at higher altitudes of the Mojave Desert because cooler, semiarid climates support greater biomass and fuel build up (Humphrey 1974). Extensive burning has recently occurred in the high eastern Mojave Desert. In 1978-1985 fires denuded >10,000 ha of pinyon-juniper woodland, Joshua Tree woodland, and blackbrush scrub on the north flank of the Granite Mountains, southeastern Providence range, and the central Midhills near Hole in the Wall. In 1991-92 fires as large as 3000 ha occurred east of the

Midhills and in Lanfair Valley.

Because postfire successions of desert scrub communities appear to require perhaps a great amount of time—at scales from decades to centuries—the present outbreak of fires may result in directional change in the abundance and species composition of desert perennials unless fire intervals are very long. There are no postfire succession studies in high eastern desert ecosystems of California. To provide some

Table 1. Postfire succession chronosequence of perennial cover of Joshua Tree Woodland in Joshua Tree National Monument (percent)¹.

Site/ TSF (yr) Species)	Off-burn													
	HID 1	QW3 3	NWE 4	SHP1 6	QWS 8	QWN 8	RAN1 9	RAN2 9	QM2 10	RAN3 20	QW2 21	LHV 47	RAN5 47	
Ag					1.3	1.6			3.6		1.5	0.5		
Cr	6.6			1.3					3.0		2.2			
En	3.4		2.0	1.3	4.3	2.5	4.2		9.5		4.1	4.0		
Ef					0.7				3.8		0.5			
Hr	8.6		17.4	10.8	15.8	6.2	30.1		0.6		40.4	15.0		
Hs	6.4		2.3	5.7	1.2	0.8			17.1		7.8	0.3		
Lt			6.7	16.0		20.3								
La			2.4									0.5		
Lc	3.9				0.6	0.2	0.6				1.7	0.5		
Oa				0.5										
Oe	0.3		1.4		0.6	0.4			2.8		1.5	1.1		
Or	0.9		0.5		1.5				2.9		0.1	0.2		
Sx			5.5						2.9		1.7	4.5		
Sm									0.5					
Ss	0.2		2.0	2.0	0.7						0.5	4.8		
Yb	6.4		4.4	2.5	15.8	11.4	4.2		8.6		6.1	4.2		
Ys														
O	3.2		4.8		2.3	10.2			1.0		0.2	7.2		
Tot.	37.6		45.8	39.1	44.8	53.6	38.7		53.4		59.8	40.9		
Site/ TSF (yr) Species)	On-burn													
	HID 1	QW3 3	NWE 4	SHP1 6	QWS 8	QWN 8	RAN1 9	RAN2 9	QM2 10	RAN3 20	QW2 21	LHV 47	RAN5 47	
Ag					3.8	2.5			3.4		1.7			
Cr									1.2					
En	0.3	0.5		0.6	2.3	1.4	3.7	3.3	3.5		0.9	3.7	0.6	
Ef				1.0	0.5				2.8					
Hr	0.1	14.7	5.7	11.7	45.4	5.9	14.5	27.8	1.5	27.7	38.8	34.7	16.1	
Hs	0.3	0.7	1.2	6.8	8.5		1.3		2.9	3.7		4.1	5.0	
Lt			1.4	0.8										
La			1.0		1.2					0.7				
Lc	2.0	3.0	2.9	4.0	4.0	2.4	1.3			1.0	1.7	2.5	8.4	
Oa			0.1						0.3					
Oe							0.6	0.8			1.0	1.0		
Or			0.5											
Sx		3.0	0.9	0.6					1.7	4.8		1.0		
Sm									9.7					
Ss		1.5		0.2	2.0			0.6	8.7		4.5	5.6	1.6	
Yb	2.6		0.4			0.2	2.2	4.4	1.1			0.2	0.1	
Ys		0.5					3.5			1.8		0.2		
O	0.6	2.2	0.4		0.4	1.1	0.9		1.4		1.4	3.0	5.3	
Tot.	5.9	20.8	14.4	21.5	62.3	13.5	28.0	33.9	35.8	39.8	49.3	48.8	37.1	

¹Species: Ag, *Acacia greggii*; Cr, *Coleogyne ramosissima*; En, *Ephedra nevadensis*; Ef, *Eriogonum fasciculatum*; Hr, *Hilaria rigida*; Hs, *Hymenoclea salsola*; Lt, *Larrea tridentata*; La, *Lycium andersonii*; Lc, *Lycium cooperi*; Oa, *Opuntia acanthocarpa*; Oe, *Opuntia echinocarpa*; Or, *Opuntia ramosissima*; Sx, *Salazaria mexicana*; Ss, *Stipa speciosa*; Yb, *Yucca brevifolia*; Ys, *Yucca schidigera*; O, Other

Table 2. Postfire succession chronosequence of perennial density of Joshua Tree Woodland in Joshua Tree National Monument (ha⁻¹)¹.

Site/ TSF (yr) Species)	Off-burn												
	HID 1	QW3 3	NWE 4	SHP1 6	QWS 8	QWN 8	RAN1 9	RAN2 9	QM2 10	RAN3 20	QW2 21	LHV 47	RAN5 47
Ag					140	88			62		117	81	
Cr	632			55							117		
En	316		70	110	420	176	116		279		351	324	
Ef									124				
Hr	1581		1540	938	1190	352	957				2574	1053	
Hs	527		140	497	140	264			434		702		
Lt			210	386		353							
La			70									81	
Lc	211				70	88	29				117		
Oa			70										
Oe	105		70		70	88			31		468	81	
Or	211								62				
Sx			420						31			243	
Sm													
Ss	105		70	55	70							567	
Yb	211		140	165	280	176	58		124		117	162	
Ys													
Tot.	4217		2800	2206	2800	1766	1157		1242		4694	3268	

Site/ TSF (yr) Species)	On-burn												
	HID 1	QW3 3	NWE 4	SHP1 6	QWS 8	QWN 8	RAN1 9	RAN2 9	QM2 10	RAN3 20	QW2 21	LHV 47	RAN5 47
Ag					137	39			27		145		
Cr													
En	18	75		50	137	39	170	240	67		145	572	188
Ef													
Hr		2265	588	1100	3699	510	850	1980		1261	4350	3289	1375
Hs	9	225	100	600	1096				40	196		429	375
Lt			50	150									
La			50	100	137					49			
Lc	18	75	100		274	59	34			49	145	143	62
Oa										49			
Oe							68						
Or													
Sx		150	75						27	196			
Sm									54				
Ss		75						60	307		725	1144	250
Yb	9					19	68	120					62
Ys			75					136			147		
Tot.	81	3021	1024	2029	5495	784	1369	2400	535	1947	5813	5739	2500
Yb Rcr.	0	0	0	0	0	0	0	20	10	0	10	20	40

Species: see Table 1.

estimate of probable postfire recovery in the eastern deserts, this study summarizes the fire history and postfire succession in Joshua Tree woodland (JTW) and blackbrush scrub (BBS) of Joshua Tree National Monument, by means of sampling chronosequences of postfire successions. The monument has experienced a similar fire history to the eastern desert ranges as fires, carried by indigenous vegetation, denuded extensive areas between 1978-1987. My hypothesis is that plant adaptations to burning are better developed in JTW and BBS than in creosote bush scrub such that these ecosystems are stable under recurrent fire.

Study Area

Joshua Tree National Monument, located 150 km east of Los Angeles, encompasses the Little San Bernardino Mountains of the eastern Transverse Range geomorphic province. The range is an uplifted block that forms a precipitous southern escarpment along the San Andreas

fault. The summit is a broad plateau averaging 1,100-1,500 m altitude with scattered peaks as high as 1,800 m. Lithologies in the JTNM plateau are mesozoic granitics, as well as precambrian igneous and metamorphic rocks. The climate is mediterranean with cool winters with light precipitation and warm, dry summers. The North American monsoon causes scattered thundershowers from July to September. Mean annual precipitation averages 15-25 cm (Climatological Data, California 1976-1985).

JTW is widespread on deep pervious soils overlying alluvial fans of the plateau. BBS grows primarily on steep hillslopes with precambrian igneous and metamorphic substrate, and adjacent thin, well-drained alluvium. Pinyon juniper woodland, occurs on hillslopes with granitic substrate. The entire area supports ground cover of spring annuals and ephemerals. Important genera contributing to this annual complex include *Oenothera*, *Mentzelia*, *Gilia*, *Eschscholtzia*, *Malacothrix*, *Eriophyllum*, *Castilleja*, and *Calochortus*. Exotic annuals found in JTW and BBS include *Bromus tectorum*, *B. rubens*, *Schismus barbatus*, and *Erodium cicutarium*.

Methods

Fire history was reconstructed from perimeter data (on file, Joshua Tree National Monument), and from aerial photographs in 1952, 1970-75, and landsat imagery (Minnich 1983). In JTW and BBS fires leave landscape scars that persist several decades, making it possible to produce perimeters having greater accuracy than JTNM file data. Landscape scars were also used to establish sampling locations across fire boundaries. Postfire succession was estimated by sampling burns along chronosequences as old as 47 yr in JTW and 21 yr in BBS. In the chronosequences, spatial vegetation patterns are used to deduce temporal vegetation change on the assumption that age is the primary ecologically significance among sampling sites (Johnson and Gutsell 1994). Seventeen sites near fire boundaries were sampled: 13 sites in JTW growing mostly on gentle alluvium, and 4 sites in BBS on steep, rocky hillslopes. At each site two 100-m parallel transects were employed on the unburned and unburned sides of fire boundaries. Vegetation in unburned transects is assumed to represent the prefire state of vegetation. The line intercept

method was used to obtain percent shrub cover (Canfield 1942). Along the intercept, the density of perennial species was obtained by the point-center quarter method (Cottam and Curtis 1956), using points every 10-m. Transects were initiated by a blind rock toss. Within the burns, I observed species resprouted, established from seedlings, or survived the burn. Fire-killed stems were recorded for arboreal *Yucca brevifolia* and *Y. schidigera*.

Results

Mature JTW consists of open stands of *Yucca brevifolia* 5-15 m tall (50-200 stem ha⁻¹, 5-15% cover) with a low, nearly contiguous understory 0.5-1.5 m tall of perennial bunch grasses (*Hilaria rigida*, *Stipa speciosa*), small subshrubs (*Ephedra nevadensis*, *Lycium cooperi*, *L. andersonii*, *Hymenoclea salsola*), and stem-succulents (*Opuntia acanthocarpa*, *O. echinocarpa*, Tables 1 and 2). Perennial cover varies from 35 to 60%, and stem densities vary from 1000 to 5000 stems ha⁻¹. Highest stand cover and densities occur in sites with abundant *Hilaria rigida* (cover, 5-40%; densities, 300- 2500 stems ha⁻¹). The cover of understory shrub species each is < 5%, and densities < 200 ha⁻¹. *Larrea tridentata* is an important component of JTW at three sites below 1,200 m.

BBS is dominated by dense low stands (<0.5 m tall) of *Coleogyne ramosissima* (15-30% cover, 600-2000 stems ha⁻¹, Tables 3 and 4). *Coleogyne ramosissima* grows with perennial grasses *Hilaria rigida* and *Stipa speciosa* and several subshrubs *Ephedra nevadensis*, *Eriogonum fasciculatum*, *Lycium cooperi* and *Viguiera deltoidea*. BBS contains an open arboreal layer of *Yucca brevifolia*, *Y. schidigera*, and *Juniperus californica* (5-10 m tall) which together make up 5-20% cover.

Fire history

About 5,000 ha of JTW and BBS were burned between 1978 and 1987. Small burns occurred in Lost Horse Valley and Quail Wash in 1942, 1967, and 1968. Nearly all fires were initiated by lightning from summer thunderstorms. Monument records indicate that fires often began as "sleepers" which stored in large fuels for several days before establishing flame fronts in drier weather. The frequency of large fires (>10 ha) represents a small percentage of total lightning fires (average, 10 yr⁻¹) which usually burn out at < 1-ha.

Table 3. Postfire succession chronosequence of perennial cover, Blackbrush scrub (percent)¹.

Site Species	Off-burn				On-burn				
	SHP2	JUN2	QM1	RAN4	TSF ¹ (yr)	SHP2	JUN2	QM1	RAN4
					6	9	10	20	
<i>Ambrosia dumosa</i>	4.4				6.4				
<i>Coleogyne ramosissima</i>	15.8	30.6	34.1	16.9	0.2		1.7	0.7	
<i>Encelia virginensis</i>	2.0								
<i>Ephedra nevadensis</i>	0.7	3.8	0.7	1.7		1.1	2.5	1.2	
<i>Eriogonum fasciculatum</i>		1.4		1.6		9.9		0.9	
<i>Hilaria rigida</i>	1.7		5.4	18.7		0.7		16.3	
<i>Hymenoclea salsola</i>			1.5	1.7				10.0	
<i>Juniperus californica</i>		8.9	10.0	3.3				0.6	
<i>Lycium cooperi</i>	1.4			1.0		2.5	2.6	3.2	
<i>Opuntia acanthocarpa</i>				1.7				2.3	
<i>O. echinocarpa</i>	1.4	0.3							
<i>Prunus fasciculata</i>			2.3				6.0		
<i>Salazaria mexicana</i>				0.9	0.9	13.6	3.4	7.1	
<i>Salvia mohavensis</i>							2.9		
<i>Stipa speciosa</i>	3.5	6.0			6.1	12.3	11.1		
<i>Viguiera deltoidea</i>	3.1				1.4		15.3		
<i>Yucca brevifolia</i>	6.1	1.2	3.9	4.3	0.9			1.0	
<i>Y. schidigera</i>	5.6		2.3	7.1	0.2		2.5	7.1	
Other species	2.0		2.3	1.9	4.5		6.8	0.4	
Total	42.1	47.4	59.9	54.7	20.6	38.7	49.5	50.8	

1. TSF, time-since fire

Fire reports (file data, Joshua Tree National Monument) indicate that blazes in JTW were propagated by a continuous layer of native perennial shrubs and bunch grasses. *Hilaria rigida* and *Stipa speciosa* were most responsible for sustained fire spread in JTW. Fuels produced by exotic annual herbs were insignificant. Wind is not necessary to sustain flame fronts in JTW due to the abundance of flashy fuels, although high spread rates occur with high winds and low humidity. Fire reports indicate that flames move through BBS only in high winds. Hence, fires are usually brief and cover small areas. While *Coleogyne ramosissima* is a low shrub, the density of fine branches provides fuel for flame lengths that may exceed 3 m. However, wind speeds > 5 m s⁻¹ tend to over-ventilate the upper branches and allow the fine dead

Table 4. Postfire succession chronosequence of perennial density of Blackbrush-California Juniper scrub in Joshua Tree National Monument (ha⁻¹).

Site/Species	Off-burn				On-burn				
	SHP2	JUN2	QM1	RAN4	TSF ¹ (yr)	SHP2	JUN2	QM1	RAN4
					6	9	10	20	
<i>Ambrosia dumosa</i>	762				1392				
<i>Coleogyne ramosissima</i>	2159	2062	672	949					
<i>Encelia virginensis</i>	127				706				
<i>Ephedra nevadensis</i>	254	290		146		160	87	65	
<i>Eriogonum fasciculatum</i>		194		146		800			
<i>Hilaria rigida</i>			96	1022		80		975	
<i>Hymenoclea salsola</i>				73				585	
<i>Juniperus californica</i>		194	72	73					
<i>Lycium cooperi</i>							58	65	
<i>Opuntia acanthocarpa</i>				146				190	
<i>O. echinocarpa</i>									
<i>Prunus fasciculata</i>							145		
<i>Salazaria mexicana</i>				73		116	632	377	650
<i>Salvia mohavensis</i>						116		87	
<i>Stipa speciosa</i>	508	970			1750	1520	319		
<i>Viguiera deltoidea</i>	127				464				
<i>Yucca brevifolia</i>	127	97	48	73					
<i>Y. schidigera</i>	381		72	146				87	260
Total	5102	3906	960	2924	4630	3193	1160	2790	
<i>Y. brevifolia</i> recruits						0	10	0	10

1. TSF, time-since-fire.

Table 5. Summary of fire weather data.

Burn site	Year	Ignition source	Sample sites	Vegetation sampled	Relative Humidity (%)	Wind dir/Speed (ms ⁻¹)
Randolph	1942	anthropogenic	LHV, RAN5	JTW	-	-
Quail Wash	1967	lightning	QW2	JTW	-	-
Randolph	1968	lightning	RAN3, RAN4	JTW/BBS	-	SW 4
Quail Mtn.	1978	lightning	QM1, QM2	JTW/BBS	20	W 0-10
Randolph	1979	lightning	RAN1, RAN2	JTW	-	-
Juniper Fl.	1979	lightning	JUN2	BBS	-	-
Quail Wash	1980	lightning	QWS, QWN,	JTW	-	-
Sheep Pass	1982	anthropogenic	SHP1, SHP2	JTW	17-30	SW 4-9
Northwest Entrance	1984	anthropogenic	NWE	JTW	27-32	NE 6-10
Quail Wash	1985	lightning	QW3	JTW	38	NE 3-5
Hidden V.	1987	lightning	HID	JTW	-	-

underbranches to burn out before critical ignition temperatures for live fuels are reached. Active spread of flame fronts usually occurred during the afternoon with winds <10 ms⁻¹ and relative humidities of 20-50%. Flame fronts died out in cool, humid weather at night (Table 5). Several burns ran into rocky barriers.

Fires typically denuded the shrub layer, leaving only an arboreal layer of *Yucca brevifolia*, *Y. schidigera*, and *Juniperus californica* skeletons. Fire damage is often species-specific due to individual species morphologies. *Larrea tridentata* tends to scorch rather than burn because of its relatively tall stature, spreading branch habit, high foliar fuel moisture, and low dead fuel content of its canopy (Brown and Minnich 1986). Drought-deciduous shrubs and bunch grasses including *Hymenoclea salsola*, *Ephedra nevadensis*, *Lycium spp.*, *Encelia virginensis*, *Salazaria mexicana*, and *Coleogyne ramosissima* burn to the ground because of their small stature, proximity of canopies with ground fuels, finely divided branching habits, and abundance of small leaves (high fuel continuity). The perennial grasses *Hilaria rigida* and *Stipa speciosa* are usually completely consumed because tufts build up of dead organic matter within the interior of the clumps. Stem-succulents in such genera as *Opuntia* and *Echinocereus* are scorched because of their high fuel moisture content, but flame temperatures apparently destroy cell walls, resulting in eventual uncontrolled desiccation of stems. Despite their tall stature above flame lengths, the canopies of arboreal leaf-succulents *Yucca schidigera* and *Y. brevifolia* are frequently destroyed because flames arising from ground fuels are propagated into the canopy by persistent dead leaves along their trunks.

Succession Chronosequences

In the landscape, burns are defined by conspicuous gaps in cover that can be seen from a distance, especially among the more arboreal or dominant species such as *Larrea tridentata*, *Yucca brevifolia* and *Coleogyne ramosissima*. Fire lines are accompanied by discrete shifts in species composition and biomass. JTW succession is dominated by resprouting of shrubs and perennial grasses with few species changes, while BBS succession is associated with species replacements caused by high mortality of *Coleogyne ramosissima*. There is little evidence of recolonization of arboreal species in either JTW or BBS. Differences in onburn/offburn densities along chronosequences reflects inter-site variability related to habitat rather than to successional processes.

Joshua Tree Woodland.

Burns throughout the chronosequence are dominated by vegetative reestablishment of perennial grasses and shrubs, with most species returning to preburn cover and densities within a few years (Tables 1 and 2). All post-burn stands are dominated by perennial bunch grasses *Hilaria rigida* and *Stipa speciosa*, mixed with scattered cover of *Ephedra nevadensis*, *Lycium cooperi*, and *Hymenoclea*

salsola. *Hilaria rigida* develops general cover by sending up numerous stems which develop into tuft-like clumps from rhizomes emerging from subsurface root network. The clumps arise from tillers developing from the culms. *H. rigida* cover was 10-45% within 10 years, with tuft densities ranging 588 to 3699 stems ha⁻¹. *Stipa speciosa* and other suffrutescent shrubs form an additional 5-15% ground cover. In *Stipa speciosa*, apical meristems close to the ground in the interior of the clumps survive the flames that consume the leaves. Stems do not form rhizomes; hence clump growth and densities were not substantially greater than on off-burn samples. However, it locally increases ground cover as a consequence of tillers establishing new tufts outside the culm, increasing the surface area over which each culm occupies. The large size of *S. speciosa* tufts in burns may also reflect the addition of soil nutrients from ash. *Ephedra nevadensis*, *Lycium cooperi*, and *Hymenoclea salsola* persist by adventitious resprouting from stem tissue of root crowns. These species maintained preburn cover and densities because fire mortality is rare. *Larrea tridentata* in early succession had directly survived fires with green canopy. On burn densities were much lower than offburn densities due to high fire mortality (Table 6). Most recruitment in early succession consisted of *Hymenoclea salsola* and *Salazaria mexicana*. *Opuntia echinocarpa* and *O. ramosissima* had established new stems in >20 yr stands from seed or vegetatively from living joints dispersed from other individuals by animals.

Total site stem densities had reached preburn levels of 500 to 5495 stems ha⁻¹ within 3-6 yr because most resprouting and seedling establishment occurs during the first few growing seasons. Total cover of the shrub layer increases for 20 yr as a consequence of the growth of individual stems, ranging from 6-20% in burns of <6 yr, 28-35% in burns 6-10 yr, and 37-49% in burns 20-47 yr.

The most conspicuous vegetation change in Joshua Tree woodland was the destruction of *Yucca brevifolia*. Surviving individuals either resprouted by sending up "pups" from rhizomes along primary root axes extending radially from the root crown (17% of stems), or had sufficiently tall canopies to withstand the heat of ground fires (14% of stems, Table 6). 64 to 95% of *Y. brevifolia* stems were fatally damaged in all but one of the sample sites. Low mortality at RAN1 (23%) was related to an unusually low intensity understory burn (file data, Joshua Tree National Monument). Resprouts were most frequent where the

proportion of surviving stems was also greatest. Hence, it appears that the resprouting rates of *Y. brevifolia* are inversely related to fire intensities.

Yucca brevifolia recruitment appears to be slow but continuous. No seedlings or saplings were seen in burns < 10 years old and recruitment densities are typically <40 ha⁻¹ in older burns. Even the 47-yr Randolph Flat burn had only widely scattered saplings mostly < 3 m height.

Blackbrush Scrub. Fires in blackbrush result in a discrete transformation in postburn species composition (Tables 3 and 4). *Coleogyne ramosissima* was entirely fire-killed because it is an obligate non-sprouter; surviving individuals exist on local unburned "fire islands" resulting from reticulate fire spread. Early succession was dominated by *Viguiera deltoidea*, *Eriogonum fasciculatum*, *Salvia mohavensis*, and *Salazaria mexicana*, all of which established as seedlings (char was not observed on root axes). Except for *S. mexicana*, these shrubs are obligate nonsprouters or facultative sprouters. However, the low frequency of *S. mexicana* in off-burn samples indicates that it was not present before the burns. Resprouting shrubs in BBS were similar to those in JTJW (*Hilaria rigida*, *Stipa speciosa*, *Hymenoclea salsola*, *Ephedra nevadensis*, *Prunus fasciculata*, *Lycium cooperi*), and maintained preburn densities. The total cover and density of "resprouting" species was smaller than perennials establishing from seed. Total understory cover increases from 20% at 6 yr to 50% at 20 years (Table 3).

Most *Yucca brevifolia* and *Juniperus californica* sustained heavy mortality, although partially damaged *Juniperus californica* survived the burn. While the smaller stature of *Yucca schidigera* exposed arboreal leaf blades to greater flame temperatures than in taller *Y. brevifolia*, *Y. schidigera* experienced higher resprouting rates of 75% of individuals (Table 6). Resprouts consisted of basal rosettes emerging from subsurface rhizomes. Except for the removal of *Coleogyne ramosissima*, species differences between plots appear to be more due to habitat than to succession species changes. *Eriogonum fasciculatum* is abundant on shallow alluvium at Juniper flats, while *Viguiera deltoidea*, *Salazaria mexicana*, and *Salvia mohavensis* tend to be most abundant on rocky slopes. The only late successional recruitment in 20 yr burns was the establishment of *Opuntia* spp., either seedlings or by vegetative establishment from dispersed joints. A few *Juniperus californica* seedlings were found on 20 yr Randolph plot. *Coleogyne ramosissima* seedlings were not observed in any burn.

Table 6. Postfire response of *Yucca brevifolia* and *Y. schidigera* (percent of stems ha⁻¹)¹.

Site	Time-since-fire	<i>Yucca brevifolia</i>				<i>Yucca schidigera</i>				<i>Juniperus californica</i>				<i>Larrea tridentata</i>			
		n	U	R	M	n	U	R	M	n	U	R	M	n	U	R	M
Joshua Tree Woodland																	
HID	1	220	9	27	64												
HID1	2	40	5	0	95												
QW3	3	90	0	29	71	20	0	100	0								
NWE	4	110	18	18	64									190	0	21	79
SHP1	6	230	13	13	74									390	10	21	69
QWN	8	80	0	13	87									140	0	0	100
QW5	8	160	0	6	94												
RAN2	9	90	11	33	56												
RAN1	9	220	50	27	23												
RAN6	10	50	8	28	64												
QM2	10	210	10	10	80					30	0	0	100				
RAN3	20	80	0	13	87	310	0	77	23								
QW2	21	120	0	0	100	20	0	100	0								
LHV	46	120	25	0	75												
All sites		1730	14	16	70	350	0	80	20	30	0	0	100	720	6	17	77
Blackbrush scrub																	
SHP2	6	20	0	50	50	200	0	85	15	10	0	0	100				
JUN2	9	110	9	0	91					80	12	0	88				
QM1	10	60	0	0	100	130	0	100	0								
RAN4	20	130	38	8	54	250	16	84	0								
All sites		320	19	7	75	580	7	84	9	90	11		89				

¹ U, unburned; R, resprout; M, mortality; M%, percent mortality.

Discussion

Whereas fire may have been absent from creosote bush scrub in hyperarid lower deserts before the arrival of European exotic grasses and forbs (Humphrey 1974; Brown and Minnich 1986), recurrent fire occurs in JTJW and BBS of the Mojave Desert because the semi-arid climate of cool temperatures and light winter precipitation supports a subcontinuous cover of fuels provided by indigenous perennial grasses and subshrubs. Within the ground layer *Hilaria rigida* and *Stipa speciosa* are particularly flammable, consisting of hemispherical tuft-like clumps with cores dead grass, with ample air space among the blades. Subshrub canopies also contain an abundance of fine stems and small leaves. Although limited productivity delays the establishment of canopy in blackbrush scrub, it burns in high temperature, high wind velocity and low relative humidity despite the open spacing of plants (Callison et al, 1985). *Erodium cicutarium* and native forbs provide only limited fuels in both communities because cured stems shatter into fine parts which fall to the ground and blow away. Exotic *Bromus tectorum* is not a significant fuel despite the prohibition of grazing on the monument.

Postfire succession studies of mudflows, flash floods, and human land-uses such as mining operations, agricultural clearing, and transmission line construction in high desert ecosystems of eastern California and Nevada, show that long-lived, poorly competitive species low reproductive capacities are eliminated by disturbance. Succession models indicate that early colonization is characterized by species with high growth rates, high reproductive capacities, relatively uniform, low stature, and short life spans (Wells 1961; Vasek et al., 1975a,b; Vasek 1979-80, 1980, 1983; Carpenter et al., 1986; Webb et al., 1987, 1988). Most colonizers commonly grow along washes, sandy sites, and

steep slopes subject to recurrent fluvial and aeolian disturbances (Vasek 1979-80, 1980; Brown and Minnich 1986). With time, the community increases in height, structural complexity and biomass as early successional species are replaced by later successional species.

Succession models have not weighed the role of stochastic processes that are likely to be more significant in secondary successions resulting from fire, including the intensity of disturbances (amount of canopy removal), seed bed conditions, in seed dispersal and fluctuations in establishment and growth rates due to climatic variability (Noble and Slatyer 1980). Most importantly, autogenic succession models developed from studies of geomorphic and human disturbances which result in total denudation, do not account for the carryover vegetation in fire disturbances which may lead to quicker vegetation recovery.

In JTW postfire succession is characterized by the vegetative persistence of extant stems through fires, and rapid postfire succession. Preburn density achieved in 6 yr and preburn cover within 10 years. *Hilaria rigida* can recover most of its preburn biomass within a few growing seasons following fire when adequate soil moisture exists because of its capacity to establish above-ground tufts from rhizomes. *Stipa speciosa* establishes preburn stem densities because meristemic tissue is concentrated near the surface or underground where temperatures are relatively low during fire. Suffrutescent shrubs persist by root-crown resprouting (*Lycium andersonii*, *L. cooperi*, *Ephedra nevadensis*, *Yucca schidigera*). There appears to be little recruitment during succession except for *Hymenoclea salsola* and continuous recruitment of fleshy fruited shrubs, such as *Lycium spp.*, *Ephedra spp.*, and *Opuntia echinocarpa*. *O. echinocarpa* also establishes new stems from fallen and/or transported joints.

Little is known about fire effects on *Yucca brevifolia* other than the fact the species is intolerant of even low intensity fires (Humphrey 1974). High fire mortality in Joshua Tree National Monument reflects variable and unpredictable resprouting and heavy rodent browsing. Succession chronosequences indicate that *Y. brevifolia* may reach preburn densities, cover and stature in perhaps a century. Although seed germination does not require scarification or leaching, and is not season-dependent (Went 1948), postfire recruitment is scarce before 10 yr and stem densities are 25-50% of those in 47-yr Randolph burnose (Table 2, Phillips et al., 1980).

Postfire succession in BBS is similar to findings in Nevada. *Coleogyne ramosissima* is highly susceptible to fire because it is an obligate nonsprouter (Beatley 1966), and is slow to reinvade after fires Webb et al (1988). The time required for its reestablishment to dominance is unknown (Bowns and West 1976). Following fire, denuded sites are colonized by abundant seedlings of suffrutescent shrubs, including *Hymenoclea salsola*, *Salazaria mexicana*, *Viguiera deltoidea*, *Happlopappus cooperi* and *Salvia mojavnensis*, and the persistence of obligate sprouters *Lycium spp.*, *Ephedra spp.*, similar to successions in Death Valley (Webb et al., 1987; 1988). Recruitment was apparently encouraged by the removal of *C. ramosissima* combined with mineral soil conditions. The even-size appearance of *S. mexicana* suggests that stems are even-aged, having established from seed caches. Seed of *Eriogonum fasciculatum* and *S. mohavensis* are

also locally dispersed by animals. *E. fasciculatum* may have also established from seed caches (O'Leary 1990). *V. deltoidea* seedlings germinate from wind dispersed seed from nearby unburned populations. The onburn/offburn convergence of stem densities within 6 and 9 yr contrasts with findings in Nevada where burns were dominated by forbs and grasses for 12 yr. The shrub layer was dominated by resprouts of *Ephedra viridus*, *Prunus fasciculata*, and *Eriodictyon* seedlings (Callison et al., 1985). The ground cover of *Coleogyne ramosissima* was only 1.7% compared to offburn cover of 38%. Although seedlings often emerge from rodent caches, it is a weak competitor and recolonization is unpredictable (Bowns and West 1976). Best seedling production in pulses following above average rainfall and reduced herbivore pressure (Beatley 1975, 1980; Wallace and Romney 1972).

Fire regimes

Available fire history records are insufficient to establish the nature of long-term fire regimes. However, succession chronosequences do permit the estimation of minimum fire intervals necessary to breach successions before the establishment of long-lived, self-maintaining communities. Studies of postfire succession in debris flows and human disturbances show that the recovery of long-lived perennials may take several centuries (Webb et al 1987, 1988; Prose and Metzger 1985; Knapp 1992). Such lengthy succession times are unlikely after fire because: (1) vegetation recovery is not dependent on primary successional processes associated with soil development; and (2) there is substantial carry over vegetation through fires. Still, chronosequences in Joshua Tree National Monument show that BBS stands as old as 21 yr show little reestablishment of *Coleogyne ramosissima*, similar to findings in Nevada (Callison et al 1985). The time required for complete recovery is unknown, and may be dependent on autogenic processes in which very long-lived species such as *C. ramosissima* (Bowns and West 1976; Christensen and Brown 1963) accumulate biomass, stabilize productivity at maintenance levels, and have greater adaptation to environmental stress in later succession (summarized in Webb 1988). Selection for competitive ability results in dominance by one or a few long-lived species. Reduced moisture loss through evaporation with increasing ground cover is offset by increased transpiration. Thus, competition for the moisture decreases for species that are competitively superior and long-lived gain dominance. *C. ramosissima* may also alter soils chemistry by increasing soil N and available P.

JTW presents a paradox because the ecosystem comprises a resilient understory that achieves preburn status within 5 yr, and a maladapted overstory of *Yucca brevifolia* that achieves preburn status in perhaps a century. The coexistence of these vegetation strata may be explained by two models: 1) understory fires normally burn at low intensities, leaving arboreal *Y. brevifolia* canopies unburned. 2) Understory fires kill *Y. brevifolia* but fire intervals are long. Evidences support the second model. While *Yucca brevifolia* possesses a thin bark layer and living cambium is found throughout the bole, characteristic of monocots, understory fires are propagated into the canopy by a shag of leaf blades along the bole, much like palms. Only one of 13 sample sites had high *Y. brevifolia* survivorship. However, fire cycle periods, or the amount of time required for the burn area

Table 7. Annual Precipitation at Victorville (cm)¹.
(Source: USDC, 1976-85)

Year	Victorville		Twentynine Palms		Mitchell's Cavern	
	Total	PON ¹	Total	PON ¹	Total	PON ¹
1976-77	21.1	166	14.1	140	18.1	65
1977-78	32.7	257	23.0	228	54.6	194
1978-79	22.8	180	13.8	137	31.6	113
1979-80	24.3	191	22.3	221	75.8	270
1980-81	7.4	58	11.2	111	25.7	92
1981-82	15.3	120	5.3	52	25.3	90
1982-83	32.7	257	12.2	220	49.9	178
1983-84	9.4	74	21.7	215	33.4	119
1984-85	17.1	135	18.1	179	52.4	187

1. July 1 to June 30.

(1,200 ha) to equal vegetation area (25,000 ha) for the period 1942-1995 (Johnson and Gutsell 1994), are estimated to be in the scale of thousands of years at current burning rates.

From these evidences, it is clear the JTW and BBS would be stable if fire recurrence intervals are longer than our chronosequences. Perhaps, the most significant evidence supporting long fire intervals is the exclusive occurrence of fires during wet years, as deduced by Humphrey (1974). In creosote bush scrub, most burning is phased with above-normal precipitation because fires are carried by flashy, exotic fuels whose productivity varies proportionately with rainfall and above-ground biomass is mostly decomposed annually (Brown and Minnich 1986; Rogers 1986; Rogers and Vint 1987; Schmid and Rogers 1988). Fires carried by the subshrub/bunchgrass layer in Joshua Tree National Monument were also phased with above-normal precipitation (Table 7). Earlier burns in 1942, 1967, and 1968 also followed wet years (USDC, 1942, 1967-1968).

This trend is unexpected because the growth of canopy and establishment of subcontiguous stands by perennials over long time scales should encourage fire occurrence rates that are driven by fuel build-up rather than annual precipitation, as in California chaparral (Minnich 1983, 1989). I hypothesize the correlation between fire incidence and rainfall is manifested in spurts of rapid growth and fuel accumulation, which increase short-term regional fire probability. Fuel build-up over long succession times is then diminished by dieback of the ground layer as a consequence of drought. For example, *Hilaria rigida* experiences heavy dieback under deficient precipitation as a consequence of low productivity and breakdown of dry leaf blades. Perhaps mortality of suffrutescent subshrubs decreases understory. Periodic dieback is also consistent with strong resprouting traits of understory species, which is an evolutionarily conservative adaptation unrelated to fire (Wells 1969), but may be critical in an environment of recurrent severe drought, flash floods, and aeolian disturbance. Although biological decomposition is inefficient in deserts (Olson 1963), I propose that dead fuels may also be shredded by wind action, and blown away. Aeolian saltation of fuels may break down undecomposed litter into nonflammable size particles that can be utilized by microfauna. It is doubtful that the occurrence of large fires is generated by the temporal distribution of ignitions. Lightning detection rates resulting from summer thunderstorms average 0.2-0.5 km⁻¹ yr⁻¹ in the mountains of southern California (Minnich et al, 1993). Clearly many discharges have failed to initiate fires

during dry years. Most large fires in Joshua Tree National Monument were set by lightning.

It is concluded that large burns are restricted to special conditions of unusually wet weather and fuel build-up. Site fire intervals are long because: (1) wet weather is rare; and (2) not all sites are burned during a wet period, as during 1978-84. As a consequence, the current rates of burning are insufficient to produce directional changes in JTW and BBS.

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Rhinoceros in Lanfair Valley

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Setting

The Hackberry Mountains are a sequence of volcanic rocks that accumulated at the eastern margin of the Woods Mountains caldera. They lie west of the Vontrigger Hills at the southern margin of Lanfair Valley, 30 miles west-northwest of Needles and 30 miles south-southwest of Searchlight. Access to the Hackberry Mountains is by Ivanpah Road leading north from Goffs..

Crystalline basement rocks in the eastern Mojave Desert include Proterozoic gneissic rocks intruded by Mesozoic plutons of the Teutonia Batholith (older than 100 Ma, Beckerman and others, 1982) and, to the south, granitoid plutons that date to around 80 Ma (Miller and others, 1991). Through portions of the first half of the Tertiary period these rocks were exposed to weathering and erosion (Miller, this volume). Pediments and terrain with relatively low relief were developed and can be seen preserved under volcanic deposits related to extensional tectonics. The Peach Springs Tuff (PST) (18.5 Ma, Nielson and others, 1990) constrains time and topography in the Lanfair Valley area. It often sits directly on the mid Tertiary erosional surface.

The Peach Springs Tuff crops out north and west of the Hackberry Mountains at the Providence Mountains, at Pinto Mountain, and at Barnwell. The Hackberry Mountains are composed of Woods Mountains volcanic center (WMVC) volcanic rocks that overlie the PST (McCurry and others, this volume). The WMVC rocks consist of precaldera Hackberry Spring volcanic rocks (18.5 to 17.8 Ma), caldera-forming Wild Horse Mesa Tuff, ignimbrites (17.75 to 17.73 Ma), and post-caldera volcanics of Tortoise Shell Mountain, dating to 17.6 Ma (McCurry and others, this volume). Basalt flows as young as 10.2 Ma overlie and provide a terminal constraint on the Woods Mountains volcanic center.

The margins of the Woods Mountains volcanic center contain fluvial and lacustrine sediments. The Summit Spring Formation (McCurry, this volume) provides clast source and current directions that help describe the surrounding terrain. The Winkler Formation, lying between the PST and the WHMT, provides a record of plant and animal life, as do the Miocene older lacustrine sediments (McCurry, 1985) at Hackberry Wash. The latter produce a fauna that corresponds to portions of the eruptive sequence of the Woods

Mountains volcanic center. North of the Hackberry Wash locality, silicified sediments (TYLS of McCurry, 1985) apparently overlie and are younger than the WHMT. A second phase of ash rich lacustrine deposits ponded in the same area as the 18 Ma Hackberry Wash sediments. This suggests that the eastern margin of the WMVC was relatively stable between 18.5 Ma and 17.6 Ma.

Paleontology

In the 1960s, hobbyists collecting jasper and opalite from Hackberry Wash located the first vertebrate fossils from Miocene rocks of the Eastern Mojave Desert (Maggie McShan, Needles, personal communication to Reynolds, 1995). Brought to the attention of John F. Lance, the fossils were deposited in the collections of the University of Arizona, Department of Geology.

Systematic investigations and cyclic prospecting of fossil localities by the San Bernardino County Museum in the 1970s and 1980s at Hackberry Wash, Wild Horse Mesa, the Little Piutes and Sacramento Mountains produced additional faunal and floral remains. Local faunas have been recovered from the Hackberry Wash locality (SBCM 1.26.1) associated with the 17.8 WHMT and from the Wild Horse Mesa locality (SBCM 1.18.2) above the Peach Springs Tuff (18.5 Ma) and below the Wild Horse Mesa Tuff (17.8 Ma). Thus, two roughly concurrent lacustrine basins on the southeast and northwest margins of the Woods Mountains



Figure 1. Rhinoceros (*Menoceros* sp.) lower jaw. R.E. Reynolds photograph.

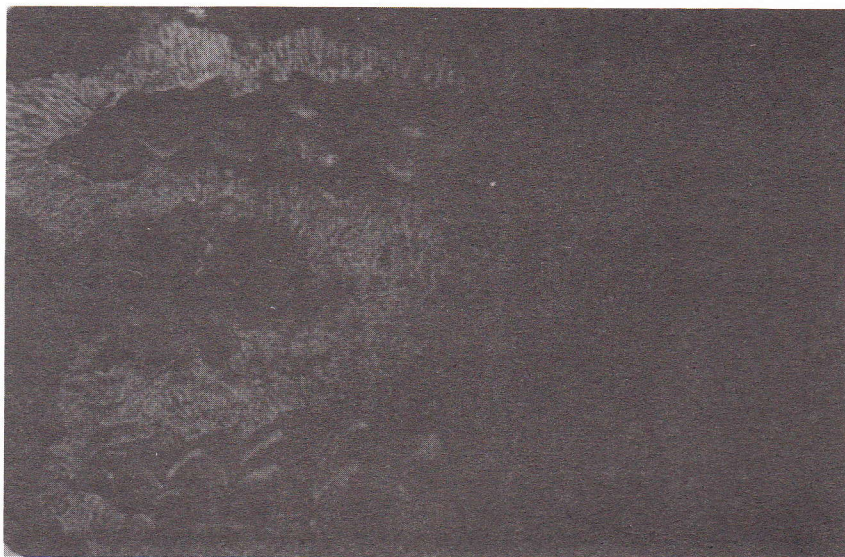


Figure 2. Camel (*Miolabis* sp.) maxillary dentition. R.E. Reynolds photograph.

volcanic center have produce fossil faunas. The Hackberry Wash fossil locality is in sediments that represent two superposed lake deposits. Debris from a breccia dome probably created ponds along a drainage. The lacustrine sediments crop out for approximately one mile along Hackberry Wash and the two superposed lacustrine horizons appear to represent two different but temporally related ponding events. The Mt. St. Helens eruption provides a similar scenario: 1) debris flows blocked a drainage and formed a pond, and 2) rainfall washed carcasses of animals suffocated by volcanic ash into the small lake. At Hackberry, these lake sediments, rich in volcanic ash, were consolidated and compressed, and ground water percolation silicified the sediments around the fossils to form resistant jasper and opal.

Sediments deposited above the Peach Springs Tuff and below the Wild Horse Mesa Tuff contains two faunas. The Hackberry fauna consists of mammals that are useful for dating the lacustrine sediments. The WHM fauna contains rare plants and ichnofossils (fossil tracks).

Methods

Vertebrate fossils at Hackberry Wash are found in silicified shale interbedded with jasper. Most fossils were removed by splitting the shale. In one place, five feet of jasper had to be removed to recover a rhino skull. Shale surfaces were scraped and shale slabs were allowed to weather into catchment basins to accumulate concentrations of finegrained sediments. At the wash site, the sediments were reduced by saturation with kerosene and replacement with water to facilitate the decomposition of the siliceous silt. The silt was then screen washed through 20-mesh and 40-mesh screen. The clean concentrate was processed through zinc bromide, a heavy liquid, which causes the permineralized fossils to sink along with grains of metallic oxides. Manual sorting under binocular microscopes produced the small mammal bones and teeth from these silicified sediments.

Elements of the Hackberry fauna were compared to

specimens from Hemingfordian localities in the collections of the San Bernardino County Museum, the University of California, Riverside, and the Nebraska State Museum..

Fossils from the Wild Horse Mesa were collected on the surface or by splitting shale.

Wild Horse Mesa Fauna

Sequoia langsdorfi

Wood

Conifer needles (A. Sanders, U.C.

Riverside herbarium, pers. comm. to Reynolds, 1987.)

Ostracods

Flamingo footprints

Discussion of Fauna

Footprints of a large flamingo like waterbird are found in the Winkler Fm. at the north end of Wild Horse Mesa. The wading bird left tracks on the margin of the pond in water less than 1.5 feet deep and probably only a few inches deep. Similar tracks were left along shallow lakes over a wide portion of the Mojave/ Colorado extensional province during this time. Large wading bird tracks are helpful in defining environmental conditions during early Miocene extension but are probably not useful as temporal indicators. Flamingo tracks have been found at Boron (SBCM 1.137.1, Whistler, 1984) and Whipple Mountains (SBCM 1.36.x, Beratan, 1992). Wading bird tracks of Hemingfordian LMA are found in the Little Piute Mountains to the southeast (Reynolds and Knoll, 1992).

Plant debris and logs of *Sequoia langsdorfi* (Hazard, 1954) and conifer suggest their source area was a well drained, local highland.

Hackberry Local Fauna

Ochotonidae

Canidae

Tomarctus sp.

Felidae

Camelidae

Miolabis

Aepycamelus?

Dromomerycidae

Aletomeryx sp.

Rhinocerotidae

Menoceras

Heteromyidae

cf. *Trogomys*

Discussion of Fauna

Ochotonidae.

A pika tooth was recovered from concentrate. This partial dP/4 is quite narrow (1.3 mm) in transverse dimension. Three ochotonids are known from the Hemingfordian LMA (Savage and Russell, 1983): *Oreolagus*, *Cuyamalagus*, and *Gripholagomys*. If size comparisons between deciduous premolars and adult premolars are relevant, then this specimen might be *Gripholagomys lavocati* (Green, 1972), the smallest of the Hemingfordian pikas. Until better material is available, however, the specimen is referred only to family.

Heteromyidae: *Trogomys* sp.

One rodent taxon is represented by three lower molars. It is a mesohypsodont heteromyid with crown height higher than *Perognathus furlongi* and lower than *Cupidinimus*

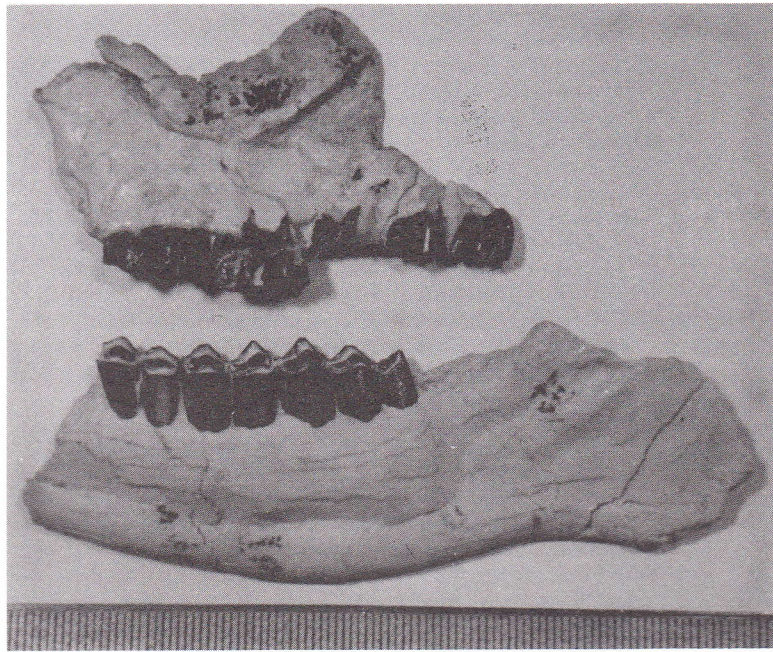


Figure 3. *Aletomeryx* sp. upper and lower jaws. R.E. Reynolds photo.

boronensis or *C. halli*. The crown height compares best with *Peridiomys* sp. (Lindsay, 1972; Reynolds, 1991b) and with a species of *Trogomys* from the Hemingfordian LMA Crowder Formation in Cajon Pass (Reynolds, 1991a). The Hackberry specimens are worn so that the central valley extends lingually only halfway across the tooth. Lingually, the metalophid and hypolophid are worn to distinct lobes: a "W" shape, as in *Trogomys*, rather than the single-lobed "U" shape characteristic of wear in *Peridiomys*. The anterolingual cingulum connection to the protoconid is angular in the Hackberry material, as in *Tromomys* sp. On these bases, the Hackberry heteromyid is tentatively referred to *Trogomys* sp. The *Trogomys* from Cajon Pass will be described as a new genus that is morphologically distinct from and later in time than *Trogomys rupedimenthi* (Reeder, 1960). The Crowder *Trogomys* occurs with *Menoceras* (see below).

Carnivora.

Carnivores from Hackberry include a coyote-size dog and a bobcat-size cat. The felid remains include most of a forelimb. The dog, *Tomarctus* sp., is represented by a lower jaw, isolated teeth, and carpals and phalanges. The size and morphology of dentition matches forms from the Runningwater Formation in the collections of the Nebraska State Museum which range in age from 18.8 to 17.5 Ma.

Artiodactyla.

Hackberry yielded an antelope-like cervoid and two camels.

The cervoid compares well with *Aletomeryx* sp. from the Runningwater Formation of western Nebraska. The Hackberry specimen is somewhat higher crowned with a more rapid rate of eruption of the lower molars.

The medium size camel is similar morphologically to, but two-thirds the size of, *Hesperocamelus alexandrae* (Davidson, 1923) from the Barstovian LMA of California. The premolars are unreduced and approximately equal

in length. The molars have rounded labial cusps and no lingual stylids. Nomenclature problems between *Aepycamelus* and *Hesperocamelus* are unresolved. Pending further review, the Hackberry specimen is probably best referred to a Hemingfordian *Aepycamelus* species on the basis of size and dental morphology.

Two Hemingfordian LMA taxa of small camels are found in the Mojave Desert. They occur in the Little Piute Range, southeast of Hackberry (Reynolds and Nance, 1992), with *Menoceras* in the Cady Mountains of the central Mojave Desert (Miller, 1980), and as far west as Cajon Pass, where they occur with *Menoceras* in the Cajon and Crowder formations. *Michenia* sp. is 20% smaller than *Miolabis*. *Michenia* has a robust lower jaw, and a short diastema with a prominent, double-rooted P1/ (Frick and Taylor, 1971; Honey and Taylor, 1978). *Miolabis* has a slender, longer jaw with a long diastema. The *Miolabis tenuis* from Hackberry compares well with the early Barstovian LMA *Miolabis* from the Cajon Formation. The Barstovian forms, however, are more derived, with stylids stronger and premolars slightly more reduced. The metapodials of *Miolabis* remain unfused while those of *Michenia* are fused over the proximal 2/3

or their length. *Miolabis tenuis* and *Michenia agatensis* occur together with *Menoceras* in the Hemingfordian LMA portion of the Crowder Formation (Reynolds, 1991a).

Rhinocerotidae.

Rhinoceros fossils are rarely found in California. Rhino material from Hackberry includes a humerus, a patella, articulated metacarpals, isolated lower jaws, and a somewhat flattened skull with an articulated lower jaw. *Menoceras* was the size of a modern tapir. It had a large, possibly prehensile upper lip, and a dual horn on the tip of the snout. The limbs and metapodials were relatively long, suggesting that it was an efficient runner.

Teeth in the lower jaw of *Menoceras* are higher crowned than *Diceratherium*. A modest cingulum is present on all lower premolars and molars. On the posterior-labial side of the tooth, the cingulum parallels the base of the enamel. When it reaches the antero-labial portion of the protoconid,



Figure 4. Carnivore (*Tomarctus* lower jaw. R. E. Reynolds photograph.

it runs dorsally up that surface, almost at right angles to the horizontal portion. Similarly, the lingual cingulum runs sharply vertical on the anterior and posterior margin of the cheek teeth. The hypsodonty, narrow cheek teeth, and the configuration of the cingulum separate *Menoceras falkenbachi* from *Diceratherium*. The size and morphology of the Hackberry material is suggestive of *M. falkenbachi* rather than the smaller *M. arikarensis*. The lower jaws of rhinos from Hackberry compare closely to specimens of *Menoceras falkenbachi* in the Nebraska State Museum collections that were recovered from the Bridgeport Quarries in Nebraska and date between 18.8 and 18.0 Ma.

SBCM collections contain *Menoceras* remains from similar age sediments that lie to the west of Hackberry and east of the San Andreas Fault in the Mojave Province. The Horse Quarry in the Cady Mountains (SBCM 5-58-20, RV 6631) has produced a *Menoceras* patella. In Cajon Pass, the lower Crowder Formation (Reynolds, 1991) has produced an elongate right metacarpal II, and Hemingfordian LMA sediments from Unit 3 of the Cajon Formation yielded a partial molar (Wagner and Reynolds, 1983; Woodburne and Golz, 1972; Woodburne, 1991). Thus, *Menoceras falkenbachi* in Southern California deposits appears to be a reliable indicator of the early Hemingfordian LMA of Southern California.

Summary

The Hackberry Wash local fauna (SBCM 1-26-1) contains taxa that compare well with mid continent forms from a period between 18.8 to 18 Ma. The Hackberry fauna is constrained in age between the 18.5 Ma Peach Springs Tuff and the 17.8 Ma Wild Horse Mesa Tuff dated by McCurry (this volume). They also compare favorably with forms from the Hemingfordian LMA Crowder Formation in a portion of that formation older than 17.5 Ma (Weldon, 1985; Reynolds, 1991a). The presence of *Menoceras*, certain artiodactyls, *Tomarctus*, and rodents is useful in dating isolated faunas in southern California.

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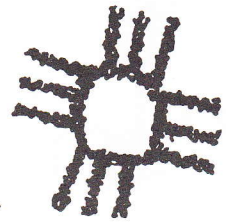
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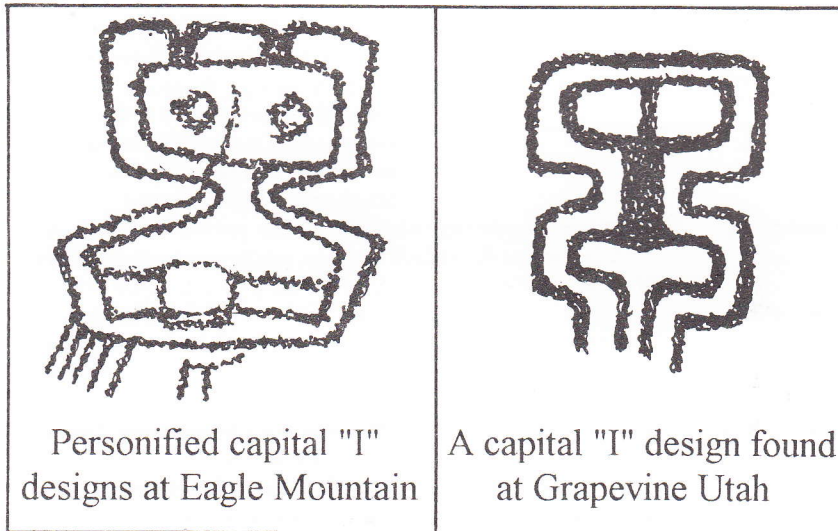
Petroglyphs at the Eagle Mountain Site

Wilson G. Turner, 10443 Lundene Drive, Whittier CA 90601



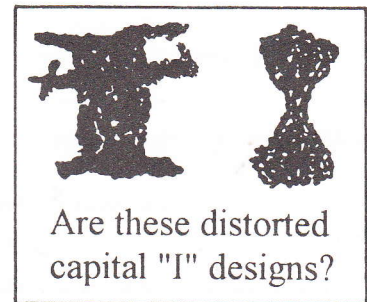
Nearly every vestige of the town of Lanfair, located in Lanfair Valley at the crossing of the Old Government Road and a dirt road running north and south called Ivanpah Road or Lanfair Road, has been removed. There remains a lone telephone booth isolated near this intersection in the "middle of nowhere," and believe it or not I have had to wait my turn to make a call home.

The petroglyphs of the area are very interesting. Although a few resemble the typical Mojave desert glyphs, many are much more like the glyphs along the Colorado River. The Capital "I" design, for example, is found in the Nevada Grapevine site. (Fig. 1) Some of the patterns are very reminiscent of the work done by the Navajo, such as the sun disk which has four short



Personified capital "I" designs at Eagle Mountain

A capital "I" design found at Grapevine Utah



Are these distorted capital "I" designs?

Figure 1. "Capital I" designs at Eagle Mountain and Grapevine, Utah.

Slightly northeast of this junction is a small hill named Eagle Mountain. One section of it is almost entirely made up of basaltic boulders which prehistoric people covered with petroglyphs. The boulders have exfoliated badly from the high temperatures and freezing weather. This has caused great damage to most of the carvings.

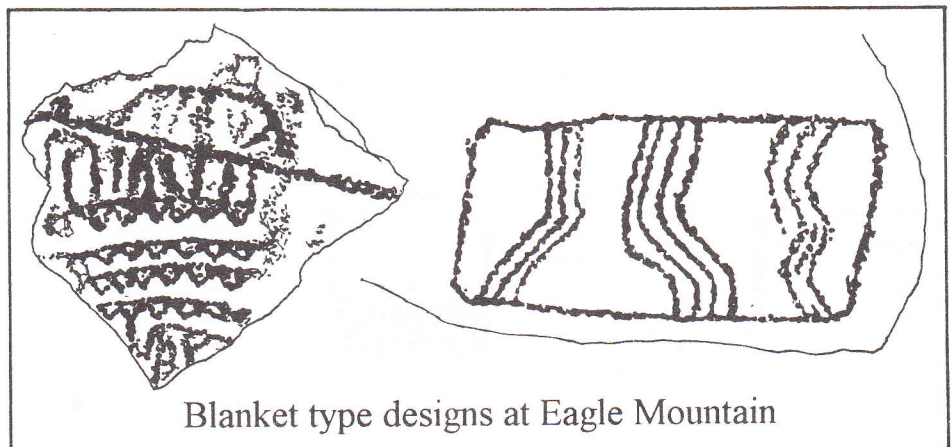
There is an interesting spring, or well, at the site. The rectilinear shape of the boulders forming its sides are of natural origin, although it is apparent the water source has been developed. The last time I was there the water level was so low that for all intents and purposes it was unattainable.

Near the well, and around the base of this western point of Eagle Mountain, are grinding slicks where the native peoples, living in or passing through this area, prepared their food. These polished surfaces are all on bedrock.

lines extending from each of the cardinal directions. There are a few blanket-like patterns which seem to be adjusted to just fit the face of the rock they are carved on. (Fig. 2) We also tend to equate these with the Four Corners area. Although we do not know that there is any connection between this site and the Indians of the Four

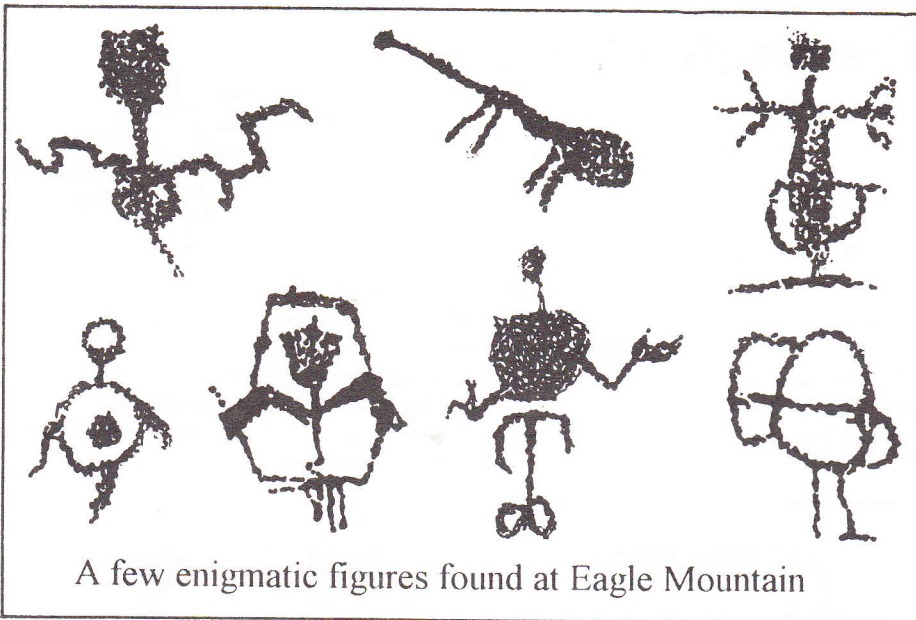
Corners, the turquoise mines about 45 miles to the northwest were mined by the Pueblo Indians who probably came through the Lanfair area to get to the excavations.

There are a number of enigmatic little figures at this site unlike any I have found elsewhere. I can't call them anthropomorphic or zoomorphic for I cannot even be sure



Blanket type designs at Eagle Mountain

Figure 2. Blanket-like patterns at Eagle Mountain, designed to fit the rock face.

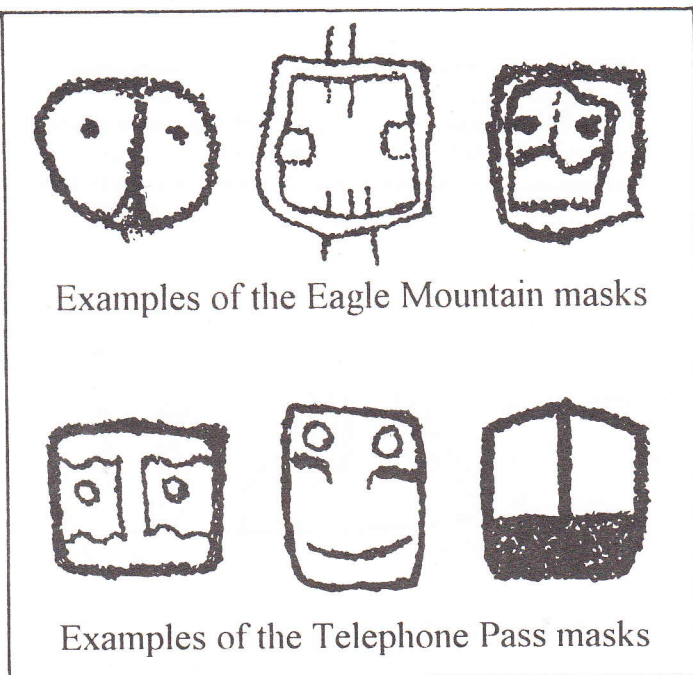


A few enigmatic figures found at Eagle Mountain

Figure 3. Figures at Eagle Mountain petroglyph site.

that they are life forms. Some seem to have heads and some don't. (Fig. 3)

There is also a sequence of so-called masks and pseudo masks here. (Fig. 4.) I find this type of design many places throughout the Mojave Desert. Although a rectangle or circle is the usual basic form used in creating a petroglyph mask, there seem to be many different overall shapes which can be used. In Fig. 1 the capital "I" has been given eyes. Some "masks" located are between here and the Turquoise Mines near Baker, but there are more in the Eagle Mountain area than most other places in the Mojave except for the Telephone Pass site (Fig. 5).



Examples of the Eagle Mountain masks

Examples of the Telephone Pass masks

Figure 4. Mask figures at Telephone Pass site.

Telephone Pass is about six miles to the southeast in the Piute Range (Fig. 5). It could have been visited by the same Indian groups as those of the Eagle Mountain site, for it too would lie on the route the turquoise miners could have followed.

All groups of glyphs at any locale were not necessarily left by the same people. A few years ago, for example, Bob Reynolds and I were able to date some very old petroglyphs in Southern California. We showed through the carbon 14 dating process that some of the glyphs at a Salton Sea site had been created over 9,000 years BP, even though there are many younger glyphs in the same area. This would lead us to assume that petroglyphs in Southern California could



Figure 5. Glyph at Telephone Pass.

have been made over a similarly long period of time. A. L. Kroeber states that when the first white men came through the Mojave Desert and asked the Indians who made the petroglyphs, the answer was "The old ones". To me this translates to, "We don't know just who did them because they were here when we arrived." It is my opinion that the Mojave Desert petroglyphs have not been made by any of the known tribal Indians (such as the Chemehuevi, Paiute or the Mojave) in the last thousand years or more.

Another group which traveled through here and left its mark is modern man. There aren't many of these historic designs, but sometimes the glyphs allow us to trace the identity of the carver and the purpose of his presence in the area. In Black Canyon, for example, the inscription "A. Tillman Sep 30 1874 San Francisco Cala" allowed us to identify a mule skinner from the days of the twenty mule



Figure 6. Historic glyph along Old Spanish Trail.

teams who followed a trail used centuries earlier by Native Americans. Along the old Morro Trail of New Mexico, some of the conquering Spanish, like Agustyn de Ynoios, left signatures and inscriptions which allow us to identify their historic pathways (Fig. 6). In the eastern Mojave Desert, historic glyphs at Fort Piute and Rock Springs recall nineteenth century surveyors, soldiers, and travelers along the Old Government Road.

Leiser Ray Mine, San Bernardino County, California

John Jenkins, *Section of Earth Sciences, San Bernardino County Museum, Redlands CA 92373*

The Leiser Ray mine, also known as the Vanadium Gold Co. (Newman 23:310-311) and the California Comstock (Eric 48:301), is situated in the Camp Signal mining district about 8 miles northeast of Goffs in eastern San Bernardino County, California. Other mines in the area include the U.S. Arbor and the tungsten mines: Kinsman, Lombard & Main group, Mathilde, and Old Glory.

The mine may have been worked primarily for vanadium, but copper, lead, silver, and gold were also produced. Exploration of this deposit began prior to 1891 (Crossman, 1891:18). Much of the mining was done between 1905 and 1915 (Cloudman, Huguein, and Merrill 1917:75-78). Latest mining was done by the California Comstock Gold Mines Ltd. in 1936-1937. Mills were built at the site, but all have been removed.

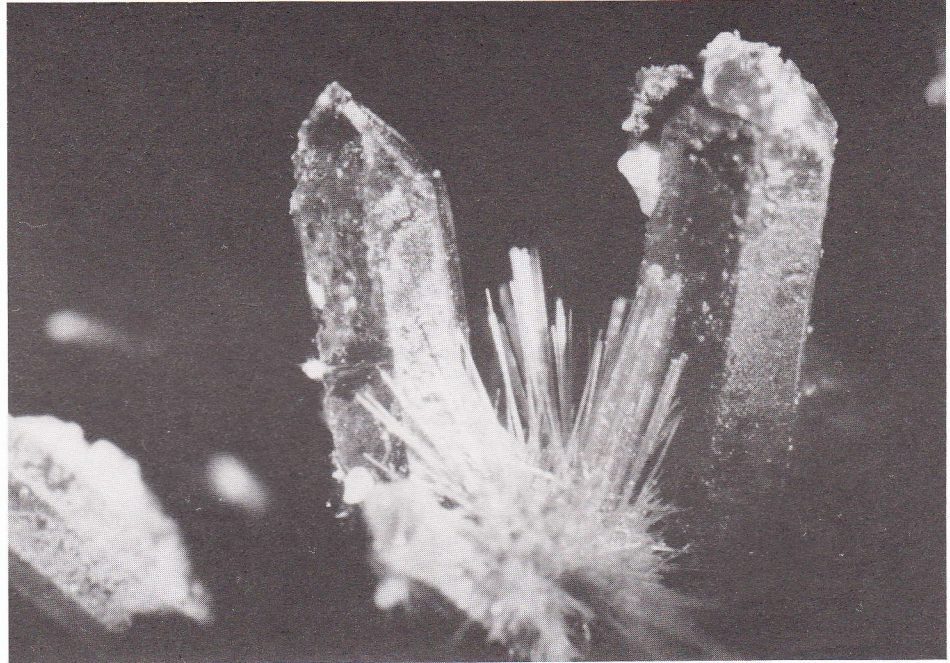


Figure 1. Quartz and olivenite, 30x magnification. John Jenkins photograph.

There is a great discrepancy in descriptions of the mine workings. The California Journal of Mines and Geology (Vol. 9, Jan.-Apr. 1953) reports development to 925 feet by vertical shaft with levels at 212, 250, and 300 feet, and six others at intervals of 100 feet, with a total of over 6000 feet of lateal workings. Another shaft 130 ft. deep is located 187 feet south. Two quartz veins 4 to 12 feet wide in granite were worked. Mineralization was along seams in the quartz. Despite several attempts, the recovery of vanadium was never profitable.

Hewett, in Geology and Mineral Resources of the Ivanpah Quadrangle, California and Nevada (USGS Prof. Paper 275) lists the workings as two included shafts, the deepest to 200 feet,



Figure 2. Wulfenite, 40x magnification. John Jenkins photograph.

on the incline to meet a vertical vein 90 feet below the surface. A 900 foot shaft north of the vein was dug primarily for water. Water now stands at 450 ft. below the surface. Teutonia Quartz Monzonite underlies the area. Four varieties of dikes cut the monzonite. Three of the dikes are aplite and andesite and are pre mineral. The fourth dike, possibly lamprophyre, is post mineral. Two quartz veins occur here, the widest from two to seven feet wide. Mineralization is in the quartz veins. Production records for 1916-1917 report 40 tons of ore averaging 17 percent copper, 1.2 percent lead, 11 ounces of silver, and 0.15 ounces of gold per ton.

Perhaps Hewett's 900 ft. well is C.J.'s 925 ft. vertical shaft. C.J.'s information may have been collected prior to work on the inclined shafts described by Hewett.

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Table 1. Mineral Assemblage, Leiser Ray Mine, San Bernardino County, California

Anglesite	PbSO ₄	An earthy halo around galena. Greyish.
Calcite	CaCO ₃	Colorless rhombohedrons
Cerussite	PbCO ₃	Colorless orthorhombic singles and twins
Chrysocolla	(Cu,Al) ₂ H ₂ Si ₂ O ₅ (OH) ₄ .nH ₂ O	Monoclinic, green, vitreous, non-crystalline
Cuprodescloisite (=Mottramite)	PbCu(VO ₄)(OH)	Orth. Color pale yellow-green to black
Endlichite (=vanadinite)	Pb ₃ (VO ₄) ₃ Cl	Hex. silvery needles.
Fluorite	CaF ₂	Cub. colorless to pale purple
Galena	PbS	Cubic. Color metallic lead grey
Goethite	a-FeO(OH)	Brown; replaces pyrite
Hematite	a-Fe ₂ O ₃	Trig. Black, ruby red in very thin sections
Hubnerite	MnWO ₄	Mon. brown
Hyalite opal	SiO ₂ .nH ₂ O	Colorless, transparent, amorphous
Jarosite	KFe ₃ (SO ₄) ₂ (OH) ₆	Rhom. Brown. occurs in an exposed dike
Malachite	Cu ₂ (CO ₃)(OH) ₂	Mon. Bright green crystals
Mimetite	(PbCl)Pb ₄ (AsO ₄) ₃	Hex. pale yellow
Olivenite	Cu ₂ AsO ₄ (OH)	Orh. Olive green opaque, in tufts of fine hairs to flat crystals
Quartz	SiO ₂	Rhom. (Trig.). Mostly massive; some crystals
Sphalerite	ZnS	Isometric-tetrahedral
Vanadinite	Pb ₃ (VO ₄) ₃ Cl	Hex. yellow
Wulfenite	PbMoO ₄	Tet. Light orange; several habits
Unknown		Yellow crystalline crust on quartz
Unknown		Black needles on yellow crystalline unknown crust (above). Similar to plattnerite; may be "endlichite"
Unknown		White, micaceous, similar to sericite

A History of Piute Pass

Arda Haenszel, *Research Associate, San Bernardino County Museum, 2024 Orange Tree Lane, Redlands, CA 92374*

The earliest recorded reference to Piute Pass was made by Father Francisco Garces in 1776. On his outward trek across the Mojave Desert he had traveled the Mojave Trail almost directly westward from the Mojave Villages at present Needles. Then, returning to the Colorado River, he branched northeast from his outgoing route at the Devil's Playground when the Indians begged him to visit their rancherias, crossed the Providence Range possible through Macedonia and Wild Horse Canyons, and, passing Rock Spring, made a curve around the north end of Lanfair Valley. Then, after following an Indian trail south along the west foot of the Piute Range, he turned eastward through the mountains to a place where there was abundant water.¹ The Mojaves called it Ahakuvilya, and the Chemehuevis Pa'ash,² but Garces did not refer to it by name. He did say that it was Chemehuevi territory. Piute Pass is the only place in the whole area where there is a lot of water, however, and the direction and distance are correct for reaching Sacramento Springs, where he rejoined his outward route.

Perhaps we can say that Garces was not the first to record the pass, though, for Malcolm Rogers of the San Diego Museum of Man, examining the site in the 1920s, found evidence in the petroglyph styles there (Figure 1) of temporary camping by San Diegito I people from 9000 B.C. Then there was temporary camping by Amargosa II people, whose petroglyphs are most numerous, around 3000 B.C. Pre-historic Yumans also did a lot of rock writing there about 1000 A.D. The Chemehuevis, he found, didn't make many petroglyphs, but evidence of their settled occupation is suggested by Garces and confirmed by the reports of Whipple and Mollhausen in 1854.³

At the edge of the settled Mojaves' agricultural territory in the Colorado River Valley, the Chemehuevis, with a migratory culture on the desert, still were known to cultivate a few areas in the eastern Mojave where there was enough water. And Piute Springs was one of them. Noted in 1854 were the presence of many tortoise shells and signs that the Chemehuevis raised corn and melons. The latter were typical of Chemehuevi settlements as late as the 1880s.⁴

Father Garces was the first white visitor to Piute Pass. Fifty years later he was followed by the first American, Jedediah Smith.

On their way west in 1826, having struck the Colorado, trapper Jedediah Smith and his party traveled down it to the Indian settlements near the site of later Fort Mojave, about 15 miles north of Needles. After some trading, he tried to follow a trail the Mojaves told him about, starting from Beaver Lake on the western bank and heading directly west, probably through Pictograph Canyon at the north end of the Dead Mountains, to what must have been Piute Springs, where he camped. But his horse was stolen, and the next day he lost the trail in Lanfair Valley.



Figure 1. *Petroglyph near Piute Springs, October 5, 1930. Haenszel collection.*

Returning to the river after another night at Piute, he was able, in a second attempt, to enlist the aid of some run-away Mission Indians as guides, and they took him for the third time to Piute Springs, the regular first stop on the proper route, and then on through the pass.

The next year, 1827, Smith found himself again on the Colorado River and short of provisions. This time the Mojaves were unfriendly, and it is thought that a recent visit by another group of white men had turned them against whites. Smith's suspicions were realized when the Mojaves attacked his party as it was divided in crossing the river. All but Smith and eight of his men were killed. The survivors on the west bank held off the attackers until nightfall, and then fled westward in the darkness to Piute Springs, where they rested all the next day before going on.

In his brief accounts of the two trips Smith gave only enough details to indicate his routes, and there is no description of Piute Pass.

In the 1830s and 40s white travel slowly increased on the early Spanish Trail down the Muddy and Colorado Rivers to the Mojave Villages.⁶ This trail followed much the same route that Smith had taken from the vicinity of the later Fort Mohave westward, making the first stop at Piute Springs.

The Mojaves themselves, and other desert Indians, used to traveling long distances on foot without carrying water and perhaps with only a handful of chia seeds for food, took the more direct version of the Mojave Trail running parallel a few miles south.⁷ But most of the white travelers, New Mexicans and a few American traders and trappers, used the more roundabout Piute Springs, Rock Spring, Marl Springs route because of their riding and pack animals that needed a water supply more abundant and regularly spaced. True, in 1829-1830 Antonio Armijo, directed by the governor of New Mexico, and his scout, Rafael Rivera, had worked out a shorter, diagonal route from the vicinity of Las Vegas to a junction with the Mojave River trail at a point east of the

present Daggett. This diverted some of the increasing traffic, though the old route down the Colorado and west to the Mojave River via Piute Pass was still used.

But the gold rush to California caused a sudden expansion in the need for overland routes, especially those in the south which avoided Sierra snow. The ultimate solution would be the railroad, it was believed. So in 1853-54 Lt. A. W. Whipple led a survey part across the far west along the 35th parallel, one of several such expeditions seeking a practical route for a railroad.

Coming from the Needles area, the Whipple expedition followed Piute Wash around the south end of the Dead Mountains to Sacramento Springs, then angled northwest to Piute Pass. Whipple had engaged Mojave chiefs Iretaba and Cairook as guides, and they steered him to the "white man's" route because he had a large party, many animals, and even a light instrument wagon.

The railroad surveys were not only charged with seeking a route suitable for a railroad. They were also directed to report in detail the kind of country they passed through — the geography, geology, natural history, and inhabitants. The command and escort was military, but the party included civilian scientists as well. And since there were no cameras to record the details, an artist, Baldwin Mollhausen, was engaged. Of special value for historians was Lt. Whipple's personal interest in anthropology and his careful observation and notes.

The official government report of the expedition is divided into sections, four of which deal with Piute Pass. Whipple was the first to refer to the site by that name. In his "Preliminary Report" he wrote,

We encamped upon a pretty rivulet, which watered a small valley that had been converted by the mountain Pai-utes into a luxuriant garden. Passing the crest of a hill, and by a gradual ascent over wide prairies of rich grama grass, we reached a rocky glen (Rock Spring).¹⁰

In the "Itinerary" section we find evidence of recent Indian

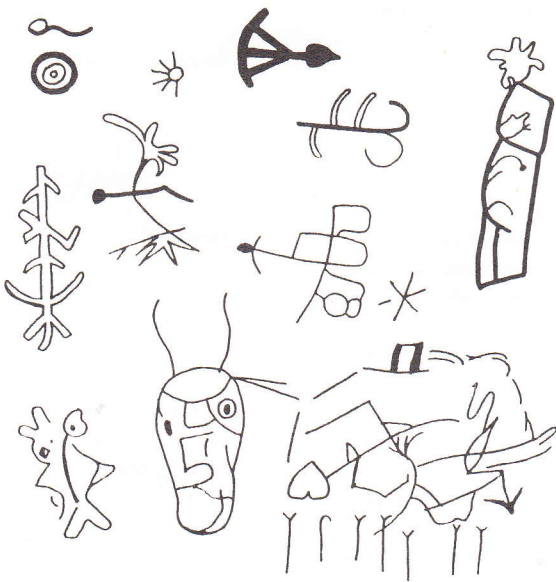


Figure 2. Petroglyphs at Piute Springs in 1853-54 as illustrated in Whipple's Report.

occupation.

Mar. 3, Camp 137

Continuing the survey northwest about 9 miles (from Sacramento Springs) over the smooth gravelly slope, we reached, at the point of a mountain, Paiute creek, a finely flowing stream of water. Finding good grass also, we encamped. A little basin of rich soil still contains stubble of wheat and corn, raised by the Paiutes of the mountains. Rude huts, with rinds of melons and squashes scattered around, show the place to have been but recently deserted. Upon the rocks, blackened by volcanic heat, there are many Indian hieroglyphics. Some of the more simple have been copied. Others are too complicated or too much defaced by time to be deciphered. . . .

Mar. 4, Camp 138

At 8:00 A.M. we filled our canteens and started. For the sake of the instrument wagon, the guide led us up the creek to a deep ravine, from which he ascended and passed over the crest of a sharp dividing ridge to a plain of great extent. He afterwards told us that the pack-train should have kept the ravine, and saved the hill."¹¹

From the section "Topographic Features" comes a description of the route through the pass.

We approach the base of a cluster of sharp crested hills from 800 to 900 feet in height, at the rock base of which flows a rivulet called Paiute creek. Upon its borders, near Camp 137, are patches of fertile soil, which have been cultivated by Indians producing corn and melons. There are cedar trees and grass upon the hill sides. The stream flows southeast The ascent has been gradual, except near the entrance to the creek, where several rough ravines were crossed. . . .

From Camp 137 the trail ascended a tortuous ravine to the head of one of the branches of Paiute creek, and then mounted to the crest of the ridge, about 830 feet above the cultivated fields in the valley. From this point sketches were taken, showing a wide gap between the hills upon the left, and upon the right a low valley, 1/4 mile wide, appearing to drain the waters of the plain, which lay extended towards the west, into the same great valley that receives Paiute creek The course of the trail is nearly magnetic west across a vast plain extending about 20 miles to Camp 139 at Rock Spring.¹²

There were two routes to Lanfair Valley through the upper pass. One was the main gorge on the north, in which rise the springs, which is too narrow for wagons. The other was the south fork canyon trail over the summit, to which the chiefs directed the party.

The "Report Upon the Indian Tribes" describes the Chemehuevis.

West of the Rio Colorado we enter the range of the widely extended Utah nation. Those that roam over the region traversed by us, call themselves Paiutes, and are closely allied to those that massacred the party of the lamented Capt. Gunnison. This band probably does not number above 300 persons. Though supposed to maintain a scanty and precarious subsistence, principally upon roots, they are probably distinct from the Diggers of California. We passed through one little valley of theirs, at Paiute creek, where wheat and melons had been cultivated.

Plates in this section of the Report contain sketches of Chemehuevi Indians, and of their implements, while the text

gives an extended description of their appearance and customs. Plate 38 (Figure 2) contains a group of petroglyph designs sketched from the rocks in Piute Pass.

Mollhausen was a bit more literary in his book about the trip.

Where the mountains began to tower up high above us, we discovered the first traces of water, as a small brook trickled over a few acres of land, and then vanished again in the sand at the end of the valley. Reeds and rushes must at one time have grown luxuriantly at this spot, for on the top of the banks we saw many heaps of them, which had apparently served the Indians as couches. In the valley itself the reeds had been burnt away, but green sprouts bursting here and there from the ground through the black ashes, announced the approach of spring. We thought we had now reached the spot where it would be best to encamp, but the Indians led us still deeper into the mountains, until we came to a small valley, which at that time of year, when it showed only a withered vegetation of grass and shrubs, had no particular attraction; but lying as it did, hidden among lofty rocks, and surrounded by a dreary inhospitable desert, might, later in the spring, or in the hot summer, appear to a traveler coming upon it accidentally as a marvel of fertility. It seems that the Indians cultivate there fields of maize and wheat; everything indicated that at certain seasons it must present an animated appearance, and the number of turtle shells lying about showed this to be a favorite food of the natives of the country

We found ourselves very agreeably situated on the banks of this full-flowing brook, where there was grass for our cattle in the valley, and even on the declivities of the nearest rocks; the sun shone warmly and pleasantly; the dry wind that had troubled us so much on the preceding day could not find us out here, and the soft sand on which we spread our blankets, was a very agreeable change from the sharp stones on which we had to rest our wearied limbs the preceding night, a couch that had left them still sore and stiff.

On the 2nd of March we left the valley with its pleasant spring, and following it when it became a ravine deeper into the mountains, came to its end at a steep ridge. Slowly the Expedition climbed this, until at the summit of the height a wide prospect to the west unexpectedly opened to us. A dreary and inanimate plain lay before us, the monotony of which was not much relieved by the yuccas that

stood pretty thickly towards the west; but what did essentially improve the prospect, was the rocky range of hills that rose behind them, and beyond these again, white glittering peaks, which we took for the southern points of the Sierra Nevada. Descending from the mountain ridge into the plain, we followed our guides in a southwesterly direction, over a tolerably good road.¹⁴

The account in the manuscript diary of Lt. David Stanley, Whipple's Quartermaster, is mainly concerned with the road.

Mar. 3, (1854)

. . . . up a canon in the mountains on the left of the valley we found a good spring of running water and encamped. Little grass — old Indian camp — bad canon

Mar. 4

We moved up the little canon in which we encamped one mile and ascending about four hundred feet, we saw directly to the west a wide plain stretching out, over which our day's march lay.¹⁵

This, then, was Piute Pass in 1854 (Figure 3).

Though actual construction of a railroad would come later, there was an immediate need for a wagon road along the 35th parallel. In 1858 Lt. Edward F. Beale made inspection trips westward and eastward along the route and was awarded a contract for construction of such a road. It was to meet the Colorado River at a point about 15 miles above present Needles that came to be called Beale's Crossing.¹⁶

As the road building party worked its way across Arizona in 1859, Beale arranged to have Samuel Bishop bring supplies by wagon from Los Angeles. As Bishop and his band of hardy frontiersmen reached the Colorado, however, they were attacked by the Mojaves, and withdrew to Piute Springs to reconnoiter. United, the members of Bishop's party were a match for the Mojaves, but they could not



Figure 3. Government Road at the summit of Piute Pass, looking west across Lanfair Valley. Haenzel photograph, April 1965.



Figure 4. *The Government Road in Piute Pass. Wagon wheel ruts are visible at canyon forks. Haenszel photograph, 1963.*

defend themselves while divided in ferrying the supplies across the river.

Meanwhile, because of repeated and disastrous attacks by Indians on immigrant parties and other travelers along the Mojave River and in the vicinity of the Colorado, Col. William Hoffman had assembled a huge force at Ft. Yuma, and was even then marching up river to subdue the Indians in Mojave Valley and establish a fort there for protection of travelers.

Bishop sent a message to Hoffman asking for military protection in crossing his supplies.

While waiting for the answer, Bishop set his men to work building the road through Piute Pass (Figure 4), making that stretch the first part of the Mojave Road to be built in California. It must have been at this time that Bishop carved his name in fancy letters on a boulder forming part of a wall in the canyon.

Hoffman refused to send any troops ahead to help Bishop. Beale badly needed the supplies, so Bishop could not wait for Hoffman to arrive. He decided to send his wagons back to Los Angeles, load what he could on pack animals, and cache the rest at Piute Springs for later recovery by Beale. The pack train was then able to use a trail recently reported by Aubry, and cross the river at a point above the Mojave territory.

Hoffman and the troops finally arrived, overpowered the Indians, and established Ft. Mojave. Leaving some soldiers to

man the fort, he took a few back with him to Yuma by steamboat, and sent the rest across the Mojave Desert by the Whipple route.

The overland party, by this time short of rations, discovered the Bishop cache when they got to Piute Springs, and proceeded to steal it. The road builders, coming along later, even more in need of the supplies by the time they reached Piute, found the cupboard bare and had considerable hardship in making their way on to the settlements. Beale¹⁷ was never able to get financial restitution from the army.

It is interesting to note that camels were prominent in the Beale party of 1859 as part of a government experiment in the use of them on the American desert. Almost 100 years later, George Irwin found the remains of a camel pack saddle in the pass.

The first forts, really mere redoubts, were built along the Government Road to Fort Mojave in 1860 under the direction of Maj. James Carleton. They were Camp Cady, west of Afton Canyon, and Hancock Redoubt at Soda Springs on the west edge of Soda Lake. Camps Marl Springs and Rock Spring were added later, as Indian attacks along the road were fairly frequent at this time. These rather rude outposts were built as stations for small groups of soldiers who served as escorts for travelers on the road. They included mail and express riders, stage drivers and passengers, wagon freighters, miners, and immigrants.

In the winter of 1867-68, Lt. J. B. Hardenburgh was sent with a detail to build the last of these outposts on Piute Creek. Having not the faintest idea how to go about it, he and the troops were marking time when there arrived a military inspection party led by Maj. Gen. Irvin McDowell, Commander of the Department of California, and Maj. Henry Martyn Robert, Chief Engineer of the Division of the Pacific. They immediately sized up the bewildered lieutenant's predicament and stopped over a day to remedy the situation. They located a site, designed the building, and instructed the men how to begin. Taking the unfortunate Hardenburgh along with them to Ft. Mojave,¹⁸ they sent back a knowledgeable sergeant to direct the work.



Figure 5. *Fort Piute in October, 1930. Haenszel collection.*

Organized by the commander, designed by the top engineer in the West, and built by efficient and highly-motivated troops, Ft. Piute was the best-planned and best-built of all the outposts on the Mojave Road. Constructed of rock on a site overlooking the road, it commanded the east mouth of the pass (Figure 5). In the plan there were three connected rooms, the one in the center intended for a corral, and those on either side for living quarters for the men. The east room, of which merely the foundation remains, was the smallest. Somewhat larger, the west room contained a fireplace. The entrance seems to have been to the middle room on the north side through a doorway which was protected by an L-shaped rock wall.

This fine building was occupied by troops mostly during the period of its construction, however. In January 1868 the San Bernardino *Guardian* reported that a lieutenant and 18 men were stationed at Piute Springs, with a relay of horses for mail escort.¹⁹ But later that year the official mail route was shifted from the Mojave Road to the Bradshaw Road farther south, and Ft. Mojave was then supplied by river steamers.

From 1868, Ft. Piute served basically as a relay station for remaining traffic, which changed somewhat in character as mines were opened all over the desert in the 1870s and 1880s. It was probably at this time that the outer corrals were built.

In 1883 the Southern Pacific Railroad completed a line from Mojave to Needles to meet the Santa Fe building westward to the Colorado River. It followed a more direct route from the Mojave Sink to the river, somewhat south of the Mojave Road, as Whipple had tentatively suggested. Unlike the wagon road, and like the Indians, the rails did not need to wander from water hole to water hole. Local roads, and ruts broken out along the route by railroad construction gangs, eventually were augmented and joined into a cross-country route paralleling the tracks. Water and help could be obtained at the little railroad stations at more frequent intervals, and the Mojave Road fell into disuse, except by occasional travelers or in certain sections. The new route evolved, in the second decade of this century, into the National Old Trails Highway. Piute Springs was bypassed.²⁰

Not entirely abandoned, however, Piute Springs in the 1890s was on occasion the camp of Indians, miners, and some travelers. Needles pioneer, Charles Battye, told of an intended meeting of friends Frank Howard, Johnny Madden, and himself "at Howard's mining claims in the Piute Springs country." This was a tale with a sad ending, for they found Johnny's body on the trail from Vanderbilt.²¹ Likewise in the 1890s, Piute Spring was the home of Johnny Moss, known for the discovery of several important mines in the east Mojave.²²

In the early 1900s the road over the hill was said to be impassable to vehicles. But access to



Figure 6. Mud and wattle hut at Piute Springs, 1928. Haenszel collection.

the canyon over the Mojave Road from the east was still open. I know that to be a fact because around 1920 my parents and I were driven over it in a Hupmobile by friends from Searchlight, where my father served as the doctor for the Santa Fe Railroad and the town. At that time my father photographed a mud and wattle house built near the mesquite grove in Chemehuevi fashion (Figure 6), as well as the remains of the fort and the corrals. In 1924 the Chief Engineer of the California Highway Department Section VIII, E. Q. Sullivan, camped at Piute Springs while resurveying parts of the old road.²³

In the 1920s the park-like wooded area beside the stream in lower Piute Canyon was a recreation spot for desert residents. The Gus Swearingens and their friends from Needles visited the fort in 1922 and took pictures of it (Figure 7). And in 1925 I remember an outing on which my girl friend and I were treated by friends from Searchlight to a picnic of chicken barbequed there in the canyon beside the stream.²⁴

The area at the mouth of the canyon below the fort was homesteaded in 1928 by Thomas Van Slyke, a miner. During World War II, when Patton's trigger-happy trainees in their armored vehicles ranged up into the Piute area practicing for the African campaign, legend has it that Van Slyke stood off



Figure 7. The Swearingen family and friends beside the fireplace at Fort Piute in 1922. Haenszel collection.



Figure 8. Although the walls were crumbling, the loopholes remained at Fort Piute in 1922. Haenszel collection.

the African campaign, legend has it that Van Slyke stood off a crew that wanted to use the old stone fort building as a military target.²⁵

The George Irwin family bought the ranch from Van Slyke in 1944, and for two years tried raising turkeys. It was an ill-fated venture, for the local coyotes grew fat, and the Irwins, though they loved the place, were forced to leave the remote spot for health reasons.²⁶

A number of exceptional photographs of the Piute Pass area taken by Walter Fiss are included in a WPA study of the Mojave Road directed by Josephine Rumble which appeared in 1939.²⁷

Informed of the outstanding value of the scenic site in the fields of geology, natural history, prehistory, and history, in 1968 the Bureau of Land Management made a preliminary survey of the pass, but had difficulty trying to furnish protection for the area.

Through the efforts of the San Bernardino County Museum in 1973 the entire pass was accepted for the National Register of Historic Places as a historical district.²⁸

In 1974 vandals toppled the stone containing Bishop's inscription down the cliff and into the stream bed, breaking it into three pieces. When this was discovered by Dennis Casebier, the Bureau promptly raised the fragments by helicopter, fastened them together, and eventually put them on display at the Mojave River Valley Museum in Barstow until such time as the inscription would be returned to its original place at Piute under proper protection.

The earliest known photograph of the fort and corrals was taken by a Needles man possibly around 1919. This picture shows the walls of the middle and west rooms virtually intact, though the east room had begun to disintegrate. The fireplace wall of the west room, still complete, indicated that the structure had had a peaked roof. By 1922, photographs show that the wall had crumbled down about half way to the top of the fireplace, but all the walls were high enough to preserve the loopholes (Figure 8). Disintegration had proceeded slowly for about 40 years (Figure 9). However increased visitation by large and small parties, including both deliberate and unintentional vandals, in a steady stream since the 1960s, has done more extensive damage to the structure than in the 100 years

preceding.

There are a number of historic remains still to be seen in Piute Pass. The old road is still visible for much of the route through the pass and across adjoining Piute and Lanfair Valleys. Save where it passes the fort, it ran in or beside the stream through the lower canyon. But still apparent are the wagon wheel ruts worn six feet apart in the soft red formation where the road came up out of the stream bed into the south fork canyon. Approaching the summit, there are even three separate alignments, indicating efforts to improve the grade at the steepest part. Piute Hill was notorious as the worst hill on the road in California.²⁹ The steep parts have been so eroded by storms that, though their course is still clear, here and there it is impossible to travel on them even on foot.

Beside the fort itself there are remains of other possibly historic structures still visible in the lower canyon. No trace of the Chemehuevi hut remains, but there are remnants of at least three Indian "house rings" below the fort (Figure 10). These may date from the 1850s when



Figure 9. The fireplace at Fort Piute in February, 1963. Haenszel photograph.

Whipple mentioned the presence of brush huts.

Just below the fort also is a rectangular masonry wall of patinated native rocks two or three feet high which might have served as an outlying breastwork supporting the fort, or as a small corral. Halfway up the hillside from it, and overlooking the fort and the road approaching from the east, is another low wall which may well have been a breastwork.

Across from the mesquite grove there is a stone corral with a gateway facing the road. But the walls of the corral leading down the hill from the fort have crumbled, and the lower part has been washed away by floods in the tributary draw.

"Bishop's Wall," an L-shaped rock structure whose long side parallels the trail, and short side makes a right angle toward the edge of the creek bed, is the deepest mystery. Artifacts indicating a possible aboriginal camp or work site have been found within the L, under the trees and beside the creek. Obviously the spot was known to the natives. Did they build the wall for protection against either weather or possible enemies? A number of petroglyphs occupy prominent exposed faces of boulders forming the wall, and there is at least one metate slick on the top surface of one of the rocks. Did Bishop's men or other white visitors build the wall for some purpose, carefully arranging the decorated boulders to display the petroglyphs? In the upper canyon there is a precedent for this, as there are instances in which rocks used to shore up the road have been placed so as to show the petroglyphs on them to advantage.

This much is known about the history of Piute Pass. The place was unique enough with its flowing stream and terrible hill to attract attention of visitors through the years and cause them to remember and comment on the place. But the outlying rock structures are not the only mysteries. A rock-covered grave with a nameless headstone indicating burial by white men, was discovered near the summit in 1962. The overlying rocks were removed and a much disintegrated skull was partially uncovered. It had been broken by a blow on the head. Because of their fragile nature, the skull and the rest of the skeleton were not excavated, and the grave was returned to its original appearance. There were accidents on the road down this terrible hill. Did some traveler suffer a fatal fall? Or was he the victim of an unsolved murder? No historical record of his death has been found. No one knows who or what he was, what happened to him, or when he died.³⁰

It was a hard land, and life in this remote part of it was precarious. The strong survived, and some of them told us about their experiences at Piute. To this extent can be reconstructed a history of Piute Pass.

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Grandview Gorge: Research Involving the Mid Hills Tectonic Block

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There is a spectacular chasm at the eastern margin of Lanfair Valley on the west side of the Piute Range. The expansive flat surface of Lanfair Valley is deceptive and it is easy to drive by this picturesque gorge — which we did on several trips — unless you consult a topographic map and ask, "What is a hole this size doing in Lanfair Valley?"

During a San Bernardino County Museum trip to investigate "Quaternary lake sediments" (Jennings, 1961) on the west side of the Piute Range, we discovered a canyon 5 miles long, 1½ miles wide and 200 feet deep. To distinguish this astounding north-south exposure from the east-west trending Piute Gorge to the east, we dubbed it Grandview Gorge in 1968.

The development of the gorge is controlled by headward erosion constrained on the east by a north-south fault on the west side of the Piute Range (Nielson and Nakata, 1993). The more ragged west margin of the gorge is controlled by up to 27 feet of cliff forming pedogenic carbonates. Stream incision through the playa sediments down to a black volcanic flow controls the depth of the gorge.

The research potential of Piute Gorge has yet to be realized. A recent study of petrocalcic horizons capping the playa sequence indicates that the soil carbonates may have formed in the early Pleistocene, or as long ago as 2 Ma (Katzenstein et al., this volume). The playa sediments were deposited against a fault that postdates and cuts volcanic rock of the Piute Range (8.0 ± 0.6 Ma, Nielson and Nakata, 1993). A lot may have happened during the six million years between the last volcanic flow and the development of the pedogenic carbonate below the stable surface of Lanfair Valley, and clues to the timing may be recorded in Grandview Gorge.

The north-south fault on the west side of the Piute Range is parallel to Mesozoic faults in the area (Nielson, this volume). This fault caused playa sediments to pond west of the Piute Range. The playa sediments contain four horizons of pedogenic carbonate kernels (Figure 1) beneath the massive carbonate cap, indicating that there were stable periods during playa deposition when soils formed across the playa surface. Rather than being finely laminated, some of the playa silts are 10 to 30 feet thick. This may suggest rapid filling during each wet period. It is possible that it took the playal basin no longer to fill than the length of time it took the carbonate cap to form. The age of the north-south faulting is important to the tectonic modeling of the area (Nielson, this volume) and to regional tilting involving the Mid Hills Tectonic Block (Miller, this volume).

The playa sediments contain clasts which indicate the source area for the drainage that filled the playa. The playa sediments are more than 180 feet thick (Figure 1) and clasts throughout indicate a source from the volcanic rocks

of the Castle Mountains and the Willow Wash paleo valley (Miller, this volume). However, the top of the playa section coarsens markedly. These cross-bedded sands and gravels are cemented by pedogenic carbonates (Katzenstein et al., this volume). As clast size and abundance increases, a population of clasts from the Caruthers Canyon and Live Oak Canyon area of the New York Mountains becomes apparent. These clasts include Live Oak granodiorite, Mid Hills adamellite (Beckerman, 1982), quartz dike rock with sulfide minerals, quartz dike rock with limonite pseudomorphs, monzonite porphyry and aplite dikes, and metamorphosed limestone and marble. This influx of central New York Mountains clasts might reflect drainage changes, perhaps due to Pliocene tilting of the Mid Hills tectonic block (Miller, this volume) or might be in response to latest uplift on the Cedar Canyon Fault which elevated the central New York Mountains.

Pleistocene sediments are exposed in wash banks up to two miles west of Grandview Gorge. Poorly indurated sands and gravels approximately 30 feet thick have been deposited above the prominent pedogenic carbonate surface. These sediments contain carbonate filled root casts and minor carbonate kernels. These Pleistocene sediments also contain a suite of clasts indicative of the Caruthers Canyon - Live Oak Canyon source area. This suggests that from the late Pliocene, prior to 2 Ma (Katzenstein et al., this volume), to the mid? Pleistocene, the major source of clasts in this area was from the northwest in the central New York Mountains. Today, active washes that cut these sediments carry clasts from the eastern New York Mountains and the Castle Peak area. This clast mixing may be the result of deainage meander on a flat alluvial plain.

Questions regarding timing of regional tilting and tectonic activity may be answered by looking deeper into information available at Grandview Gorge. Questions that remain unanswered include:

- When was the west Piute Range fault active?
- What is the amount and direction of offset on this fault?
- What is the age of the volcanic flow beneath the playa sediments?
- What was the rate of playa filling?
- Can a time-diagnostic fossil fauna be produced from the playa paleosols?
- Where are the volcanic ashes? The Bishop Ash is preserved in coarse fans of the western Providence Mountains. Where is that tuff, the 2.4 M.A. Huckleberry Ridge Ash, and the 3.4 M.A. Nomlaki tuff (Sarna - Wojcicki, 1976)?
- Can magnetostratigraphy be applied to the section to constrain the times of faulting and rate of playa filling?

After traveling miles across the flat surface of Lanfair

Valley, your curiosity can only be inspired when you stop at the edge of this astounding gorge at the east end of the Mid Hills tectonic block—a gorge that so far is filled with more questions than answers.

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A Preliminary Assessment of Calcic Soil Development at Piute Gorge, Fort Piute Wilderness, California

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Introduction

At the eastern margin of Lanfair Valley, Piute Creek has incised in the valley sediments and through the Piute Range to create Piute Gorge. Lanfair Valley sediments exposed in the gorge consist of fine-grained, poorly-cemented sandstones and mudstones of Plio-Pleistocene (?) age (Nielsen et al., 1987). These sandstones and mudstones are capped with a calcium carbonate cemented conglomerate suggesting the presence of a soil of considerable age. The top of the gorge is approximately 1100 m in elevation with an annual temperature range from -12.2° to 37.7° C and mean annual precipitation varying from 5.6 to 17.5 cm (Nielsen et al., 1987). The vegetation is creosote scrub, including creosote, ambrosia, indigo bush, brittle bush, buckwheat, cacti, and yucca (Nielsen et al., 1987).

The objective of our preliminary study is to constrain the age of the soil developed in the valley sediments through correlative dating of pedogenic carbonate morphology. Gile et al. (1966) observed that carbonate morphology changes with time, thus allowing calcic soils to be a useful indicator of relative age. Relative age correlations assume that the factors affecting the rate of carbonate accumulation — specifically, carbonate influx rate, precipitation, evaporation, temperature, available water-holding capacity and partial pressure of CO_2 — are similar on a regional basis (for further discussion, refer to

Table 1: Stages and diagnostic characteristics of pedogenic carbonate morphology observed in calcic soils in the southwestern United States (after Bachman and Machette, 1977, and Machette, 1985)

Stage	Diagnostic morphologic characteristics in gravelly parent material
I	Thin, discontinuous coatings on pebbles, usually on undersides
II	Continuous, thin to thick coatings on tops and undersides of pebbles
III	Massive accumulations between clasts, becomes cemented in advanced forms
IV	Thin (<0.2 cm) to moderately thick (1 cm) laminae in upper part of KM horizon. Thin laminae may drape over fractured surfaces.
V	Thick laminae (>1 cm) and thin to thick pisolites. Vertical faces and fractures coated with laminated carbonate (case-hardened surface).
VI	Multiple generations of laminae, breccia, and pisolites, recemented. Many case-hardened surfaces.

Table 2: Soil horizons and depths along with pertinent horizon properties.

Horizon	Depth (cm)	Carbonate Morphology	Structure	% Gravel
Bk	0-49	minor, filamentous	coarse, angular blocky	25%
2Bk	49-103	minor, filamentous	prismatic to platy	20%
2Bk2	103-161	58 cm thick laminate	massive plates	
2Bk3	161-205	nodules	nodular calcite 1-3 cm in diameter	<2%
2Bk4	205-325	nodules	nodular calcite 7-8 cm in diameter	
2Bk5	324-431	coatings	coarse angular blocky	5-10%
3Ck	431-491	minor	unconsolidated	
4Bkb	491-691	nodules	modular carbonate (3 cm mean)	

McFadden and Tinsley, 1985). Gile et al. (1966) described a four-stage evolution of pedogenic carbonate morphology (Stages I-IV). Bachman and Machette (1977; Table 1) described two additional stages (Stages V & VI).

Carbonate Morphology of the Soil

A soil profile was described on a natural exposure on the south side of Piute Gorge. Table 2 shows soil horizon designations and selected soil properties including carbonate morphology. A 58 cm thick laminar horizon occurs at a depth of 103 cm. Underlying the laminar cap are horizons with nodular calcium carbonate, ranging in size up to 7-8

Table 3. Stages and age ranges for Vidal Junction, CA/Overton, NV showing ages in millions of years (after Machette, 1985)

Stage	Age Range for Vidal Junction, CA/Overton, NV
I	Holocene to mid-Late Pleistocene (0-0.0)
II	mid-Late Pleistocene to Late Pleistocene (.05 - 0.12)
III	Late Pleistocene of mid-Middle Pleistocene (0.12 - 0.3)
IV	mid-Middle Pleistocene to late-Early Pleistocene (0.3-1.0)
V	late-Early Pleistocene to late Pliocene (1.0 - 2.0)
VI	late Pliocene to late Miocene (2.0 - 5.0?)

cm in diameter. This soil is overlain by material which appears to represent a later pulse of deposition.

Discussion

Although laminar layers have been described as having formed by other processes (Lattman, 1973), the nodular carbonate horizons that underlie the laminar horizon (2Bk3 and below) suggest that the carbonate is pedogenic in origin. Nodules that appear to engulf pedogenic structure suggest an upward migration of carbonate precipitation with decreasing permeability, again, consistent with a pedogenic origin for the carbonate.

The 58 cm thick laminar horizon at Piute Gorge represents carbonate morphology equivalent to Stage V. The suggested age for correlative Stage V calcic soils ranges from approximately 0.25 Ma for the Jornada I surface (Gile et al., 1996; 1981) to 2 Ma surfaces near Vidal Junction, CA and Overton, NV (Machette, 1985). A correlation in age to the Vidal Junction site suggests that the Piute Gorge surface may be as old as late Pliocene (Table 3). The Vidal Junction site is approximately 50 km to the southeast of Piute Gorge and has similar climatic conditions. However, it is unknown if other calcic soil forming factors such as dust flux, available water-holding capacity and partial pressure of CO₂ are similar between Piute Gorge and Vidal Junction. Thus, the correlation is tentative and subject to reinterpretation if future workers establish that the rate of formation of calcic soils varies between the two locations, or discover other evidence (i.e. paleontological).

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Summary of the Stratigraphy, Structure and Mineralization in the Eastern New York Mountains, San Bernardino County, California

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Introduction

The New York Mountains are one of the major mountain ranges of the eastern Mojave Desert (Figure 1), and have attracted prospectors and geologists for over 100 years. Several mines in the area have produced lead, silver, zinc, copper, gold, tungsten, and sericite mica. The eastern New York Mountains contains the largest known undeveloped high brightness, high purity limestone deposits in the southwestern United States.

Geology of the New York Mountains region was originally mapped by D.F. Hewett (1956) in the 1920's. Burchfiel and Davis (1977) recognized much of the stratigraphic succession, and many of the structural complexities in the area. Brown presented detailed stratigraphy and 1:6,000 scale geologic maps (1986), and the geology and genesis of the limestone deposits (1989). Beckerman and others (1982) reported on the intrusive rocks, and Miller and Wooden (1993) produced a reconnaissance map of the northern New York Mountains

and Castle Mountains area. This summary will focus primarily on the Paleozoic stratigraphy, Mesozoic structure, and mineralization in the eastern New York Mountains.

Regional Geologic Setting

Oldest rocks in the area are Precambrian metaigneous and metasedimentary gneisses which are considered part of the autochthonous basement terrane. More than 6000 feet of Paleozoic and Mesozoic sedimentary and volcanic rocks are present in the New York Mountains (Burchfiel and Davis 1977) (Figure 2). Although metamorphosed, the Paleozoic sedimentary rocks can be correlated with shallow water Cordilleran inner miogeoclinal and cratonal platform deposits (Figure 3) exposed in adjacent ranges (Brown 1986).

Intrusion of plutonic rocks in the area occurred during Late Cretaceous time (Beckerman 1982). Tertiary volcanic rocks (adjacent to the map area), old alluvium, stream terrace deposits, younger and recent alluvium were deposited during Cenozoic time (Hewett 1956).

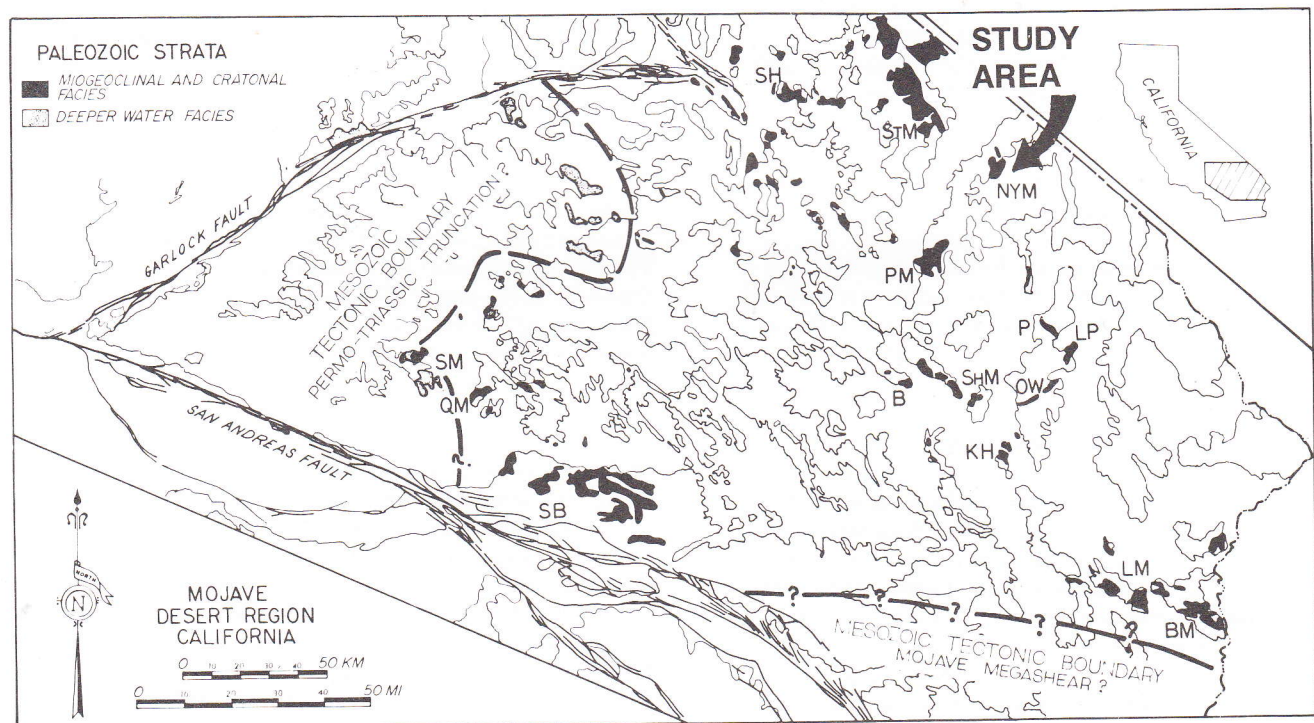


Figure 1. Index map of the Mojave Desert region showing location of the New York Mountains. Map also shows location of Paleozoic strata. SB=San Bernardino Mts., SM=Shadow Mts., QM=Quartzite Mtn., B=Bristol Mts., ShM=Ship Mts., P=Piute Mts., LP=Little Piute Mts., OW=Old Woman Mts., KH=Kilbeck Hills, LM=Little Maria Mts., BM=Big Maria Mts., PM=Providence Mts., NYM=New York Mts. (this paper), STM=Striped Mts., SH=Silurian Hills

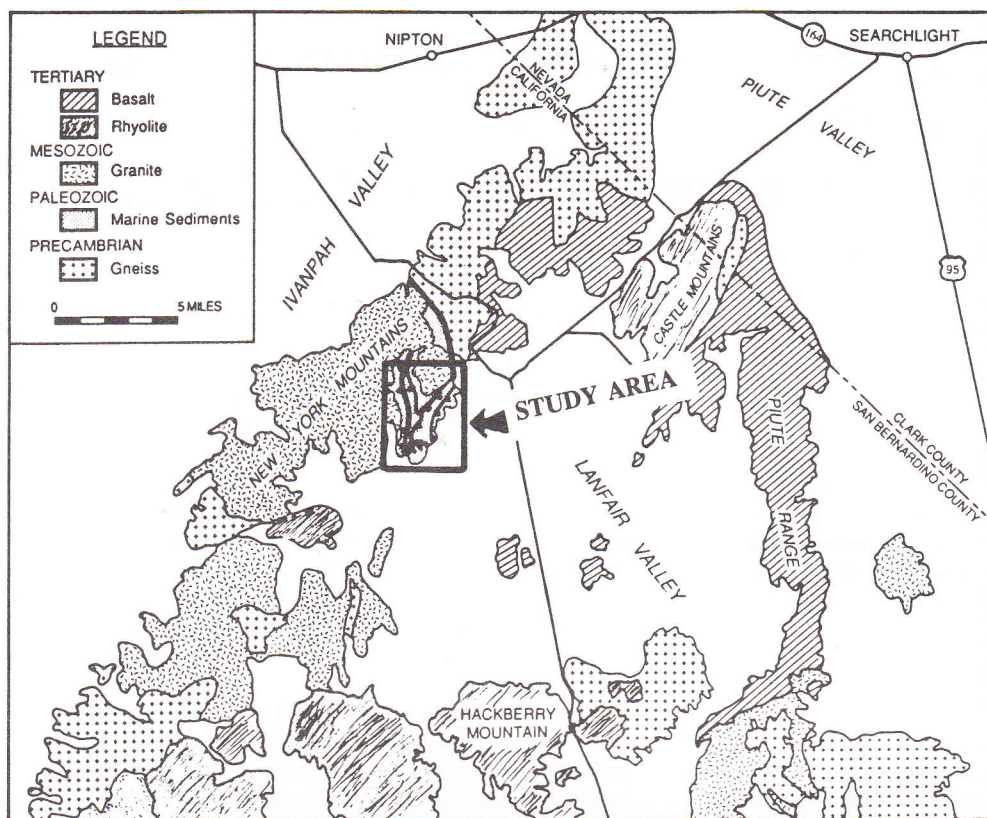


Figure 2. Generalized regional geologic map (modified after Linder 1989).

Structurally, rocks in the New York Mountains are exceedingly complex. Structures belonging to several deformational events have been recognized, and correlation of deformational events in the New York Mountains with those in adjacent ranges suggest that structures of both early and late Mesozoic age are present (Burchfiel and Davis 1977). Mesozoic structures in the New York Mountains are part of the foreland fold and thrust belt of the Cordilleran orogeny (Burchfiel and Davis (1977).

Tertiary volcanic rocks are cut by several sets of high angle faults. Several ages of alluvium and stream terraces, suggest several periods of uplift and erosion may have occurred during Cenozoic time (Brown 1989). No Quaternary faults are known (Miller and Wooden 1993).

Stratigraphy

Metamorphosed Paleozoic and Mesozoic sedimentary and volcanic rocks with a tectonic thickness of more than 6000 feet are present in the eastern New York Mountains. The sedimentary rocks have been correlated with formations in adjacent ranges, which are typical shallow water inner miogeoclinal and cratonal platform deposits (Burchfiel and Davis 1977, Brown 1986).

Paleozoic formations present in the New York Mountains include the Cambrian Tapeats Sandstone, Carrara Formation, Bonanza King Formation, and Nopah Formation; Devonian Sultan Limestone; the Mississippian Monte Cristo Limestone; and the Pennsylvanian-Permian Bird Spring Formation. Calc-silicate rock and quartzite inferred to unconformably overlie the Bird Spring Formation have been

correlated with the Moenkopi Formation of Triassic age (Burchfiel and Davis 1977, Brown 1986). Mesozoic metavolcanic rocks (Sidewinder Volcanics or Delfonte Volcanics equivalents) and sedimentary rocks unconformably overlie the Moenkopi Formation (Burchfiel and Davis 1977, Brown 1986).

Figure 4 is a composite stratigraphic column depicting the various Paleozoic formations and members which have been differentiated. Table 1. contains brief descriptions of the various Paleozoic stratigraphic units. Detailed mapping (Brown 1986) has allowed 30 or more map units to be differentiated, and it has been shown that the Paleozoic formations in the eastern New York Mountains are virtually identical to the inner miogeoclinal-cratonal platform Paleozoic rocks in

several other ranges within the Mojave region (Figure 3), (Brown 1986, Brown 1991).

Although the sequence is incomplete in the New York Mountains, the lower part of the Paleozoic section is comprised of brown, conglomeritic and pebbly, cross-bedded quartzite correlated with the Tapeats Sandstone, and interbedded calc-silicate and pelitic hornfels correlated with the Carrara Formation of Lower Cambrian age.

The middle part of the Paleozoic section is composed dominantly of a thick grey to white dolomite sequence. Visually and chemically distinctive thin banded dolomitic limestone of the Lower Member of the Bonanza King Formation of Lower Cambrian age, includes a brown hornfelsic marker bed at the top of the unit, which is also present in a number of mountain ranges in the Mojave (Brown 1982, 1986). Recognition of brown hornfels and impure silty limestone of the Dunderberg Shale Member of the Nopah Formation, has allowed several Cambrian dolomite dominated Formations and members to be differentiated (Table 1) (Brown 1986).

A regionally widespread unconformity is present between the Cambrian Nopah Formation and Devonian Sultan Limestone. Stromatoporoids in dolomite of the Lower Sultan Limestone may represent the Ironsides Member, although it has not been mapped separately (Brown 1986).

Thin bedded, white calcite marble of the Crystal Pass Member of the Sultan Limestone of Devonian age marks a change to a limestone dominated section. A complete section of Monte Cristo Limestone of Mississippian age is present, including the Yellowpine Member and the Arrowhead Member, a dark grey cherty limestone marker bed less than

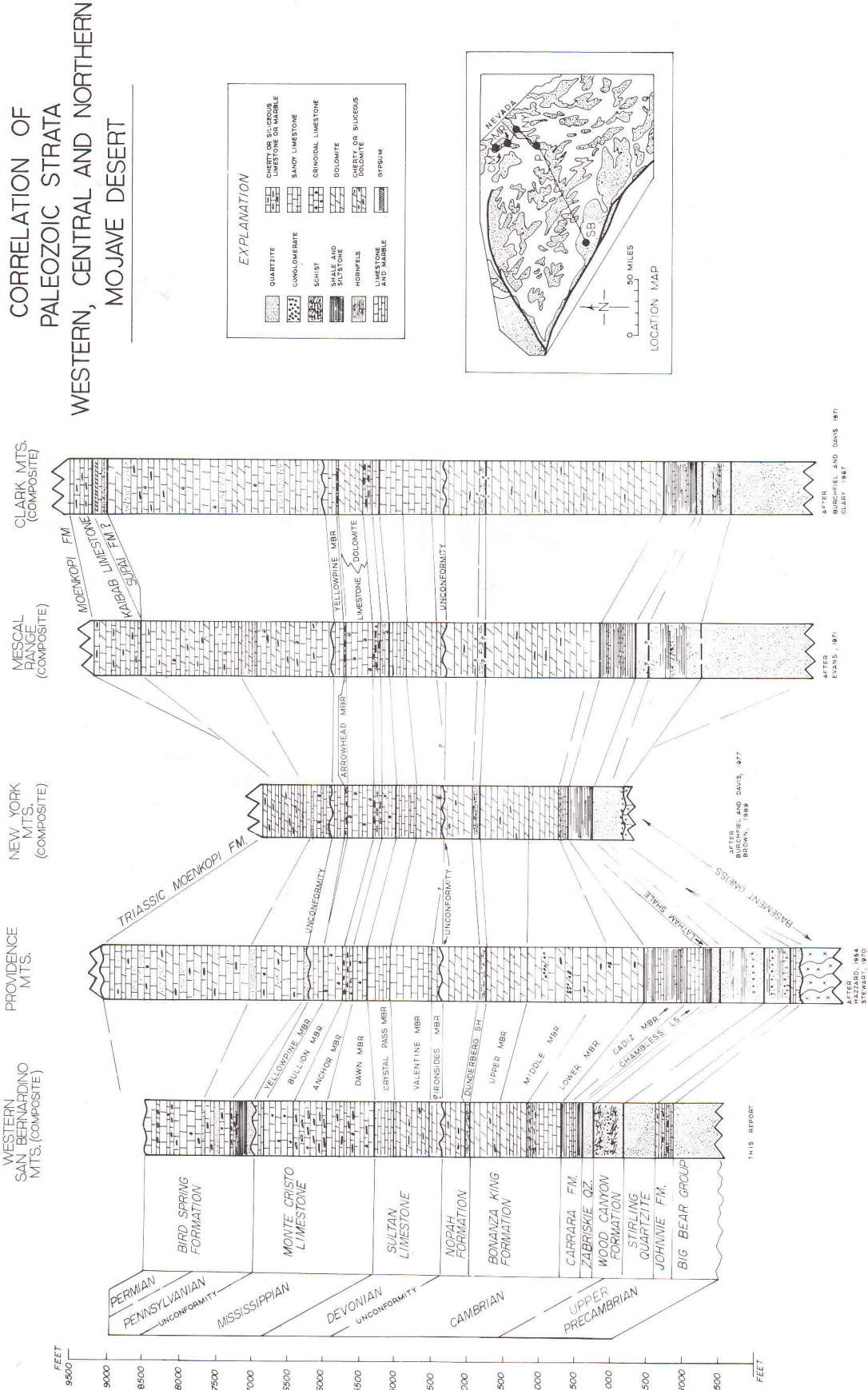


Figure 3. Correlation of Paleozoic strata from the San Bernardino Mountains to the northeastern Mojave Desert region (from Brown, 1989).

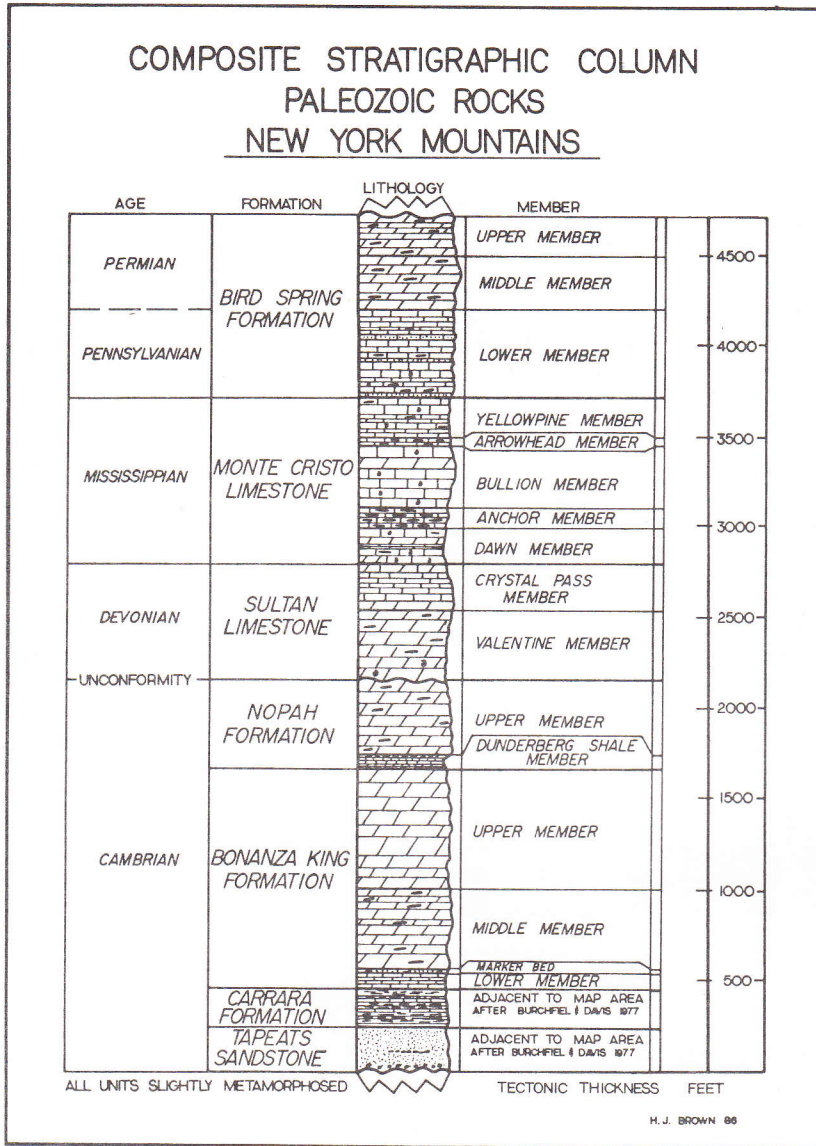


Figure 4. Composite stratigraphic column of Paleozoic rocks in the New York Mountains (from Brown 1989).

20 feet thick. The Bullion Member, a prominent thick bedded to massive, pure bioclastic limestone, is virtually identical to outcrops in the San Bernardino Mountains. Dark grey limestone which forms the Dawn Member, and the distinctive and very cherty Anchor Members, are carbon copies of exposures in adjacent ranges (Brown 1991). Brown colored secondary dolomitization of the Crystal Pass member of the Sultan and portions of the Monte Cristo Limestone is present in several exposures (Brown 1986). Similar dolomitization is prominent in these same units in the Clark Mountains to the north, and in the Goodsprings region near Las Vegas (Brown unpub.)

Rocks of the Bird Spring Formation overlie the Monte Cristo Limestone. Bird Spring rocks in the New York Mountains are similar to Bird Spring in other areas, but also have several differences. In many ranges an unconformity is present between the Monte Cristo and Bird Spring. This may

not be the case in the New York Mountains. The lower unit of the Bird Spring in the New York Mountains is composed of medium bedded white calcite marble with brown silty and sandy limestone beds, which are very similar to the lower unit of the Bird Spring in the San Bernardino Mountains. However most exposures in the San Bernardino Mountains, and many other ranges contain a basal clastic and carbonate unit (Indian Springs Member), which has not been recognized in the New York Mountains. The lower Bird Spring contact in the New York Mountains appears conformable, but a tectonite marble marker bed is present along some contacts, suggesting a tectonic contact, or that the unconformity is not everywhere present (Brown 1986).

The upper members of the Bird Spring in the New York Mountains are white and grey dolomite (Brown 1986). In many other ranges in the Mojave including the San Bernardino Mountains for example, the Bird Spring is largely limestone, with only a small dolomite component. Rocks assigned to the upper Bird Spring in the New York Mountains are very similar to portions of the Cambrian dolomite sequence. In the San Bernardino Mountains a major Mesozoic age thrust fault has juxtaposed the Bird Spring and Cambrian Bonanza King Formation in many places (Brown unpub.). It is possible that a (major) tectonic contact is also present in the New York Mountains, although it has not been mapped as such.

In conclusion, although the rocks are metamorphosed, and there are some minor differences and or complications, the Paleozoic section exposed in the New York Mountains is very similar to, and is correlated with virtually identical exposures of Cordilleran inner miogeoclinal and cratonal platform rocks exposed in several adjacent ranges in the Mojave Desert region.

Unconformably overlying the Bird Spring Formation is a sequence of calc-silicate,

hornfels, quartzite and impure carbonate rock correlated with the Moenkopi Formation of Triassic age (Burchfiel and Davis 1977, Brown 1986). Similar but unmetamorphosed rocks correlated with the Moenkopi Formation have been recognized in some adjacent ranges (Providence Mountains and Clark Mountains) (Hewett 1956, Burchfiel and Davis 1977).

Mesozoic Volcanic and Sedimentary Rocks

Mesozoic metavolcanic rocks are prominent in the Sagamore Canyon area. The rocks are fine grained, metamorphosed, and strongly foliated, and volcanic texture is generally only observed on weathered surfaces. The sequence includes interbeds of volcanic breccia or agglomerate, and thin metasedimentary siltstones which were probably water laid tuffs. Sericite muscovite schist may have formed from aluminous tuffs. The base of the

DESCRIPTION OF PALEOZOIC ROCKS NEW YORK MOUNTAINS

THICKNESS	DESCRIPTION	AGE
4500	<p>BIRD SPRING FORMATION</p> <p>UPPER MEMBER Thin to medium bedded, medium to dark grey siliceous and tremolitic dolomite</p> <p>MIDDLE MEMBER Thin bedded, white to buff weathering siliceous, tremolitic dolomite</p> <p>LOWER MEMBER Thin bedded to massive, light grey to white calcite marble with occasional siliceous or cherty layers, and dolomitized pods. Unit contains several brown weathering, sandy siliceous layers. Marker beds mapped separately include red and black siliceous marble, green tinted white marble. Where unmetamorphosed, the Lower Member is slightly impure medium olive grey limestone with occasional crinoidal debris.</p>	PENNSYLVANIAN-PERMIAN(?)
4000	<p>MONTE CRISTO LIMESTONE</p> <p>YELLOWPINE MEMBER Thin to medium bedded, light grey to white calcite marble, occasionally siliceous and dolomitized pods.</p> <p>ARROWHEAD LIMESTONE MEMBER Medium grey cherty limestone. Unit is less than 20 feet thick.</p> <p>BULLION MEMBER Thick bedded to massive, light grey to white pure limestone and calcite marble. Where unmetamorphosed contains common crinoidal debris. Brown weathering secondary dolomitized zone near top of unit some places.</p> <p>ANCHOR MEMBER Medium bedded, medium to dark grey limestone with abundant chert and siliceous layers. Occasional crinoidal debris.</p> <p>DAWN MEMBER Thin to thick bedded, light to dark grey limestone, often containing abundant crinoidal debris. Base of unit often irregularly dolomitized.</p>	MISSISSIPPIAN
3500	<p>SULTAN LIMESTONE</p> <p>CRYSTAL PASS MEMBER Thin to medium bedded, pure, white calcite marble. unit contains occasional oxidized pyrite, and common iron oxide stain. Several orange weathering zones and irregular dolomitization occurs near top and base of unit. Thin bedded grey marble marker bed near base of unit.</p> <p>VALENTINE MEMBER Medium bedded, light to dark grey siliceous dolomite. Occasional stromatoporoids near base of unit may mark the Ironsides Member, not mapped separately.</p>	DEVONIAN
3000	<p>UNCONFORMITY</p> <p>NOPAH FORMATION</p> <p>UPPER MEMBER Thin bedded to massive, interbedded blue grey, light grey and white fine grained dolomite, often siliceous and or tremolitic.</p> <p>DUNDERBERG SHALE MEMBER Thin layered, brown to grey impure silty limestone and dolomite, and interbedded brown weathering hornfels.</p>	UNCONFORMITY
2500	<p>BONANZA KING FORMATION</p> <p>UPPER MEMBER Medium bedded to massive, light grey to white dolomite, occasionally siliceous and or tremolitic.</p>	CAMBRIAN
2000	<p>MIDDLE MEMBER Medium to thin bedded dark grey slightly siliceous dolomite.</p>	CAMBRIAN
1500	<p>LOWER MEMBER Thin bedded, dark grey to slightly mottled dolomitic limestone. Brown silty hornfelsic marker bed at top of unit.</p>	CAMBRIAN
1000	<p>CARRARA FORMATION Interbedded calc-silicate and pelitic hornfels. Unit outcrops adjacent to map area.</p>	CAMBRIAN
500	<p>TAPEATS SANDSTONE Brown conglomeritic and pebbly, cross bedded quartzite. unit outcrops adjacent to map area.</p>	CAMBRIAN
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Table 1. Description of Paleozoic rocks in the New York Mountains.

metavolcanic unit is marked by a stretched pebble conglomerate containing carbonate clasts (Burchfiel and Davis 1977). The volcanic sequence has been difficult to correlate. The rocks are unlike the Cretaceous Delfonte Volcanics (100m.y.) which occur in adjacent ranges (Burchfiel and Davis 1977, Fleck et al. 1994). The rocks are superficially similar to the Jurassic Sidewinder volcanics which occur in the Victorville region to the west (Brown unpub.).

Unconformably overlying the metavolcanic rocks is a 200 foot thick sequence of generally dark grey to black argillite, slate, siltstone and conglomerates. These rocks only occur in the Sagamore Canyon area, and may represent a local site of deposition of volcanic derived sediments (Burchfiel and Davis 1977).

Intrusive Rocks

Intrusive rocks in the area are considered to be part of the Teutonia batholith, a large composite batholith which is exposed over 10,000 mi² in the region. Beckerman and others (1982) recognized and described several plutons including the Mid Hills Adamellite, and the Live Oak Granodiorite, which outcrop in the eastern New York Mountains area. The Mid Hills Adamellite has yielded K-Ar cooling ages of 83-73 m.y. for biotite, which is 10 to 25 m.y. younger than the emplacement age of 93 m.y. (Miller and Wooden 1993). The Live Oak Granodiorite has yielded biotite K-Ar ages of 79.9 m.y. (Beckerman and others 1982). Rocks of the two plutons are visually similar, and Miller and Wooden (1993) indicated that contacts between the Mid Hills Adamellite and the Live Oak Granodiorite are gradational.

The Paleozoic rocks and the intrusives are cut by several east-west trending granite porphyry, aplite and pegmatite dikes. Several of the dikes appear to blend into the plutons. A dike cutting and grading into the Live Oak Granodiorite has been dated by K-Ar method at 71.7 m.y. (Burchfiel and Davis 1977). Two Amazonite bearing east-west trending pegmatites are notable.

Metamorphism of the Paleozoic Sedimentary Rocks

Most of the Paleozoic sedimentary rocks in the eastern New York Mountains are metamorphosed. Paleozoic rocks have been metamorphosed for a distance of over two miles from the intrusive contacts. Most metamorphism is mild, lower greenschist facies. Dolomite is commonly tremolitic. The Bullion Member of the Monte Cristo Limestone becomes progressively recrystallized and bleached white, and fossils show progressive destruction as intrusive contacts are approached. Within 1000 feet of the contact the rock is recrystallized to medium grained calcite marble, with crystal size up to 1/4 inch. Miller and Wooden (1993) indicated that contact metamorphism occurred at about 95 m.y. In the southern part of the area gravity slides of non-metamorphosed Paleozoic rocks are present (Burchfiel and Davis 1977).

Structure

Structure in the area is exceedingly complex, the result of numerous deformational events. Because of the complexity, there are slight differences in interpretation among workers in the area. The sequence of events presented here is based

on my detailed mapping (Brown 1986, 1989). Several east directed Mesozoic age folding and thrusting events have been recognized, which predate intrusion of Late Cretaceous plutonic rocks. In addition rocks in the area have also been subjected to at least two periods of high angle faulting, and gravity slides are also present in the area. Tertiary volcanics are cut by two episodes of high angle faults, and several periods of uplift and erosion may have occurred since Tertiary time. Table 2 shows the geologic history of the area.

The earliest deformation recognized includes north trending east vergent recumbent folds. Numerous small scale folds are common in several upper Paleozoic stratigraphic units. The major recumbent (F_1) and associated small scale folds (F_2) are interpreted to have formed during the same deformational event. Bedding plane flowage and faulting (TF_1) occurred during the early folding event. Deformation predates Late Cretaceous intrusive activity, and post dates Jurassic(?) volcanic activity. Although there is little evidence, a Pre-Late Jurassic age is suggested.

Early folding was followed by F_3 folding, which formed north-south trending and gently south plunging, moderate to tight, upright to rarely overturned (east verging) folds with 1/2 wavelengths of up to 1000 feet. Many of the F_3 folds are on the inverted limbs of earlier recumbent folds and are therefore inverted folds. Although age constraints are loose, a Late Jurassic age is suggested.

After F_3 folding, the rocks were involved in thrusting (TF_2) and high level low angle faulting events. Mesozoic Moenkopi correlatives were thrust over and truncate older folds and thrusts. Age constraints are again loose, but a Late Jurassic age is suggested for TF_2 .

After TF_2 , a hiatus in compressional tectonics occurred, and the area was subjected to high angle faulting. Dip of the north striking faults is variable, but steep. An early Cretaceous age is suggested.

East-West trending high angle mineralized (Ag-Pb-Zn-W) quartz veins formed in fractures, fissures shear zones and faults which cut across the foliation in the metavolcanic rocks in the Sagamore Canyon area. These veins are truncated by younger thrusting (TF_3), and are cut by granite porphyry dikes related to younger Late Cretaceous plutonic activity.

Following high angle faulting, thrusting was again renewed (TF_3), and the previously folded and thrust (and overturned) Paleozoic and Mesozoic sediments were thrust eastward over the Mesozoic metavolcanic rocks. Faulting was complex and several compound thrust plates formed ($TF_{3,4,5}$). The structurally highest low angle faults (TF_3) involve highly brecciated rocks, and may have occurred at or near the surface. Some of the features could be back slides or gravity slides. Recent isotopic dating by Fleck et al. (1994) indicates Sevier age thrusting which may include $TF_{3(4,5)}$ post dates the late Early Cretaceous Delfonte Volcanics (100.5 m.y.), and likely occurred between 100 and 83 m.y. In the New York Mountains the post TF_3 Mid Hills Adamellite was emplaced at 93 m.y. (Miller and Wooden 1993) and thus TF_3 may have occurred between 100 and 93 m.y., and a mid-Cretaceous age is therefore suggested.

After TF_3 thrusting, was the formation of north trending open folds (F_4), and warps, which have folded the earlier thrusts, resulting in thrusts which dip both east and west.

Intrusion of plutonic rocks occurred during Late

AGE		SEDIMENTATION	IGNEOUS	METAMORPHISM AND MINERALIZATION	FOLDING	FAULTING AND UPLIFT
Cenozoic	Quat	Recent Alluvium Stream terraces				
	Tertiary	Old Alluvium	Volcanics			NW high angle NE high angle Gravity slides Uplift
Mesozoic	Cretaceous		Pegmatite dikes Plutonic rocks and dikes (71 m.y.)	North trending Mineralization (CaF ₂ -Cu-Ag) Metamorphism		N70E high angle ? Slaughterhouse Fault TF ₆
	Jurassic	Volcanic sediments Unconformity	(meta)volcanics	E-W mineralization Ag-Pb-Zn-W	F ₄ open folds	TF ₅ TF ₄ TF ₃ N-S High angle E-W High Angle TF ₂
	Tri	Moenkopi Fm.			F ₃ North trend upright to ot F ₁ , F ₂ Recumbent east verging	TF ₁
Paleozoic	Perm	Unconformity				
	Penn	Bird Spring Fm.				
	Miss	Monte Cristo LS.				
Paleozoic	Dev	Sultan Limestone				
		Unconformity				
Paleozoic	Cambrian	Nopah Formation Bonanza King Fm. Carrara Formation Tapeats Sandstone				
		Unconformity				
PreCamb		Basement Gneiss				

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Table 2. Chronology of geologic history, eastern New York Mountains.

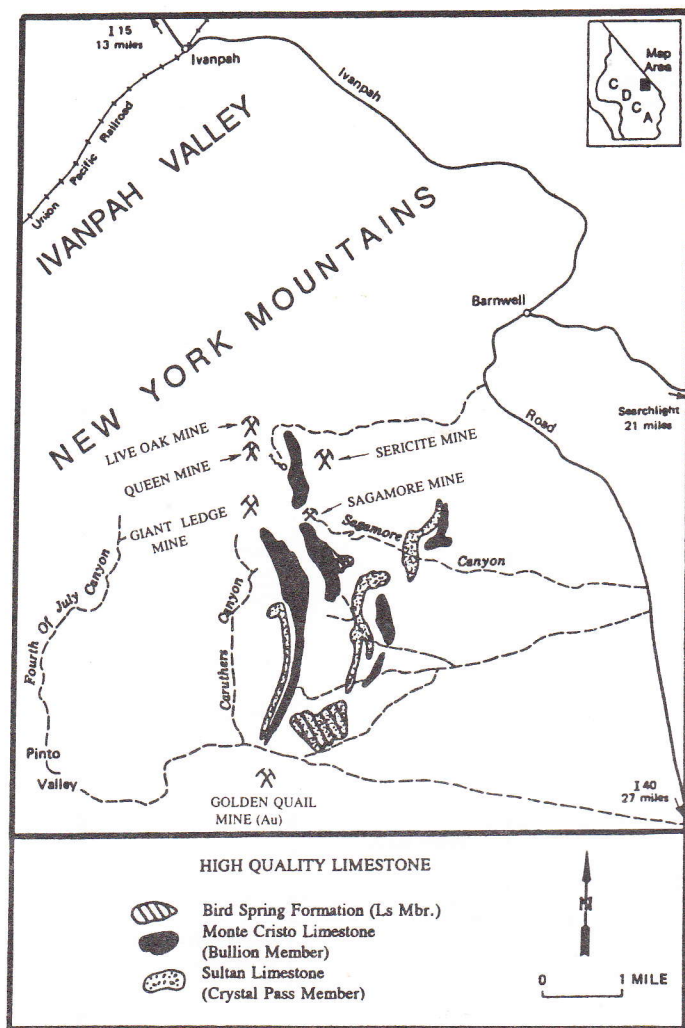


Figure 5. Eastern New York Mountains area showing major mineral deposits (from Brown, in press).

Cretaceous time (Burchfiel and Davis 1977) and post dates thrusting and folding events. Near some contacts the rocks are strongly recrystallized to marble. Skarn, north trending quartz veins, and some north trending contact mineralization (eg. Giant Ledge) formed at this time in the Paleozoic and intrusive rocks (Brown 1986), and is related to the Mid Hills Adamellite pluton. Intrusion of Amazonite bearing pegmatites may also have been related to late stage fluids of the Mid Hills Adamellite. Brecciated zones and small thrusts near some intrusive contacts may also have formed during intrusion of the pluton.

The Slaughterhouse Fault is located adjacent to the map area, and is a major high angle fault which has been traced for over 30 miles. The fault juxtaposes autochthonous Precambrian crystalline rocks and deformed and intruded Paleozoic and Mesozoic rocks. Burchfiel and Davis (1977) indicated that the fault cuts the Mid Hills Adamellite, but Miller and Wooden (1993) indicate that the fault is locally cut by the pluton. A Cretaceous age is suggested. Burchfiel and Davis (1977) speculated that the fault may represent a subduction related strike slip fault.

Following intrusive activity, the rocks were uplifted,

during which time several gravity slides occurred, particularly in the southern part of the area. Rocks involved in the sliding are not metamorphosed, and highly brecciated. Sliding may have followed rapid uplift, and may be a product of denudation tectonics. An early Cenozoic age is suggested.

In conclusion, complex structures belonging to several Mesozoic age deformational events have been recognized. Correlation of deformational events in the New York Mountains with those in adjacent ranges suggest that structures of both early and late Mesozoic age are present. Mesozoic structures in the New York Mountains are part of the foreland fold and thrust belt of the Cordilleran orogeny.

Metallic and Non-metallic Mineralization

The eastern New York Mountains are highly mineralized and several economically valuable metallic (Hewett 1956) and non-metallic mineral deposits are present in the area (Brown 1989). Gold, silver, lead, zinc, copper tungsten, and sericite have been mined in the past. Extensive high brightness limestone deposits are present but have not been mined (Figure. 5).

Metallic mineralization appears to have occurred during two or more distinct episodes. Older mineralization, the Sagamore Mine veins have yielded Pb-Zn-Ag-W (Wolframite). Sulfide mineralization occurs in several steeply dipping parallel quartz veins trending nearly east-west, hosted by Mesozoic volcanic rock. The veins appear to be truncated by and therefore predate TF_3 thrusting, and are also cut by the Late Cretaceous granitic dike rocks (Brown 1986).

Younger metallic mineralization which includes the Giant Ledge, Live Oak and Queen mines, has yielded copper, gold and fluorite. Mineralization trends north-south, and occurs along the contact of Late Cretaceous Mid Hills Adamellite and Upper Paleozoic carbonate rocks, or within the intrusive but adjacent to the carbonate contact. This episode of mineralization post dates TF_3 and post dates but closely followed and is likely related to late stage intrusion of the pluton (Brown 1986).

East-west trending granitic dikes, and east-west trending Amazonite bearing pegmatites may also be late stage offshoots of the Cretaceous pluton. Several hundred pounds of Amazonite crystals and lapidary quality material have been recovered from the Bar and Grill claim.

Several silicified and or jasperoidal Au(?) mineralized zones occur along the TF_3 thrust fault contact which has juxtaposed Paleozoic carbonate rocks over the Mesozoic volcanic sequence (Brown unpub.). Mineralization postdates thrusting but relationship to the Late Cretaceous intrusives is unknown.

Non-metallic minerals include sericite and high purity limestone deposits, and are of far greater importance than the metallic mineral resources. As much as 30,000 tons of sericite mica were mined over a period of several years from an open cut adjacent to Keystone Canyon. The deposit formed by metamorphism of aluminous ash flow tuff, which occurs in the core of a tight overturned fold, within the Mesozoic volcanic sequence.

The eastern New York Mountains contain the largest undeveloped deposits of high brightness, high purity

limestone in the southwestern United States. Major limestone deposits are present in the Sultan Limestone Crystal Pass member of Devonian age, Monte Cristo Limestone Bullion Member of Mississippian age, and Bird Spring Formation of Pennsylvanian age (Brown 1986).

Major factors influencing the genesis of the deposits include; depositional environment, post depositional activity, metamorphism, intrusion, faulting folding, uplift, erosion and preservation thru geologic time. Detailed geologic investigations have identified nine major deposits. The limestone is very pure containing >98% CaCO₃, and is also of high brightness (90->95 -270 mesh G.F.). The limestone is suitable for all currently produced high brightness, high purity limestone products. Combined indicated reserves of the nine limestone deposits are well over 100 million tons, with a gross value (existing ground limestone products, 1994 dollars) in excess of 5 billion dollars (Brown in press). Recent legislation involving the East Mojave area will have a negative effect on all future mineral exploration and or mining in the area.

Acknowledgments

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Lead-Rich Pegmatites in the New York Mountains

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The New York Mining District was organized in 1870 (Vredenburg, this volume). Ores of lead, silver, and zinc are discussed by Hewett (1956) and are inventoried by the California Division of Mines (Aubury, 1908; Cloudman and others, 1919; Eric, 1948; Jenkins, 1942; Tucker and Sampson, 1930; Wright and others, 1953). Brown (this volume) discusses the timing of intrusive relationships of deposits in this district. Many of the deposits are in Paleozoic clastic sediments and carbonate rocks which have been intruded by the plutons of the Teutonia Batholith complex.

The mining district is located in the New York Mountains, west of Barnwell and west of the trace of the north-south Slaughterhouse Fault (Miller and others, 1994), where there is a sequence of early and late Paleozoic clastic and carbonate sediments and Jurassic sandstone and volcanic rocks. The relationship of these rocks has been complicated by Mesozoic thrusting and folding, and by Late Jurassic and Cretaceous intrusion of the Teutonia Batholith. Granitic rocks of the Teutonia complex include the Mid Hills adamellite, a tan granitic rock with scant biotite that can be seen in Caruthers Canyon and Fourth of July Canyon. Live Oak Canyon contains exposures of the Live Oak granodiorite, a white rock with black euhedral crystals of biotite. The Mid Hills adamellite is dated at approximately 85 Ma, and the Live Oak granodiorite is approximately 80 Ma (Beckerman and others, 1983).

Mineralization occurred in two phases, one predating pluton intrusion and one coincident with or after the final stages of intrusion of the Mid Hills adamellite (Brown, this volume). Pegmatite dikes generally have granitic composition, that is, they consist primarily of quartz and feldspar, particularly microcline feldspar. Because they form from fluid-rich melt, crystals in pegmatites grow rapidly and often reach large sizes, while melt of similar composition with lower fluid composition will crystallize as fine-grained aplite.

In the New York Mountains, pegmatite dikes, aplite dikes, quartz dikes and monzonite porphyry dikes often trend roughly east-west. This may indicate that they were emplaced in a fracture system that was formed by north-south stresses. Because pegmatites are associated with pluton intrusion, the New York Mountains pegmatites may have formed around or after 85 Ma.

Pegmatites generally are enriched in incompatible elements that are not easily incorporated into common rock-forming minerals. Lead and bismuth are two such elements. The ionic radii of bismuth (Bi^{+3} 0.96, Bi^{+5} 0.74) and lead (Pb^{+2} 1.20, Pb^{+4} 0.84), for example, are greater

than the ionic radii of iron (Fe^{+2} 0.74, Fe^{+3} 0.64), and the former do not combine easily to form common minerals.

Two pegmatites in the New York Mountains have an unusual assemblage of minerals in part because of the presence of lead and bismuth. The assemblage of major, minor, and secondary minerals is shown in Table I.

Analysis

The pegmatite minerals and their oxidation products listed in Table I are unique in the Mojave Desert Province and several have only rare occurrences in the state of

Table 1. *New York Mountains pegmatite mineral assemblage*

* minerals inferred to be present on the basis of casts

Primary Pegmatite Minerals	
Lepidolite	$\text{K}(\text{Li}, \text{Al})_3(\text{Si}, \text{Al})_4\text{O}_{10}(\text{F}, \text{OH})_2$
Microcline, var amazonite	KAlSi_3O_8
Muscovite	$\text{KAl}_2(\text{Si}_3\text{Al})\text{O}_{10}(\text{OH}, \text{F})_2$
Quartz	SiO_2
Topaz	$\text{Al}_2\text{SiO}_4(\text{F}, \text{OH})_2$
Zinnwaldite?	$\text{K}(\text{Li}, \text{Fe}, \text{Al})(\text{AlSi}_3)\text{O}_{10}(\text{F}, \text{OH})_2$
Accessory Pegmatite Minerals	
Beryl	$\text{Be}_3\text{Al}_2\text{Si}_6\text{O}_{18}$
Cassiterite	SnO_2
* Cosalite	$\text{Pb}_2\text{Bi}_2\text{S}_5$
or	
* Galenobismutite	PbBi_2S_4
Manganocolumbite	$(\text{MnFe})(\text{Nb}, \text{Ta})_2\text{O}_6$
Monazite	$(\text{Ce}, \text{La}, \text{Nd}, \text{Th})\text{PO}_4$
Pyrite	FeS_2
Spessartine garnet	$\text{Mn}_3\text{Al}_2(\text{SiO}_4)_3$
Tourmaline, var. schorl	$\text{Na}(\text{Mg}, \text{Fe})_3\text{Al}_6(\text{BO}_3)_3(\text{Si}_6\text{O}_{18})(\text{OH})_4$
Zircon	ZrSiO_4
Secondary Minerals	
Anglesite	PbSO_3
Bismutite	$\text{Bi}_2(\text{CO}_3)_2\text{O}_2$
Coronadite	$\text{Pb}(\text{Mn}, \text{Mn})_8\text{O}_{16}$
Hollandite	$\text{Ba}(\text{Mn}, \text{Mn})_8\text{O}_{16}$
Vanadinite?	$\text{Pb}_5(\text{VO}_4)_3\text{Cl}$

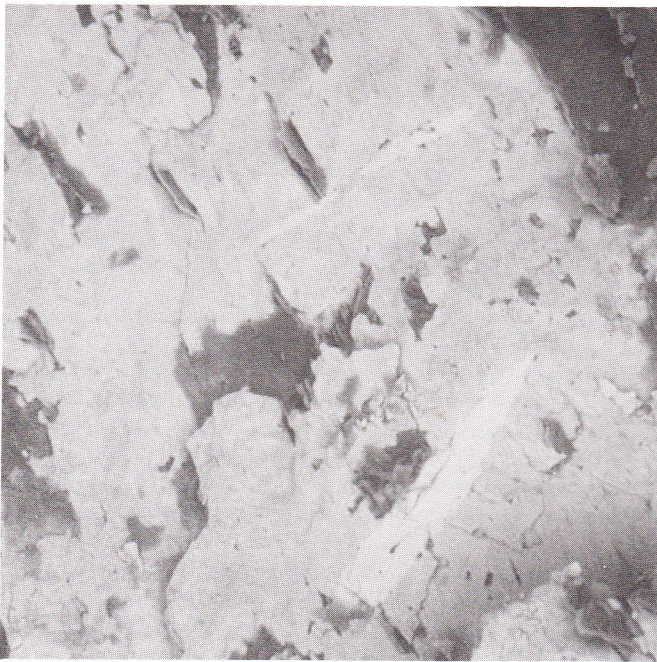


Figure 1. *Manganocolumbite*.

California. The topaz was studied by E. E. Foord, U.S. Geological Survey. The amazonite was analyzed by Hoffmeister and Rossman (1985) of the California Institute of Technology. One of us (RMH) examined a number of specimens from these pegmatites using an ETEC-Autoscan scanning electron microscope equipped with a PGT energy dispersive x-ray analyzer. This provided semi-quantitative compositional data on the remaining minerals for elements with atomic number greater than fluorine. Patricia Hunter



Figure 2. Blade-like cavities in quartz suggests *cosalite* or *galenobismutite*.

examined specimens for fluorescence. Many San Bernardino County Museum volunteers have assisted in specimen recovery over a period of several years.

Primary Pegmatite Minerals

Mica

These pegmatites contain abundant muscovite in large books. They also contain considerable purple, green, brown, and golden brown mica spanning a range of compositions. The common purple mica as well as some of the green mica are identified as lepidolite by their low aluminum to silicon ratios. The brown and golden micas are identified as zinnwaldite by having very low magnesium contents. The zinnwaldite contains manganese and iron in roughly equal amounts, reflecting the low percentage of iron in the deposit. Lepidolite and zinnwaldite have been reported from Riverside and San Diego counties (Jahns and Wright, 1951); the pegmatites in the New York Mountains may be the first record of the minerals in San Bernardino County (Pemberton, 1983; Murdoch and Webb, 1966).

Feldspar

Microcline, variety amazonite is a green feldspar classified as a semi-precious gemstone. The blue to green color is radiation-induced when the microcline contains lead. Amazonite with lead less than 1,000 parts per million (ppm) is blue; amazonite with a higher lead content is green (Hofmeister and Rossman, 1985). The analysis by these authors notes that the ratio of lead to iron is high in amazonite from the New York Mountains. The crystals in these pegmatites, generally frozen in the quartz core, are up to 10 inches in diameter. Some of the amazonite has white zonation (Foord and Martin, 1979) and rarely is intergrown with pink perthite. A single, one-inch, translucent, deep green crystal was found in an open pocket.

Amazonite in California is known from Lone Pine Station, Inyo County (Murdoch and Webb, 1966). Pale green microcline occurs at the Mesa Grande pegmatites and at Pala in San Diego County (Hofmeister and Rossman, 1985), and an angular blue detrital grain was found in Mojave River sediments near Oro Grande, San Bernardino County (San Bernardino County Museum collections). Amazonite from the Zapot mine, Mineral County, Nevada, is associated with zinnwaldite and topaz, but that deposit also contains abundant tourmaline, var. schorl, indicating that it was not a low iron deposit. A recent find of amazonite from the Rocket mine is an example of the excellent amazonite produced from Teller County, Colorado, where it is associated with fluorite and goethite (Geffner, 1995).

Quartz

Quartz has filled most pockets, creating a solid pegmatite core. A single pocket produced quartz crystals to six inches. These smokey crystals have suffered radiation damage and are dark gray to black, often with a milky exterior layer.

Topaz

Topaz is quite abundant in large, translucent, fractured masses and crystals near the core of both dikes. The equidimensional euhedral aspect of some crystals is attributed to twinning (E.E. Foord, U.S.G.S., Denver, pers. comm. to Reynolds, 1980). The crystals range in color from tan to milky to pale milky blue, and fluoresce a strong yellow color. In certain cases, what appears to be graphic granite is actually microcline with included, elongate topaz

prisms. The abundance of topaz indicates initial high concentrations of fluorine. However, there is only a modest amount of fluorite in the wall rock, suggesting that the impervious chilled margins of the dike formed rapidly and might closely reflect the bulk composition of the dikes. Crystallized topaz is known from pegmatites in San Diego County (Larson, 1977; Murdoch and Webb, 1966; Pemberton, 1983) and from the Zapot pegmatite, Mineral County, Nevada (Gordon, pers. comm. to author, 1995). The presence of topaz worldwide is discussed by Menzies (1995).

Accessory Pegmatite Minerals

Beryl

One blue-green crystal of beryl var. aquamarine was found frozen in quartz. The specimen was located in an isolated pocket down dip from the main pegmatite core. The frozen pocket, with beryl and microcline, did not contain topaz, and may represent a phase of crystallization slightly earlier than the larger core up-dip. Beryl is known from the eastern San Bernardino Mountains in San Bernardino County (SBCM collections) and is abundant in Riverside and San Diego counties (Pemberton, 1983).

Cassiterite

Submillimeter cassiterite is a fairly common accessory mineral, but is only found intergrown with manganocolumbite. A typical example is illustrated in Fig. 1. Similar textures have been reported from other localities by Ramdohr (1980). Cassiterite is also known from a contact metamorphic deposit in eastern San Bernardino County at the Evening Star mine in the Ivanpah Mountains (Tucker and Sampson, 1943; Wright and others, 1953).

Cosalite or Galenobismutite

Long, thin, striated, bladelike cavities in quartz, such as the ones illustrated in Fig. 2, are common and are invariably associated with secondary lead and bismuth minerals. This suggests that the primary lead and bismuth minerals may have been sulfosalts such as cosalite or galenobismutite. Galenobismutite occurs in sulfide rich hydrothermal veins and in pegmatites in the Darwin district, Inyo County (Pemberton, 1983). Cosalite has no record in California.

Manganocolumbite

Manganocolumbite forms elongated black, striated blades. They are frequently intergrown with cassiterite, as mentioned previously. This manganocolumbite contains variable amounts of iron, titanium, scandium, and tantalum, with scandium frequently making up a significant percentage of its composition. Manganocolumbite is common in Riverside and San Diego counties (Murdoch and Webb, 1966; Pemberton, 1983), but this occurrence is the first record for San Bernardino County.

Monazite

Monazite is present in crudely-formed, orange to orange-brown crystals. It is low in lanthanum and contains roughly equal amounts of cerium and neodymium, along with significant thorium. Monazite occurs at a number of localities in San Bernardino County, locally in the basnaesite deposits of Mountain Pass (Pemberton, 1983).

Pyrite

Submillimeter limonite pseudomorphs after a cubic mineral, as illustrated in Fig. 3, are present, and suggest that pyrite may have also been a minor primary mineral.

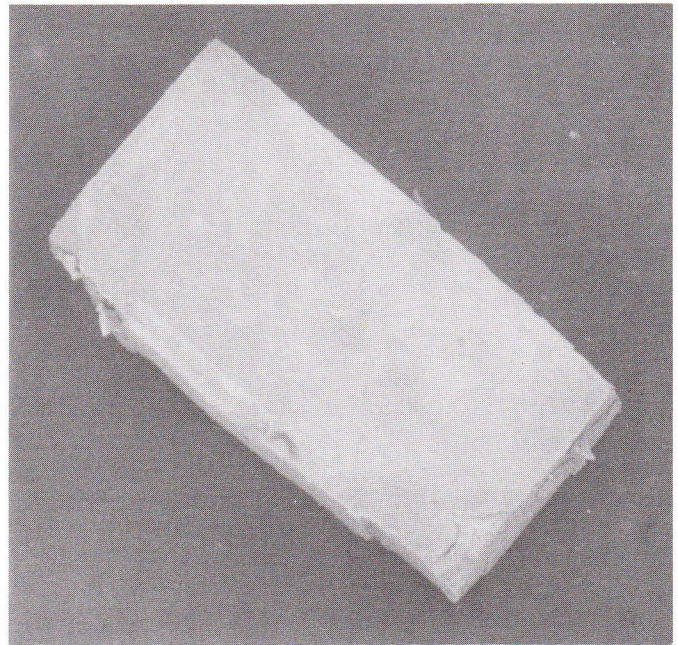


Figure 3. *Goethite pseudomorph.*

Spessartine Garnet

Spessartine is present as small, red-brown dodecahedral crystals. Spessartine garnets are found throughout California. At the Little 3 mine they occur with microcline and tourmaline (Larson, 1977).

Tourmaline var. schorl

Black, iron-rich tourmaline is almost absent from the pegmatites. Only one poorly-formed grain, about 1 cm long, has been confirmed by x-ray diffraction. The absence of tourmaline, particularly schorl, is consistent with a low iron component for the deposit. Schorl is known from



Figure 4. *Anglesite.*



Figure 5. *Bismutite*.

Wrightwood (Murdoch and Webb, 1966), the Shadow Mountains, and Coyote Lake areas of San Bernardino County (SBCM collections).

Zircon

Zircon occurs as brown, well-formed tetragonal, millimeter-size crystals and contains significant thorium. It is known from pegmatites in San Bernardino County at Marl Springs and in the Cady Mountains (Pemberton, 1983).



Figure 6. *Vanadinite*.

Secondary Minerals

Secondary minerals are oxidation products of the primary and accessory pegmatite minerals described above.

Clear *anglesite*, as illustrated in Fig. 4, is one of the crack-filling secondary lead minerals, frequently associated with vanadinite.

Bismutite occurs as white to orange, powdery-looking intergrowths of thin blades as illustrated in Fig. 5. Green bismutite is recorded from Morongo Valley in San Bernardino County, and it is widespread in pegmatites of San Diego County (Jahns and Wright, 1951; Pemberton, 1983)

Coronadite and *Hollandite*. Secondary manganese oxides are very abundant, and always contain significant lead or barium. The compositions generally correspond to coronadite or hollandite, although in some cases the lead content appears too high. Coronadite has been reported from Inyo County (Pemberton, 1983), but this is the first record of the species from San Bernardino County. Hollandite is a first record for California.

Vanadinite is common as a yellow, crack-filling secondary mineral, and is illustrated in Fig. 6. It sometimes is intergrown with anglesite and appears to have grown from it. Although not common in pegmatites, vanadinite is known widely in the eastern Mojave Desert (Murdoch and Webb, 1977; Pemberton, 1983).

Micrometer sized grains having compositions suggesting thorianite, cerianite, pyrochlore, etc. are found associated with monazite, zircon, and manganocolumbite, and in clay.

Discussion

It is interesting to speculate about the relationship between the low iron, lead-rich pegmatites and the hydrothermal mineralization that occurs elsewhere in the eastern New York mining district.

- Fluorine content is high in both. Fluorine alteration of the carbonate host rock has produced crystalline fluorite at the Hard Cash, Giant Ledge, and Bronze mines. At the Hard Cash and the Bronze, the fluorine alteration appears to take place at the same time as the deposition of primary azurite, malachite, and cerussite.
 - A secondary vanadate, mottramite, occurs at the Giant Ledge mine. It is a copper analog of the vanadinite found in the pegmatites.
 - Huebnerite is a low iron, high manganese tungstate that occurs at the Sagamore mine and the Hard Cash mine.
- Iron minerals, pyrite, and chalcopyrite, and the lead sulfide, galena, occur in abundance at the Hard Cash, Giant Ledge, and Sagamore mines. In contrast, the pegmatites are low in iron and high in lead. The pegmatite emplacement may be closely related to the younger phase of sulfide mineralization that took place along the north-south trending contact of the Mid Hills adamellite and the Paleozoic carbonate host rock. The low iron aspect of the pegmatite may indicate that the lead-silver-zinc and iron sulfides in hydrothermal veins were already deposited. Fluorine alteration of the sulfide deposits occurred at a later time, and might correspond to the emplacement of the fluorine and lead-rich pegmatites.

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Pack Mule Trails in the New York Mining District

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Introduction

When an unexpected, late-night storm near the Nevada border blanketed our sleeping bags with drifts of snow in the 1960s, a hasty, flashlight search of a topographic map for shelter led us to discover the abandoned but weatherproof cabins of the Sagamore Mine. The morning light introduced us to the glory of the New York Mountains (and our beagle to the excitement of tracking rabbits through the snow). Through the years, the New York Mountains have continued to be a locus for research and recreation. Our children, starting in infancy, learned responsibility and respect for their environment in the New Yorks. Jennifer learned word processing and data base design by reading manuals under the shade of piñon trees and pretending she had a computer. Bob perfected the art of roasting turkeys in a pit for Thanksgiving dinners in Caruthers Canyon. We have explored the mines and canyons with friends and colleagues, and we have met many people who have lived in and visited the region long before our "discovery." This paper is in part a result of observation and exploration over the past quarter century as well as research.

The New York Mountains contain many of the highest peaks in the Eastern Mojave Reserve. Steep-walled canyons, spectacular rock formations, piñon-juniper vegetation, and historic mines make the range an inviting area to explore on foot. Recreational hiking and climbing can be challenging. But it is surprisingly easy to travel on foot between historic mines on routes laid out and improved more than a century ago.

Background

The coming of the railroad to Lanfair Valley in 1893 changed the social dynamics in the New York Mining District. Miners in three different canyons had formed a self-reliant, but interdependent community that communicated and traded by a series of foot and pack mule trails. In 1893 the focus changed toward the new Nevada Southern Railroad and the food, supplies and technology that it carried.

Although the New York mining district was established in 1870 (Vredenburg and others, 1981; Vredenburg, this volume), the mining press (Crossman 1890-91) reported mining activity as early as 1861. The district was worked in 1873 and again in 1880. Eight silver mines, including the Keystone mine, were active, as were four copper mines. The mining methods and milling techniques were constantly being upgraded. Smelted bullion was shipped to San Bernardino and raw ore was taken by freight wagon a distance of 25 miles to Goffs on the A&P (later A.T. & S.F.) railroad.

In the 1890s, Isaac Blake bought up existing silver mines in Sagamore Canyon and named them the New York mine.

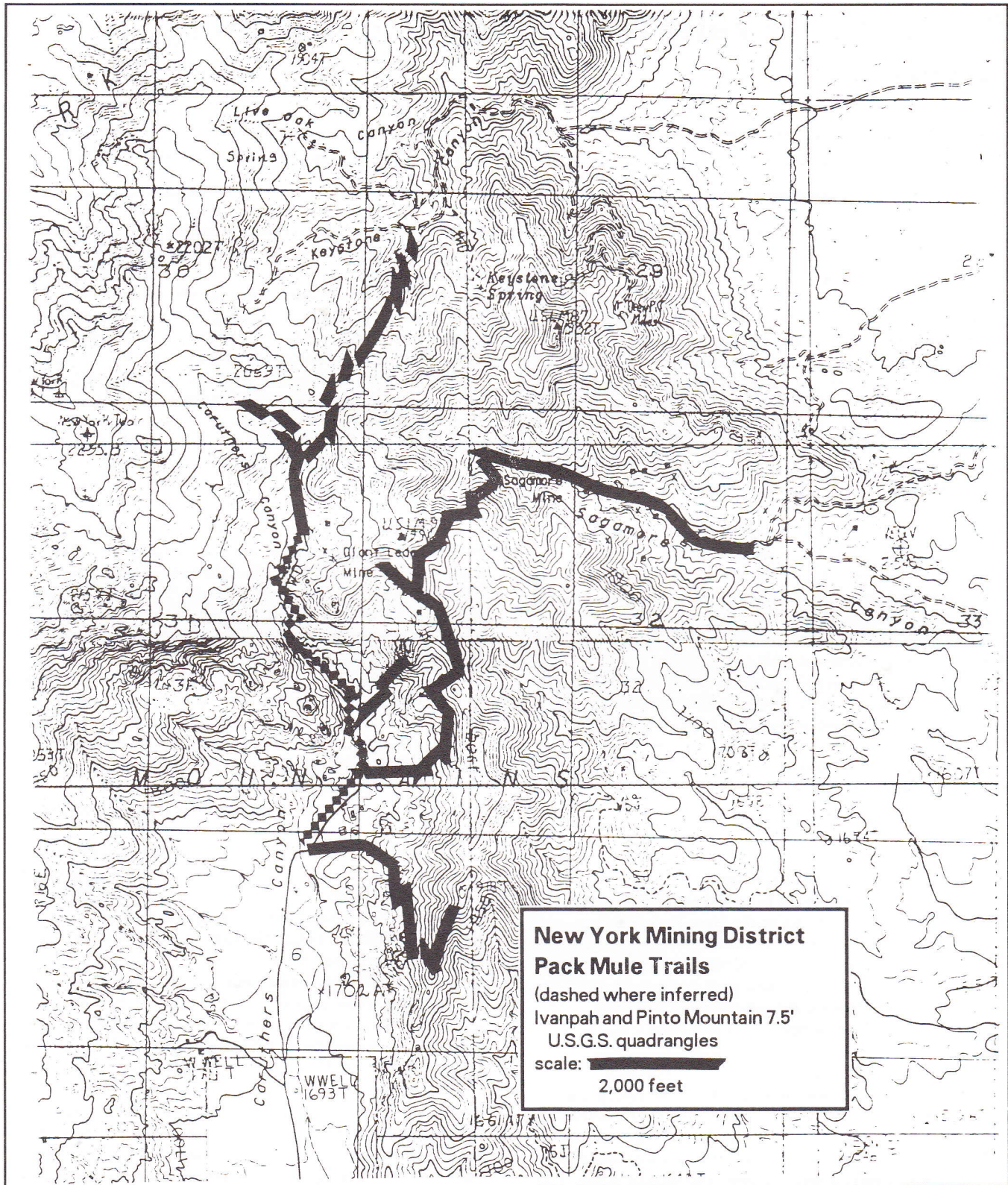
In 1893 Blake built the Nevada Southern Railroad north from Goffs to Manvel, named by Blake for the president of the Santa Fe, who in turned renamed Goffs "Blake." Blake intended to extend the line as far north as Utah, serving mines along the route and hauling ore to his reduction company in Needles (Myrick, 1963). Unfortunately for the Blake empire, silver prices fell and the New York mining district became idle (Vredenburg and others, 1981), although it has been suggested that Blake's purchase of the New York mine was less for its ore than to increase tonnage for his railroad (Myrick, 1963). Manvel remained the end of the line for the railroad as funding schemes failed and nearby bonanzas, such as Vanderbilt, floundered. Receivers took over the Nevada Southern in December, 1894, and much of Blake's property collapsed in debt immediately thereafter. A new company, the California Eastern Railway, took over the line in 1895 and at the turn of the century extended it to Leastalk (now Ivanpah). The Santa Fe loaned the funds for the expansion, and by 1901 owned the line outright—renaming Manvel "Barnwell," and changing "Blake" back to Goffs.

The New York mine was revived in 1907 when N.P. Sloan purchased it and formed the Sagamore Mining Company. In 1914, fifteen men were working at Sagamore, one of whom was Bert Sharp (Sharp, 1984). Bert and his wife, Maude, could walk four miles to their homestead north of the railroad siding of Maruba (Ledge) in about an hour. By 1917, the Sagamore mine was idle again (Vredenburg and others, 1981).

Mendenhall (1909) notes that there was no water along the Santa Fe between Vontrigger Springs and Barnwell. His 1908 map shows no roads to water sources in the New York Mountains. This probably reflects a period of time when mines were idle but before the enactment of the 1910 Homestead Act (Sharp, 1984).

Mines in the New York Mountains with names recognizable today appear in inventories by the California Division of Mines and the U.S. Geological Survey (Aubury, 1908; Bateman and Irwin, 1954; Cloudman and others, 1919; Eric, 1948; Hewett, 1956; Jenkins, 1942; Tucker and Sampson, 1930; Wright and others, 1953). They include the Bronze (Live Oak) mine in Live Oak Canyon, the Copper Queen mine in Keystone Canyon, the Sagamore (New York) mine in Sagamore Canyon, and in Caruthers Canyon the Hard Cash and Giant Ledge mines. The Gold Chief (Golden Quail) mine is at the mouth of Caruthers Canyon, south of New York Mountain Road.

The Ivanpah Quadrangle of 1912 (Marshall, 1912) shows roads in Fourth of July Canyon and Caruthers Canyon. They connected to the New York Mountain Road and ran to Ledge (later Maruba, and now the OX Ranch headquarters)(Sharp, 1984), probably a shipping siding for



the Giant Ledge mine. A road connected Sagamore Canyon and Keystone Canyon to Barnwell via Mail Spring and Lycre Well (Marshall, 1912). In 1912, no road was shown running from Sagamore to the railroad siding of Purdy, a stop on the old Nevada Southern named after a railroad man and

Blake's partner in the Nevada Reduction Company (Myrick, 1963). But this road was apparently available in 1914 when the Sharps walked to their homestead south of Purdy. A Sagamore-Purdy road is shown in the 1917 Thompson survey (Thompson, 1921).

Before the coming of the railroad in 1893, news and supplies came to the New York Mining District with freight wagons carrying 9000 pounds of supplies and equipment (Vredenburg and others, 1981). Between these occasions, miners probably visited and traded or borrowed supplies locally. When other mining camps were occupied and active, miners could socialize and trade. When camps were abandoned, they could recycle and reuse materials from them. The mines in Keystone, Sagamore, and Caruthers canyons are only about two miles apart: an easy two-hour hike.

The trails that were developed between mine camps of the New York Mining District crossed the rugged topography with gentle ascents and descents. Trails were often improved by cuts in ridges and through dikes and outcrops. Rock lagging and buttressing can still be seen where trails cross slopes with rivulets, streams, and gullies. Switchbacks were built to assist the gentle ascent to mines, saddles, and ridge tops. Trails start and end near water sources. A holding corral was built east of the Giant Ledge mine. These features indicate that the trails were specifically designed and constructed for pack mules hauling heavy supplies, lumber, and charcoal between mines. Whether the trails were developed by independent mule skinnners or as part of the community of mines in the New York mining district, they played an important part in the early activity of the district.

The Trails

In this paper, trails are described with informal place names used by campers, miners, and residents over the past quarter of a century. Mules were used in the mining district to haul sacks of ore from the mines (Frizzell, 1985) and to haul wood for mine timbers and charcoal to fuel coking operations which reduced lead ores. That the trails were used by pack mules is evidenced by panniers, a pair of "J" shaped hooks formed from one inch steel rod and mounted on the left and right side of a pack frame to support a balanced load of wood or charcoal. Panniers, some broken, have been found along the trails.

Caruthers - Keystone Trail

This trail starts at The Oaks in Caruthers Canyon. The Oaks, a former cabin site (structures still stood here in shambles in the late 1960s), is on the east side of the wash at the road crossing. This



Figure 1 Wheel barrow at the New York Peak Prospect. R.E. Reynolds photograph

is approximately the contact of bedrock with alluvial valley fill, and water often flows to this point. Historic wells are located here, and a dam was built in the middle 1900s. Forage is available on the flats. The Caruthers-Keystone Trail is partially obscured by the road which, prior to 1950, was pushed by bulldozer to the Hard Cash mine. The trail runs north, past the junction with the road to the May Barnes cabin. It clings to the west canyon wall until it crosses the wash to the east wall. A great deal of work went into buttressing the trail and subsequent roadway. The trail passes the Hard Cash mine and continues up canyon to the Upper Camp, where there was a cabin site and forge for smithing. Near the line between section 30 and 31, the trail



Figure 2. Anvil made of railroad track, wired to forked log and braced by tree and rock, at the New York Peak Prospect. R.E. Reynolds photograph.



Figure 3. Rock foundations near the Giant Ledge mine. R.E. Reynolds photograph.

branches northeast from the Blue Grotto Trail. The Caruthers-Keystone Trail switchbacks northeast to a saddle on Monument Ridge and proceeds north to the junction of ridges at the west end of Sagamore Canyon. It then follows a north-northeast trending ridge with gentle slope, crosses the 29/30 section line, and reaches the Keystone Road one canyon west of Keystone Spring. This trail is about 1.9 miles long and gains 860 feet in elevation.

Blue Grotto Trail

This short trail leaves the Caruthers-Keystone Trail near the 30/31 section line and proceeds west-northwest along the north slope of Caruthers Canyon. There is buttressing to cross rivulets, and switchbacks help gain 250 feet in altitude. The trail disappears after 1/4 mile at copper prospects and a spring.

New York Peak Prospect

The New York Peak prospect is on the north slope of Caruthers Canyon below a pinnacle that is 1/4 mile east of the highest point in the New York Mountains (VABM 7532, New York 2). We tried to locate a supply trail to the prospect but found none. We were at first very impressed that the prospectors had hauled a 150-pound, hand-forged steel wheelbarrow to the prospect, 1800 feet higher in elevation than The Oaks. We then realized that it had been assembled on site from three pieces. Some giant of a person had hauled a thirty inch long section of Nevada Southern railroad track to this site to use as an anvil! The track suggests the prospect dates later than 1893.

Giant Ledge Trail

The Giant Ledge Trail leaves the Caruthers-Keystone Trail at the junction to May Barnes cabin. After 1/4 mile, it passes the cabin, crosses the wash, and switchbacks to the Consolidated Tunnel, a gain of 300 feet in one-half mile. The Consolidated Tunnel was an exploratory effort to reach the riches of the Giant Ledge vein. It was pushed at least 1/4 mile through granite and ended when only low-grade ore was found (Frizzell, 1985).

Windy Point Trail

This trail leaves The Oaks and runs east-northeast up a

low ridge. In 1/4 mile, the topography steepens and the trail turns south for 2,000 feet. It then switchbacks to the ridge top and turns north to the Windy Point mines on Monument Ridge. Approximately one mile long, this trail gains 500 feet in elevation.

Sagamore Trail

The pack mule trail from The Oaks serviced the original Giant Ledge¹ tunnels and connected areas of wood cutting and charcoal making. The Sagamore Trail leaves the Caruthers-Keystone Trail 1/4 mile north of the Oaks, where an intermittent stream runs west to the upper dam in Caruthers Canyon. On the low slopes along the first quarter mile, the trail passes through a juniper cutting area. Stumps, stacked wood and charcoal indicate that cord wood and charcoal were hauled away for fuel and reduction ovens. The trail passes adits and prospects of the Athens, Morningstar, and Miama claims on the west slope of Monument Ridge. It passes rock foundations and turns northeast, then runs north through a saddle. A short branch runs northwest to the Giant Ledge mine. The main trail continues north along the east crest of Monument Ridge. One quarter mile north of the Giant Ledge and about one quarter mile south of USLM 92, the



Figure 4. Snow defines the Sagamore Trail. R.E. Reynolds photograph.

trail comes to a corral: two fenced pens, 8x10' and 10x10', are anchored to piñon trees. Milled lumber and wire nails suggest that the corral was repaired and still in use after the turn of the century (Sears Roebuck catalog, 1909). Two tent cabin foundations are nearby. This site is located at elevation 6340', the maximum elevation of the trail, 720 feet above The Oaks and 800 feet above the main Sagamore mine. The Sagamore Trail proceeds northeast on the east side of Monument Ridge and east of USLM 92, reaching the ridge on the south side of Sagamore Canyon. It switchbacks northeast down a ridge to the canyon bottom. This broad portion of the west end of Sagamore Canyon is a wood lot with stacked logs to 18" diameter and 3 to 3.5 feet long. Piles of charcoal remain in flat areas. The trail then runs east down the well-watered canyon for 2/3 mile to the main Sagamore workings. The Sagamore Trail is about 2.1 miles long.

Summary

The years between 1860 and 1893 were years of discovery and development in the New York mining district. Trails were developed for pack mules to haul food, equipment, and wood to the mines. The mules hauled sacks of concentrated ore and charcoal to the smelters. In fact, if distance mattered, communication via two-mile long mule trails would have been easier between Caruthers, Keystone, and Sagamore canyons than it would have been to Manvel or Ledge after the coming of the Nevada Southern.

The Nevada Southern Railroad, constructed in 1893, changed the focus of communication and trade. Supplies could be obtained at the rail sidings of Ledge and Purdy. Homesteaders entered Lanfair Valley and social gatherings became frequent after 1910. The Lanfair "boom" was short-lived (Casebier, 1987) and the Santa Fe abandoned its California Eastern branch to Goffs in 1923 (Myrick, 1963). But the mines remained, and interest in their riches never really died. May Barnes, who lived in the cabin at the foot of the Giant Ledge trail and today lives in Needles, describes their mining venture for us.



Figure 6. Corral on the Sagamore Trail. R.E. Reynolds photograph.



Figure 5. Stumps left in the wood cutting area along the Sagamore Trail. R. E. Reynolds photograph.

We were living at Colton. In 1951, Bob Engels came along and got Sam Barnes all excited about some patented mining interests in the New York Mountains about 25 miles northwest of Goffs, California. Sam was always interested in mining so he bought an interest in the property. They got Jerry Anderson, who ran a grocery store in Colton. We moved to the Mountains and a mining company was formed. The Whites, Andersons and several others joined in with Bob Engels and Sam.

The first thing they did was to repair the road that led up to what was called the Hard Cash mine and the property was patented. We were about 25 miles from Goffs where we got our mail. The Bozarth Ranch covered all the territory for miles for grazing. A cabin was built for us to live in. We lived in a very nice campground 'til our cabin was finished. The men all continued to work on the road, and gather samples. Seemed like it was a very complex ore and a lot of lime. The mine had been worked years before. There was a track coming out of a long tunnel and an ore car still there. We gathered up

some real good samples.

A Mr. Miller, who had worked for mining companies, came by and they hired him to dig out the spring which was a short distance from our cabin. They ran a long pipe from a spring to the road where they could get our water tank filled and moved near the cabin for household uses. We raised rabbits and had tomato vines. There was another mine across the mountains called Copper Queen which had some action. Finally the Giant Ledge Lead and Copper Company ran out of money. We were left up there and we ran lines and staked claims over that mountainside. Jerry Anderson bought up some of the mining interest and we moved into Needles in October, 1952. As of my knowledge Gerry Anderson purchased all the shares.

There was another ranch in a canyon beyond ours called Caruthers.

At one time the Santa Fe sent several men out there for several days to take samples. They said they had come up with some good samples but we didn't hear any more that I know of.

Acknowledgements

The authors thank May Barnes of Needles for sharing her adventures in the New York Mountains for this paper, and for the personal insights that she and her daughter, Helen Lozano, have provided about Caruthers Canyon. Larry Vredenburg kindly reviewed a draft of this paper and we appreciate his comments and suggestions.

Notes

1. U.S. Geological Survey 7.5 minute and 15 minute topographic maps label the Hard Cash mine as the Giant Ledge.

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Cenozoic Geologic Framework and Evidence for Late Cenozoic Uplift of the Castle Mountains, Castle Peaks, and Piute Range, California

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Abstract

The Castle Peaks, Castle Mountains, and Piute Range form the north and east boundaries of Lanfair Valley in southeastern California. The mountain ranges largely consist of Miocene volcanic rocks, and sedimentary deposits of Miocene to Quaternary age cover much of the valley. Most of the Miocene volcanic rocks erupted locally and were tilted during an episode of extensional (detachment) faulting. Uplift of the Castle Peaks, Castle Mountains, and Piute Range, relative to Lanfair Valley, postdated Miocene volcanism and detachment faulting, as indicated by the following lines of evidence. 1) Common clasts of Paleozoic limestone and Mesozoic granite of the Teutonia batholith in Miocene gravel deposits at high elevations in the Castle Peaks, Castle Mountains, and Piute Range were derived from sources in and near the Mescal Range to the northwest, and the southeastern New York Mountains to the southwest. Drainages from those sources presently are either obstructed by topographic barriers or flow in a direction that is inappropriate for transportation of such clast types to the deposit sites. The earliest time that the topographic barriers could have formed is late Miocene. 2) Two deep canyons in the Piute Range probably represent superposition of east-trending late Miocene drainages onto Miocene lavas and gravel deposits; thus the canyons were cut after the cessation of volcanism in late Miocene time. 3) Blockage of local drainages and accumulation of thick playa deposits at the west boundary of the Piute Range also occurred after the cessation of volcanism in the late Miocene and before the late Pleistocene, possibly in Pliocene time. The topographic barriers were created by local uplift of the northern New York Mountains and the high western ridge of the Castle Mountains, relative to Lanfair Valley. This uplift probably resulted from late Cenozoic faulting events that also caused segmentation and reorientation of drainages regionally.

Setting

The Castle Peaks, Castle Mountains, and Piute Range form prominent ridges and buttes 30 to 50 km southwest of Searchlight, Nevada (figs. 1 and 2) in northeastern San Bernardino County, California. The Castle Peaks are at the northeastern end of the northeast-trending New York Mountains and the Castle Mountains lie between the Castle Peaks and north-trending Piute Range (fig. 1); the south part of the Piute Range adjoins Homer Mountain (Spencer and Turner, 1985), near Civil-War era Fort Piute. The area lies west of the Colorado River Valley, and straddles the boundary between the northern and southern parts of the Basin and Range province. The mountain ranges border Lanfair Valley on the north and east and mostly lie within boundaries of the Mojave National Preserve, although parts of the Castle Mountains and Piute Range extend into Nevada. Seven topographic quadrangles at 1:24,000 scale cover the area: Castle Peaks, East of Grotto Hills, Hart Peak, Homer Mtn., and Signal Hill, California, and Tenmile Well and West of Juniper Mine, California and Nevada.

Geology and Structure

Pre-Cenozoic Rocks

The oldest rocks of the Castle Peaks, Castle Mountains, and Piute Range are Early

Proterozoic gneiss and foliated granite that record episodes of tectonism, metamorphism, and plutonism beginning about 2300 million years ago and continuing to about 1640 million years ago (Wooden and Miller, 1990). Rocks of

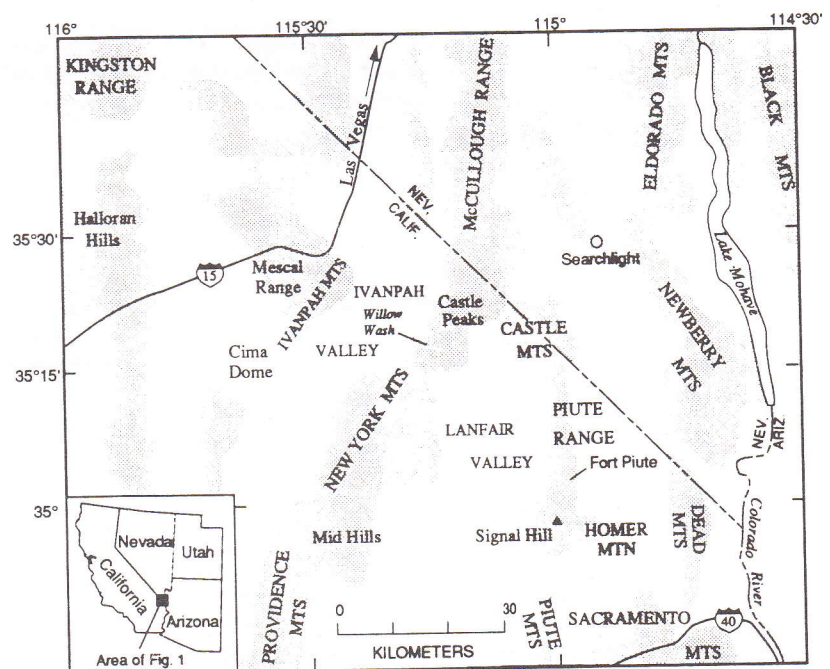


Figure 1. Map showing the location of the Castle Mountains, Castle Peaks, and Piute Range, California and Nevada.

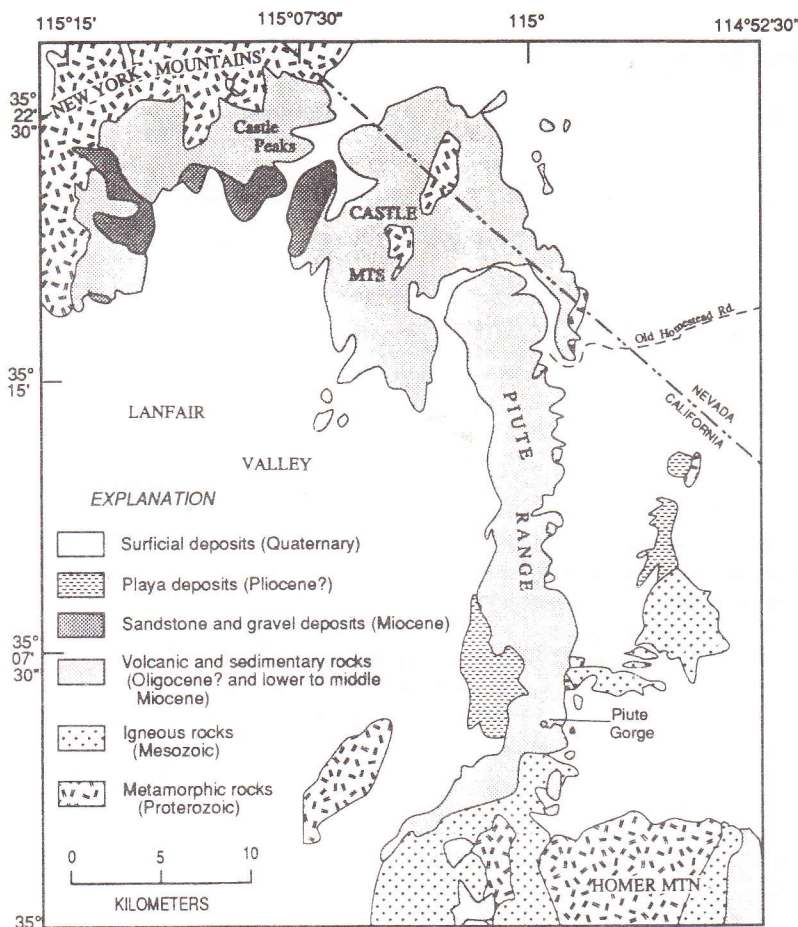


Figure 2. Generalized geology of the Castle Peaks, Castle Mountains, and Piute Range.

Paleozoic age are rare in the immediate area, although metamorphosed and complexly faulted and folded Paleozoic rocks crops out in the southeastern New York Mountains (Burchfiel and Davis, 1977; Miller and Wooden, 1993), where they are invaded by Mesozoic (Cretaceous) granite plutons (Beckerman and others, 1982; Miller and others, 1986). Other exposures of Paleozoic rocks are found in low-relief parts of Lanfair Valley. The Proterozoic to Mesozoic rocks are overlain nonconformably by Cenozoic (Tertiary and Quaternary) sedimentary and volcanic deposits.

Cenozoic Rocks

Miocene Volcanic and Sedimentary Rocks

The lowermost Tertiary (Oligocene? and Miocene) unit in the Castle Peaks, Castle Mountains, and Piute Range is locally-derived arkosic sandstone and conglomerate overlain by the Peach Springs Tuff of Young and Brennan (1974), a regionally-widespread sanidine-rich, sphen-bearing ash-flow tuff of alkali rhyolite composition (Buesch, 1993) deposited 18.5 million years ago (Nielson and others, 1990). The basal units are overlain by sequences consisting mostly of volcanic rocks that erupted locally from dikes and plugs and accumulated in local fault-bounded basins; these volcanic sequences form the skyline and dominate outcrops in each of the three mountain ranges.

The locally-erupted volcanic rocks have ages that indicate eruptions from about 19 million years ago to about 8 million years ago (Turner and Glazner, 1990; Nielson and Nakata, 1993). Each of the locally-erupted sequences is distinctive: the Castle Peaks sequence comprises piles of coarse volcanic breccia (Miller and others, 1986; Thompson, 1990; Nielson and others, 1993), the Castle Mountains sequence is dominated by light-colored rhyolite flows, tuff, and intrusive domes (Turner, 1985; Turner and Glazner, 1990; Capps and Moore, 1991; Nielson and others, 1993), and the Piute Range sequence is composed mostly of dark flows and breccia intruded by myriad dikes and sills, and rare domes (Nielson and others, 1987; Nielson and Nakata, 1993). The contrasting volcanic sequences of the Castle Mountains and Piute Range formed coevally for the most part, in close proximity, on irregular topography. Miocene sedimentary deposits overlie the Castle Peaks and Castle Mountains volcanic sequences on an angular unconformity (Balkwill, 1967; Miller and others, 1986), and also interfinger with or fill channels between or eroded into the uppermost Piute Range volcanic flows (Nielson and Nakata, 1993; Nielson and others, 1993). The sedimentary deposits comprise arkosic sandstone with lenses of coarse gravel, coarsening upsection to predominantly gravel with pebble and cobble sized clasts and very little matrix. In the western Castle Mountains basalt lava flows and a silicified rhyolite ash-flow(?) tuff of unknown ages are interbedded in the basal part of the gravel deposit. The unit can be

traced westward into the Castle Peaks, where the gravel fills a broad paleovalley (Miller and others, 1986; Miller and Wooden, 1993), called the Willow Wash paleovalley (Miller, 1995a, this volume). In the Castle Mountains the Miocene gravel deposits overlie rocks that are 14 to 12.9 million years old (Turner and Glazner, 1990) and in the Piute Range, the uppermost lava flows interbedded with the gravel deposits are 8 to 10 million years old.

Clasts in the gravel deposits are mostly Proterozoic gneiss and granite, Paleozoic to Mesozoic limestone and marble, and Mesozoic granite. A smaller proportion of clasts consist of Mesozoic metasedimentary and metavolcanic rocks or Tertiary volcanic rocks, although concentrations of andesite porphyry from the Castle Peaks volcanic sequence are found locally. Clasts of unmetamorphosed limestone and Mesozoic metavolcanic rocks probably were derived from the vicinity of the Mescal Range and Clark Mountain (D.M. Miller, 1995a, this volume); also present are granite clasts derived from parts of the Teutonia batholith that crop out in the southeastern New York Mountains (D.M. Miller, 1995a, this volume).

Late Tertiary and Quaternary Deposits

On both west and east sides of the Piute Range, substantial thicknesses of thin-bedded and fine-grained claystone were deposited in playa and (or) lacustrine

environments. A sequence of playa deposits more than 80 meters thick accumulated against faulted Piute Range lavas at the west end of Piute Gorge (fig. 2). These rocks are soft, horizontally-bedded, buff, tan, and reddish-brown claystone, siltstone, sandstone, and pebbly sandstone with gypsum and calcite interbeds, which also contain dispersed pebbles of basaltic scoria, massive basalt, and andesite derived from the Piute Range volcanic sequence. The unit is deeply dissected by the locally-developed drainage of Piute Gorge into badlands topography, and is capped by alluvium bearing a soil horizon with a thick petrocalcic layer. The thickness of the petrocalcic layer suggests that the soil developed between early and late Pleistocene (Katzenstein and others, this volume); thus the playa deposits began to accumulate in late Tertiary, probably Pliocene, time.

Another sequence of uniformly light-colored, massive claystone and siltstone crops out against a buttress of Cretaceous granite in Piute Valley, east of the Piute Range (figs. 1 and 2). The exposed thickness of these playa deposits is about 10 meters but the actual unit thickness is unknown. The strata include zones of siliceous tufa, deposited by hot-springs. Some outcrops of tufa contain fossil bison, yielding a Rancholabrean land mammal age that indicates formation in the late Pleistocene.

Everywhere in the Castle Mountains, Castle Peaks, and Piute Range and intervening valleys, unsorted Pleistocene and Holocene alluvial fan and stream channel deposits overlie the Tertiary and older rocks. Pleistocene alluvial fan and stream deposits are highly dissected to form relatively steep sided ridges 5 to 6 meters above the level of the active washes. The gently-sloping crests of the Pleistocene ridges have been stripped of soil, exposing calcified zones. Younger Pleistocene and Holocene stream deposits form broad, gently-sloping terraces 2 to 4 meters above active washes; these younger deposits overlap and in places partly bury the dissected ridges of older Pleistocene alluvium. Both the older and younger alluvial deposits consist of predominantly clay and quartz-sand matrix, with about equal proportions of clasts derived from Tertiary and pre-Tertiary rocks. Locally the younger alluvium may be largely volcanic-clast boulder conglomerate that also contains rocks derived from reworking of older alluvium.

Structural Relations

Structure of the Older Miocene Rocks

The Miocene volcanic rocks dip moderately to gently (45° to 25°), but as much as 65° locally, to both southeast and southwest. The overlying Miocene sandstone and gravel deposits dip gently to very gently (20° or less); in the Castle Peaks and Castle Mountains they overlie the tilted volcanic sequences on a distinct angular unconformity. Although at least one low-angle normal fault has been described in the northern Castle Mountains (Turner and Glazner, 1990), the Miocene volcanic sequences are mostly cut by high-angle normal faults that are oriented northeast and east-northeast (Nielson and others, 1993), although some Tertiary faults are oriented nearly due east. In addition some Miocene faults strike northwest, parallel to Mesozoic faults (Miller and others, 1986).

The steeply-dipping normal faults locally exhibit dip-slip offsets that are on the order of tens of meters; however, basal Tertiary rocks in the Castle Mountains crop out at elevations

300 to 500 meters higher than stratigraphically and structurally equivalent rocks on the east side of the Piute Range (Nielson and Nakata, 1993). This amount of offset in the basal Tertiary contact on older rocks probably is produced by a fault or zone of normal faults with strikes parallel to the straight west boundary of the Piute Range, where steep gravity and magnetic gradients are observed (U.S. Geological Survey, 1983; Mariano and others, 1986). These gravity and magnetic gradients are probably due to the contrast of Miocene volcanic and older rocks across steep faults.

Younger Miocene and Quaternary Structure and Paleotopography

Two deep east-trending canyons are cut into Miocene volcanic rocks and Miocene gravel at the south and north ends of the Piute Range. The southern canyon is Piute Gorge; the unnamed northern canyon lies at the end of Old Homestead Road (fig. 2). Playa deposits at the west end of Piute Gorge were deposited against the steep western scarp of the Piute Range, a boundary that clearly was created by steep faults. The contact between soft playa sediments and the faulted lavas is obscured by talus, and therefore provides no information about possible movement after formation of the playa deposits. The buttress contact between stratigraphically-equivalent playa strata and granite in Piute Valley to the east is also at a fault zone with east-side-up movement. Hot-springs that deposited tufa at the playa margin likely were channeled along that fault (R.E. Reynolds, oral commun., 1994), although there is no evidence of post-playa movement on the fault. No obvious fault offsets are observed in Quaternary alluvial deposits, but equivalent ridges of older alluvium are exposed at high elevations at the north end of Lanfair Valley and at lower elevations farther to the south.

Short east-striking faults are present in Miocene lavas of the Piute Range, but only faults with north-south orientations show control of drainage orientations. The canyon at Old Homestead Road shows fault control only at the east end, where it trends south, parallel to a zone of north-striking faults. The drainage of Piute Gorge is not connected to that of Lanfair Valley, but arises in Piute Range lavas directly north of the playa deposit exposures where it probably follows a north-south fault. Within Piute Gorge, the drainage follows another north-south fault for a short distance, but overall maintains its eastern course. The playa deposits at the western end of Piute Gorge are dissected by the drainage, indicating that entrenchment of the canyon continued after formation both of those deposits and the pedogenic carbonate in the overlying alluvial cap.

Discussion

Miocene Extensional Volcanism and Tectonism

In all the mountain ranges, rocks that show the greatest amount of tilting were erupted between 18.5 and about 14 million years ago, coincident with at least one event of extension that caused extreme tilting of Miocene volcanic rocks in the nearby Black Mountains of Arizona, and in the Eldorado Mountains and McCullough Range of Nevada (13 to 16 million years ago; Faulds and others, 1994). No low-angle normal (detachment) faults due to extension crop out in the region, but detachment faults are exposed to the west

in the Kingston Range (fig. 1; Reynolds, 1993) and to the east in the Black Mountains (Faulds and others, 1990; Faulds, 1993). Most likely, extensional faulting produced the volcanism and structural trends of Miocene rocks, and of many of the numerous faults that cut the Miocene rocks.

The thick sequences of locally-erupted lava, fragmental ejecta, and sedimentary rocks that form the Castle Mountains and Piute Range are elongate in a north-south direction and have limited lateral overlap relations, even though they accumulated nearly contemporaneously. For example, exposures 500-to 600-meter thick of Piute Range rocks that erupted between about 19 and 8 million years ago crop out over an east-west distance of less than 2 km, but extend for about 60 km from north to south (fig. 1). In a similar time frame (18.5 to 13 million years ago), andesitic to light-colored rhyolitic rocks and minor basalt flows were erupted in the Castle Mountains to the west. The limited extent of both the Castle Mountains and Piute Range volcanic sequences, and limited overlap of these adjacent contemporaneous sequences indicate that they accumulated in side-by-side fault-bounded troughs (Nielson and Nakata, 1993).

The relation between the Castle Peaks and Castle Mountains volcanic sequences is less clear because no overlapping or interfingering relations are exposed. A tuff unit in the upper part of the Castle Peaks sequence (Gusa and others, 1987) may be the lateral equivalent of the Wild Horse Mesa Tuff of McCurry (1988), which forms thick deposits at the west side of Lanfair Valley. If proved, this correlation would indicate that the Castle Peaks volcanic sequence is coeval with units that are more than 16 million years old, in the lower part of the Castle Mountains volcanic sequence (Nielson and others, 1993).

Evidence of late Cenozoic Faulting and Uplift

The distinctive association of far-traveled clasts in Miocene gravel deposits of the Castle Mountains, Castle Peaks, and Piute Range, superimposed drainages of the Piute Range, and ponding of late Tertiary playa deposits indicate that the Miocene paleotopography and drainage systems were quite different from those of the present time. Pre-existing topography is indicated by distinctive clasts derived from sources in the Mescal Range, Clark Mountain, and the southeastern New York Mountains, in deposits of Miocene gravel elevations above 1300 meters in the southern Castle Peaks and western Castle Mountains (fig. 2). Identical gravel is exposed in the central Castle Mountains and the northern Piute Range, at elevations between 1000 and 1100 meters. Late Miocene drainages that transported clasts to these sites were channeled eastward toward the Willow Wash paleovalley (Miller, 1995a, this volume).

At present, sediment derived from the Mescal Range and Clark Mountain cannot reach the Willow Wash paleovalley because of topographic barriers formed by Ivanpah Valley and the New York Mountains. Also, modern drainages that originate in the area of the Willow Wash paleovalley no longer transport debris into the central part of the Castle Mountains or to the Piute Range, because they are deflected to the south by a topographically high ridge of middle Miocene rhyolite domes and volcanic ejecta (average elevation, 1550 meters) that forms the western flank of the Castle Mountains. Similarly, erosional debris from the New York Mountains must be transported northward to the

Willow Wash paleovalley, whereas the modern drainages of the southeastern New York Mountains trend southeast. Miller (1995b, this volume) cites thickness trends in the Wild Horse Mesa Tuff to suggest that the depositional surface of western Lanfair Valley did slope northward in the early Miocene.

The evidence cited above indicates that late Miocene drainages in the area of Lanfair Valley were integrated with those of Ivanpah Valley to the west, and flowed generally eastward. Also, western Lanfair valley apparently drained northward. With respect to the Willow Wash paleovalley, the location of Tertiary gravel deposits in the Castle Mountains and Piute Range indicates a Miocene drainage of eastern Lanfair Valley that drained eastward. Additional evidence for east-trending drainages is supplied by the two generally east-trending Miocene drainages that were incised into Miocene lavas and gravel of the Piute Range. In spite of prominent north-striking faults in the lavas, the eastward trend of both drainages persisted as the canyons formed, which may indicate rapid downcutting.

In contrast to the apparent integration of Miocene drainages between Ivanpah and Lanfair Valleys, those of the present day are segmented by mountain ranges that form topographic barriers. Instead of a northward drainage system, most of Lanfair Valley now drains southeast. Formation of topographic barriers to the southeastward transport of clasts from the Mescal Range-Clark Mountain area to the Willow Wash paleovalley, as well as obstruction of drainage eastward from the Willow Wash paleovalley, occurred after the cessation of volcanism in late Miocene time. The deep canyons in Piute Range lavas and interbedded gravel were cut more recently than about 8 million years ago, in response to either uplift of the Piute Range or lowering of regional base level. The ponding of drainages and accumulation of playa deposits at both the western boundary of Piute Range lavas, and in Piute Valley to the east, probably represents uplift of the Piute Range relative to eastern Lanfair Valley. The time of playa formation could be as young as Pliocene.

The formation of topographic barriers to drainages that once connected the areas of the Ivanpah and Lanfair Valleys must have been caused by local uplift of the mountain ridges relative to the valleys. Reorganization of the topography and drainages of a region including Lanfair Valley, the Castle Mountains, Castle Peaks, and Piute Range, thus was most probably due to regional tectonism, and accomplished by local faulting rather than remote changes of base level. These faulting events could not have occurred earlier than late Miocene, and might have begun as late as Pliocene time. Some Piute Range faults with due north strikes, parallel to the western boundary of the range, may be related to the late Cenozoic uplift, which also reoriented drainage patterns. Variable elevations of older Quaternary alluvium at various sites around Lanfair Valley may provide evidence that such relative fault movements continued into Quaternary time.

Conclusions

Evidence from clast contents, formation of playas, and incision of east-trending canyons into upper Tertiary units, indicate that the Castle Mountains and Piute Range were uplifted in events of faulting that occurred in late Cenozoic time (Pliocene or latest Miocene), after volcanism related to

Miocene detachment faulting had ended. Topographic relations suggest that the late Miocene drainage patterns in the area of the Ivanpah and Lanfair Valleys were more regionally integrated compared to the present, and drained generally eastward. Segmentation of the regional drainage resulted from regional-scale events of local faulting and probably corresponded to tectonic uplift of the Castle Mountains, Castle Peaks, and Piute Range relative to the flanking valleys. Relative uplift of the Piute Range in late Cenozoic time ponded drainages at the east side of Lanfair Valley, and superimposed east-trending drainages across the Piute Range. These tectonic events cannot be older than late Miocene but could have begun in the Pliocene, and may have continued into middle Pleistocene time.

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Tectonic Implications of a Middle Miocene Paleovalley, Northeastern New York Mountains, California

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Abstract

The middle Miocene Willow Wash paleovalley incised 180 m into volcanic rocks south of the Castle Peaks, northeastern New York Mountains, contains deposits of eastward-flowing streams. Many clasts are derived from two widely separated areas, the Mescal Range to the northwest and the Live Oak Canyon area of the New York Mountains to the southwest. The paleovalley cuts volcanic rocks about 17.5 to 19 Ma and probably predates the inception of a 13-Ma basin to the northwest. The detrital provenance implies topography sloped southeastward from the Mescal Range during the middle Miocene, across what is presently Ivanpah Valley. Deposition may have begun when source areas in the New York Mountains and Mescal Range were uplifted, probably by reverse faulting along the Slaughterhouse and Kokoweef faults. Ivanpah Valley formed by normal and strike-slip faulting after 13 Ma.

Introduction

Hewett (1956) conducted the initial mapping of the Ivanpah Valley area (Fig. 1). He noted that Miocene volcanic rocks in this region lap against presumed highlands of older rocks and mapped the limit of this onlap. His studies have endured as the definitive regional statement on Miocene paleotopography of the northeastern Mojave Desert, although many advances have been made in understanding deposits in specific areas and in dating volcanic rocks and sedimentary deposits. Miller and others (1986) and Miller and Wooden (1993) noted the presence of a paleovalley in the northeastern New York Mountains, between Barnwell and the Castle Peaks (Fig. 2). The paleovalley is exposed at the head of Willow Wash, for which it is named in this paper. Miller and Wooden (1993) described the deposits of the paleovalley as Miocene gravel and sand deposited in a fluvial environment.

This paper presents information on the age of gravel in the Willow Wash paleovalley, sources for its clasts, and discusses the implications for tectonics and landscape evolution. My sedimentology studies were reconnaissance in nature. Subsequent study documenting stratigraphic variations in detrital composition may reveal additional details about drainage patterns and tectonic processes.

Geologic Setting

The northeastern New York Mountains are underlain by early Proterozoic granite and gneiss that are unconformably overlain by Miocene volcanic rocks. The Miocene volcanic stack consists of the Peach Springs Tuff (of Young and Brennan, 1974) at the base, a medial sequence of dacite breccia, and a capping sequence of andesite and basalt lavas (Miller and others, 1986; Nielson and others, 1993). The Peach Springs Tuff is a widely-traveled ignimbrite (Glazner

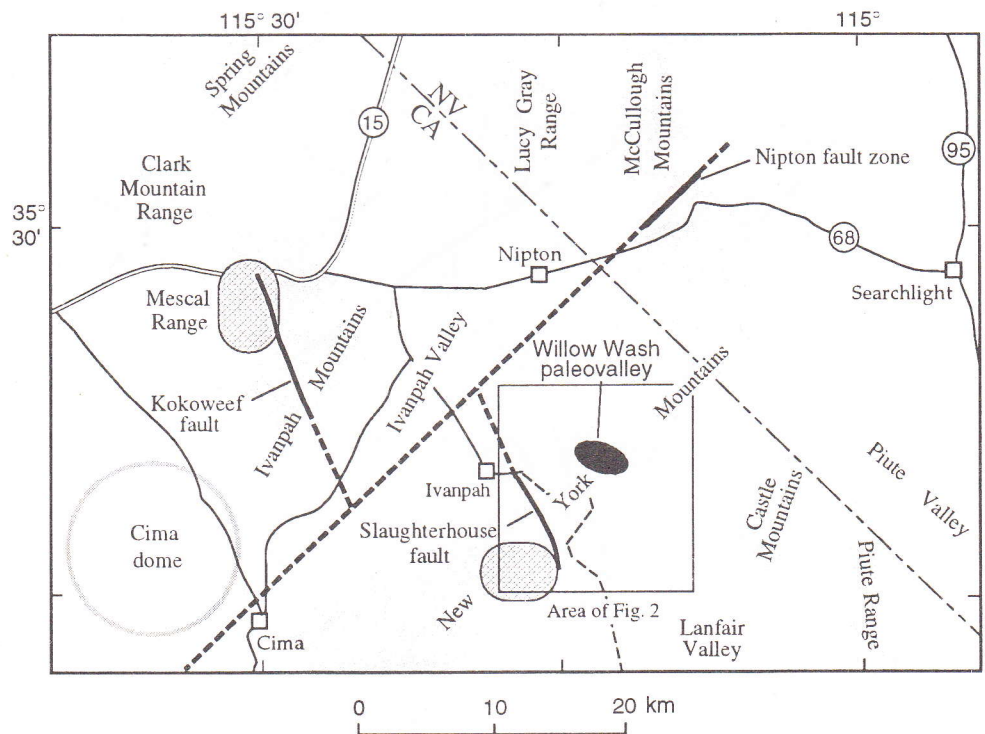


Figure 1. Location of Willow Wash paleovalley in northern New York Mountains, northeastern Mojave Desert, California. Nipton fault zone accomplished ~15 km of left-lateral separation, probably in the Late Miocene (Swanson and others, 1980; Miller and Wooden, 1993; Miller and Jachens, 1995). Ovals with cross-hatch represent locations of probable source terranes for gravel clasts in paleovalley.

and others, 1986) that dates the base of the volcanic sequence at 18.5 ± 0.5 Ma (Nielsen and others, 1990). The top of the sequence is probably slightly older than 17 Ma, on the basis of heavy minerals extracted from a tuff in the andesite and basalt sequence (Gusa and others, 1990), which suggests a correlation to the Wild Horse Mesa Tuff (17.8 Ma; McCurry, this volume); Miller and Wooden (1993) initially inferred an age younger than 16 Ma on the basis of a previously reported age for Wildhorse Mesa Tuff of 15.8 Ma (McCurry, 1988). A pre-18.5-Ma paleosurface at the base of the Miocene section and a post-17.5 Ma paleosurface at the top of the section are described below.

Paleosurfaces and gravels

The paleosurface beneath the volcanic rocks mainly coincides with the base of the Peach Springs Tuff, which crops out in a restricted, northeast-trending belt from the mouth of Live Oak Canyon through Barnwell to Dove Spring (Fig. 2). Elsewhere younger volcanic rocks form the base of the section, suggesting that Peach Springs was confined to a northeast-trending broad canyon bottom. This supposition is heightened in a few places by the presence of fluvial(?) arkosic sand beneath the Peach Springs. The ~200 m thick volcanic section between Barnwell and the Castle Peaks is much thicker than sections to the south and north,

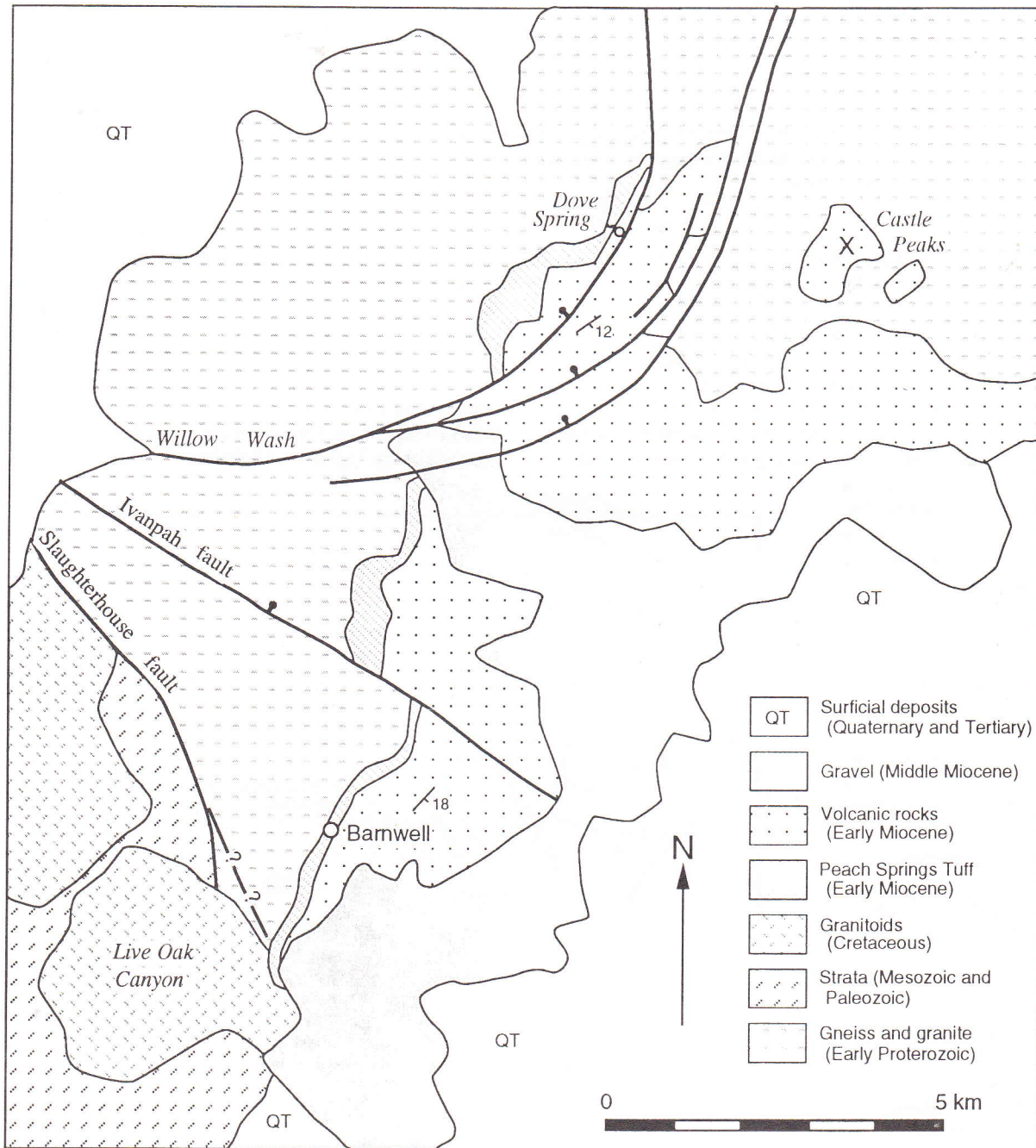


Figure 2. Simplified geologic map of northern New York Mountains showing Miocene gravel deposits of Willow Wash paleovalley (after Miller and Wooden, 1993). Gravel (gray unit) is tilted about the same as underlying volcanic rocks. Faults indicated by heavy lines; bar and ball on downthrown sides. Axis of paleovalley is represented by a narrow reentrant that projects northwestward into the center of the range.

suggesting that a topographic depression existed. However, it is also possible that the thicker section represents a volcanic build-up near a center of eruptions.

More information can be interpreted from the paleosurface that overlies the volcanic rocks. In many places, the paleosurface is approximately parallel to underlying volcanic flows. However, near the head of Willow Wash, the surface cuts down section as a paleovalley cut into Proterozoic rocks. Within and adjacent to the paleovalley, coarse gravel lies on the paleosurface.

Willow Wash Paleovalley

Where it is exposed at the crest of the New York Mountains, the Willow Wash paleovalley consists of a U-shaped channel cutting the roughly planar top on the volcanic section. The paleovalley is 1.2 kilometers wide, about 180 meters deep, and at least 2 km long, and is filled with crudely to well bedded gravel, sand, and silt. Cross-beds indicate fluvial transport to the east-southeast. Clasts in these deposits are very instructive about source terranes of the valley's tributary system. There are three principal sets of clasts: 1) locally derived angular clasts of Early Proterozoic gneiss and granitoids and Miocene volcanic rocks that are similar to the immediately adjacent basement rocks (Miller and Wooden, 1993); 2) rounded clasts of Late Proterozoic, Paleozoic, and Mesozoic rocks that range from unmetamorphosed to highly metamorphosed and foliated; and 3) rounded clasts of granitoids.

The second and third clast sets are derived from more distant sources. The second set includes clasts of unmetamorphosed Paleozoic rocks similar to limestone and dolomite of the middle Paleozoic section in the greater Mescal-Clark Mountain Range area (fig. 1). By contrast, Paleozoic strata in the nearby New York Mountains are metamorphosed. The second set of clasts also includes quartzite, schistose quartzite, and conglomerate of Cambrian to Late Proterozoic age and a small amount of red subarkosic sandstone that is probably Jurassic in age. Clasts of the Cretaceous Delfonte Volcanics were found in gravel high in the paleovalley section (R.E. Reynolds, personal commun., 1995). The clasts of Late Proterozoic quartzite, Jurassic sandstone, and Cretaceous volcanic rock pinpoint a source in the present-day Mescal Range, about 30 km to the northwest (Fig. 1). Clasts of metamorphic rocks include a suite of marble and coarse dolomite similar to metamorphosed Paleozoic strata in the southeastern New York Mountains. Common clasts of green-and-white striped hornfels are identical to the rock in the southeastern New York Mountains tentatively assigned to the Moenkopi Formation by Burchfiel and Davis (1977). Clasts of granitoids are predominantly derived from the Mid Hills Adamellite and Live Oak Granodiorite plutons of the Teutonia batholith (Beckerman and others, 1982; Miller and Wooden, 1993) in the Live Oak Canyon area (Fig. 2) of the southeastern New York Mountains. The Live Oak Granodiorite is areally restricted, indicating that the clasts of metamorphic and granitoid rocks were derived from an area about 10 km southwest of the paleovalley (Figs. 1, 2). These clasts generally are larger than those interpreted as derived from the Mescal Range.

The suite of mixed clast types strongly suggests a widely branching network of tributaries and/or a broad alluvial

plain bordered by the several reconstructed source terranes. The first alternative is preferred because rocks typical of the Cima Dome area and the southwest Ivanpah Mountains (Fig. 1) are not found, including the Ivanpah Granite and the Teutonia Adamellite. The Ivanpah Granite forms resistant highlands and is a prolific clast type in sedimentary basins east of Cima Dome. Peach Springs Tuff lies on part of Cima Dome, indicating that the rocks now exposed in that area were likely exposed during the Middle Miocene. With restoration of the Nipton fault zone (Fig. 1), sources of the Ivanpah Granite would have been closer than sources in the Mescal Range.

After the paleovalley filled, the fluvial system expanded to form a broad alluvial plain across much of the volcanic plain surrounding the paleovalley. The alluvial gravel deposit is several tens of meters thick and covers a broad area between Live Oak Canyon and Castle Peaks (Fig. 2). Nielson and others (1987, 1993) described similar gravel within the western Castle Mountains and in the northern Piute Range. The clast assemblage in this upper gravel is similar to the assemblage in the underlying paleovalley. In the Castle Mountains, the gravel lies on a paleosol(?) on Miocene volcanic rocks and in the Piute Range it occupies channel deposits (Nielson and others, 1987, 1993). The widespread Miocene gravels were eroded and then unconformably overlain by latest Pliocene(?) and Quaternary gravels (Miller and Wooden, 1993), some of which contain strongly-developed soils. This overlap by much younger gravel suggests that deposition of the Miocene gravel was followed by a long period of landscape stability.

Depositional model

The gravels in the Willow Wash paleovalley represent fluvial deposits in a deep channel. When the channel filled, the deposits formed a broad alluvial plain across the Castle Peaks and part of the Castle Mountains and northern Piute Range.

The paleovalley probably was carved by an aggrading fluvial system that overtopped a volcanic barrier. Headward erosion from the east is ruled out by a lack of local volcanic clasts at the base of the deposit, such as in the Castle Mountains. Stream deposits may have accumulated on the west side of the volcanic pile until the barrier was overtopped. Then a stream spilling across the barrier would have rapidly carved a canyon through the volcanic rock. Following incision, gravel on the west side of the barrier would have been eroded and redeposited east of the paleovalley. As depositional sites to the east aggraded, the paleovalley eventually filled with gravel. Finally, the gravel overtopped the channel and spread across the volcanic plateau north of Barnwell.

Age

The fluvial gravel in the Willow Wash paleovalley and conformably overlying alluvial gravel post-date rocks a little younger than the Wild Horse Mesa Tuff (~17.8 Ma), and significantly pre-date the present-day topography and associated late Cenozoic alluvial deposits. Nielson and others (1993) showed that, in the Castle Mountains, gravel they correlate with that in the paleovalley lies on 14.4- to 15.1-Ma volcanic rocks and contains interbeds of basalt near

the base. I suggest that the deposits they have described represent the upper, widespread part of the gravel sequence, so the middle Miocene age inferred from the Castle Mountains relations represents the upper part of the section.

Significant time elapsed between deposition of the paleovalley gravels and deposition of the locally derived Pliocene and Pleistocene gravel. Ivanpah Valley formed during this time interval, for it could not have existed in its present form at a time when streams carried clasts southeastward, depositing them in what are now the New York Mountains, Castle Mountains, and Piute Range. The Pliocene and Pleistocene gravel that overlies gravel of the Willow Wash paleovalley has a local source reflecting present-day topography.

An indirect clue for the minimum age for the gravels in Willow Wash paleovalley is given by middle Miocene (13 to 9 Ma) gravels in the Shadow Valley basin (Reynolds, 1991, 1992; Reynolds and Nance, 1988), west of the Clark Mountain Range (Fig. 1). These gravels also were derived from the Mescal Range. However, compared to the Willow Wash paleovalley, the Shadow Valley basin received a clast assemblage that corresponds more closely to the modern topography and that probably resulted from progressive downcutting of the source during basin development (Reynolds, 1992; Reynolds and Nance, 1988). These relations suggest that the Willow Wash paleovalley gravel source in the Mescal Range predated the Halloran basin. Therefore, the likely age for the Willow Wash paleovalley gravels is middle Miocene, probably between 17.5 and 13 Ma.

Tectonics

What tectonic or climatic event caused the development of the paleovalley and the larger drainage basin of which it was part? There are two main events that probably took place during the early to middle Miocene. Low-angle normal faults of the Colorado extensional corridor to the east were active, although their timing is poorly known. Elsewhere in the corridor, extension is approximately coincident with volcanic outpourings. If this relation holds locally, extension may have been about 18 to 14 Ma, the span of time for volcanism in the Castle Peaks and Castle Mountains (Nielson and others, 1993). The Castle Peaks lay west of the main Colorado River extensional corridor and was not subject to that pervasive faulting. The Castle Peaks probably underwent broad tilting and minor faulting that may have diverted drainage systems. In addition, a fairly thick pile of volcanic rocks was deposited near Barnwell over a period of about 1 to 2 million years. This accumulation may have disrupted drainage systems by changing surface topography, perhaps temporarily placing a barrier between source and sink for a previous drainage.

The clast assemblages indicate that the probable source terranes for most clasts in the Willow Wash paleovalley are located immediately southwest of the Clark Mountains fault (of Hewett, 1956), later described as the Kokoweef and Slaughterhouse faults (Fig. 1) on different sides of Ivanpah Valley (Burchfiel and Davis, 1977). The faults may have had reverse separation during the middle Miocene and highlands developed on the upthrown side of the faults likely would have been rapidly eroded. This scenario explains the clast types and the accumulation of a large amount of gravel derived from points to the northwest and

southwest of the paleovalley.

Kinematic indicators on the Slaughterhouse fault support the foregoing model, indicating that the last movement was probably down-to-the-northeast. Fault plane striae observed where the fault cuts marble of the Bird Spring Formation are nearly down-dip. Most fault planes strike N 50 W to N 70 W and dip about 65° SW, and most striae on these planes plunge steeply to the southwest. The evidence for nearly dip-slip movement conflicts with regional indicators of strike-slip separation on the fault (Burchfiel and Davis, 1977; Miller and Wooden, 1993), and the reverse separation probably occurred only during the most recent episode of faulting. Some fractures near the main fault bear subhorizontal striae that may reflect earlier episodes of strike-slip faulting. No kinematic indicators have been found on the Kokoweef fault. No other examples of reverse faulting are known for the Miocene in this region, although strike-slip faults such as the Nipton fault zone may have been active in the Miocene (Miller and Wooden, 1993) and indicate north-south shortening.

Although the timing of large-scale extension in the Colorado River corridor east of Willow Wash is not precisely known, it probably approximates the age of the footwall granite (~16 Ma; B.E. John, 1995, personal commun.) in the Newberry Mountains, and may have spanned a few million years. Most observations of detachment systems and models of depositional systems related to detachments indicate that the footwall block west of the breakaway of an east-dipping detachment fault will undergo down-to-the-west tilt. However, the information from the paleovalley gravel indicates that streams flowed eastward from the Mescal Range to the Piute Range, which lies adjacent to the inferred breakaway. Either tilts and topography changed rapidly following middle Miocene extension, or the tilt direction of the footwall block was opposite to that predicted by other studies. If the former hypothesis holds, rapid extension was probably about 18 to 16 Ma, and topography reversed for gravel deposition during the 16 to 13 Ma interval.

Willow Wash paleovalley gravels provide information on subsequent tectonism. Ivanpah Valley formed long after the paleovalley gravel was deposited. The structure of the valley is somewhat obscure because faults are mostly buried by alluvium, but north-northeast-striking faults in the western New York Mountains appear to have accomplished some of the basin formation, inasmuch as they drop the Miocene gravel and underlying volcanic rocks down to the west (Miller and Wooden, 1993). Another significant fault system in Ivanpah Valley is the Nipton fault zone that extends from northeast of the town of Nipton (Fig. 1) southwest toward Cima. The fault accomplished about 15 km of left-lateral separation of magnetic patterns in southern Ivanpah Valley (Swanson and others, 1980) and Early Proterozoic rock units northeast of Nipton (Miller and Wooden, 1993). It may have dilational *en echelon* steps that added to the development of modern Ivanpah Valley. Ivanpah Valley is bordered by long, gently-sloping alluvial fans on the east (northeastern New York Mountains) and steeper slopes on the west and southeast (Ivanpah and New York Mountains). These fans lap across eroded faults, and are unlike the steep fans of many actively faulted mountain fronts. These fans are middle to early Quaternary at oldest, indicating that southern Ivanpah Valley was not actively subsiding in the

Quaternary. The highly eroded landforms of the northeastern New York Mountains may indicate a Pliocene or late Miocene age for faulting of Ivanpah Valley.

Conclusions

A paleovalley cut through early Miocene volcanic rocks at the crest of the northeastern New York Mountains is middle Miocene in age. Clasts deposited in the valley were derived from two source terranes to the southwest and northwest. Both terranes probably were uplifted by reverse separation on the Slaughterhouse and Kokoweef faults, respectively. Eastward-sloping Miocene surfaces are opposite to those predicted for footwall rocks for the Colorado River extensional corridor to the east, and they may represent rapid relaxation of the crust after most extension took place. Ivanpah Valley formed by late Miocene faulting that reversed the topographic gradients on the western slopes of the present New York Mountains.

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Yucca schidigera in bloom near Foshay Pass. Providence Mountains distant. Front and back cover photographs by Robert E. Reynolds.

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