OLD ROUTES TO THE COLORADO

compiled by

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Cover photograph: Parker Dam at the Colorado River. *R.E. Reynolds photograph*

Back cover map: from Map of the Saline Deposits of the Southern Portion of California. *G.E. Bailey, California Division of Mines Bulletin No. 24, 1902.*

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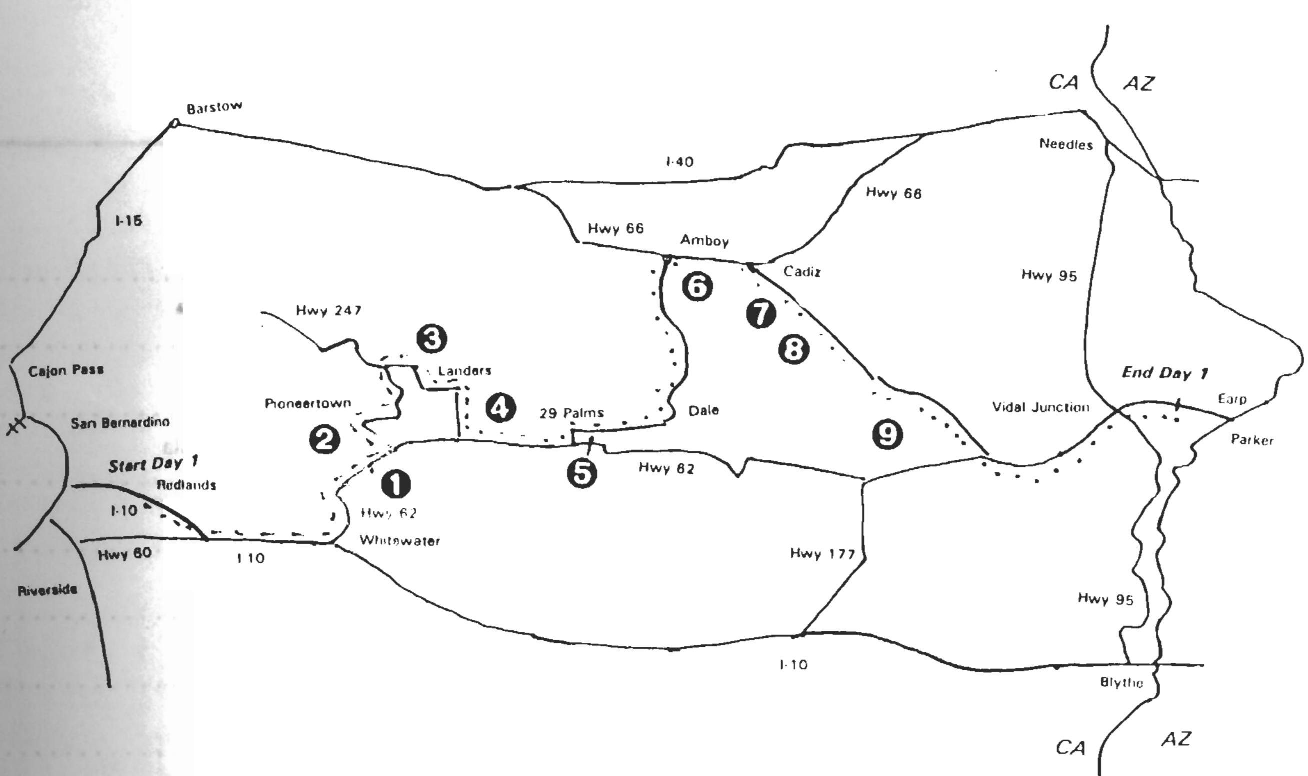

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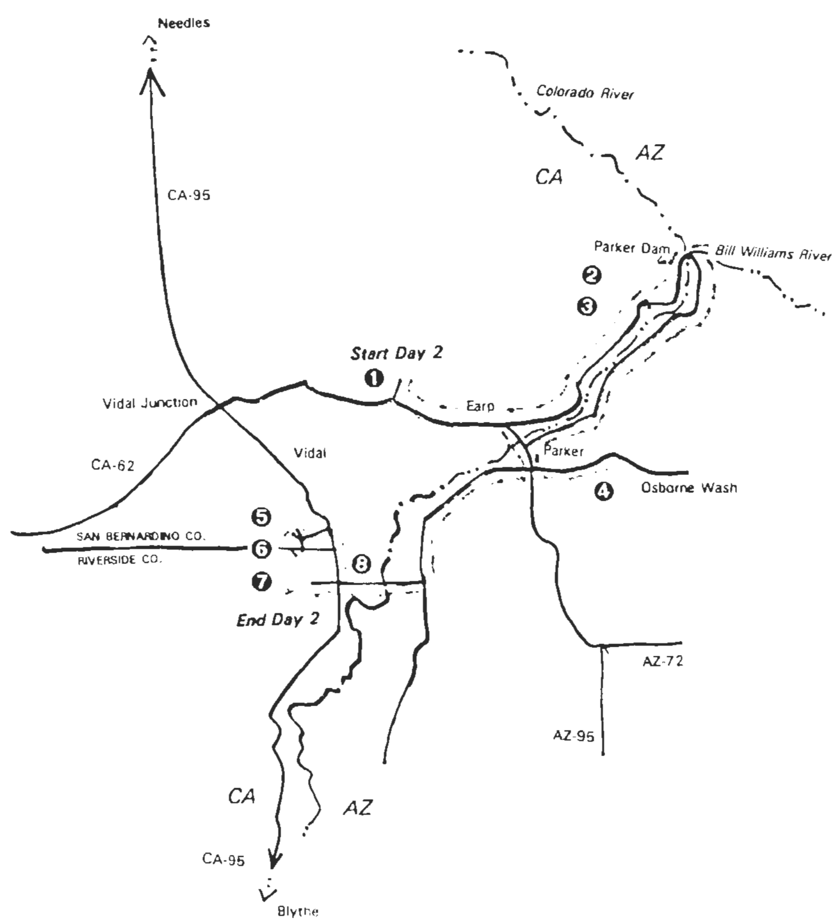

Day 2. The Colorado River Extensional Corridor

Old Routes to the Colorado

The 1992 Mojave Desert Quaternary Research Center Field Trip

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DAY 1

0.0 (0.0) START at the San Bernardino County Museum, 2024 Orange Tree Lane, Redlands. TURN RIGHT from parking lot entrance onto Orange Tree Lane, proceed to California Street.

0.2 (0.2) TURN LEFT at the stop sign onto California Street.

San Andreas Fault Segment

The San Andreas Fault system marks the boundary between the Transverse Range Province and the Penninsular Range Province. The right lateral San Andreas fault system, which includes the San Jacinto Fault to our south, controls the local geography and topography. We will be driving parallel to various components of the San Andreas system as we go southeast toward Palm Springs and Indio (Crowell, 1992).

0.4 (0.2) TURN LEFT onto Interstate 10 East, heading east toward Yucaipa and Palm Springs. Ahead you can see Yucaipa Ridge and San Bernardino Peak (elevation 10,525'). Yucaipa Ridge sits between the north branch (Mill Creek strand) and south branch (San Bernardino strand) of the San Andreas Fault.

Basement rocks similar to those found in the San Gabriel Mountains are overlain by the Mill Creek Formation where it is exposed on Yucaipa Ridge south of the mountain front. The Mill Creek Formation is a series of nonmarine Tertiary sediments deposited in a pull-apart basin (Demirer, 1985; Sadler and Demirer, 1986). The formation contains fossils which suggest an early Pliocene age (Axelrod, p.c. to Reynolds, 1985) but which may be as old as late Miocene or as young as middle Pliocene (Gibson, 1971). The Wilson Creek Fault (Matti and others, 1983 and 1985) crosses Yucaipa Ridge and separates the San Gabriel basement from basement rocks typical of the San Bernardino Mountains.

San Bernardino Peak and Mt. San Gorgonio are part of a complex of Precambrian biotite gneiss and schist, and granitoid gneiss intruded by Mesozoic quartz monzonite and granodiorite (Morton and others, 1980b). The massif was glaciated during the Pleistocene (Sharp and others, 1959; Dibblee, 1964).

2.3 (1.9) Holocene alluvium is on both sides of the freeway; to the right at about 2:00, Smiley Heights is on a Pleistocene alluvial surface with a Pleistocene soil (Reynolds and Reeder, 1986). This Pleistocene surface will be encountered repeatedly throughout the Yucaipa/Banning area.

2.6 (0.3) We are driving over terraced Pleistocene sediments near the Orange Street overpass.

3.8 (1.2) At University Street, the freeway crosses the zanja, California State Historical Landmark 43. The first irrigation project in the county, the zanja was constructed in 1819 and 1820 by Serrano and Cahuilla Indians under the guidance of Franciscan fathers from the Mission San Gabriel to develop agriculture at Guachama, the Indian rancheria near the site of the Asistencia mission branch in Old San Bernardino (Redlands) (Quinn, 1980).

4.2 (0.4) Cypress overpass. We are now driving on Pleistocene alluvium (Qoa of Morton, 1978).

4.8 (0.6) Cross the Redlands Fault, a normal fault which elevated Pleistocene alluvium on its southeast side. The trace runs southwest along Crescent Avenue across San Timoteo Canyon to join the San Jacinto fault zone.

5.1 (0.3) Ford Street off ramp. Houses at 10:00 are built on the Pleistocene erosional surface. The degree of soil development suggests the surface is of late but not terminal Pleistocene age.

6.1 (1.0) Reservoir Canyon. Cross the trace of the Crafton Fault (Reservoir Canyon Fault) offsetting Quaternary alluvium and uplifting Precambrian metamorphic and igneous basement rocks (Rogers, 1967). The Crafton Hills are a faulted complex of upper and lower plate rocks divided by the Vincent Thrust. Octavius Decatur Gass located gold-bearing quartz veins on the Yucaipa side of these hills in 1884. By 1889 the "Gold Bar Company" had developed a 60-foot tunnel and in 1890 the water-powered Yucaipa Quartz Mill had been constructed to process gold ore. The mine property was located in the canyon north of Crafton Hills College water tank; the mill site was in Dunlap Acres near 10th Street (Archer, 1976).

6.4 (0.3) Upper plate gneissic quartz diorite exposed in these road cuts, is separated from lower plate Pelona Schist by the Vincent Thrust. On the left, these exposures of the upper plate gneisses include Permo-Triassic Lowe Granodiorite and cataclasites. Pelona Schist is exposed in the road cuts to the left near the top of Reservoir Canyon.

6.8 (0.4) Marked by trees and bushes to the right, Crystal Springs comes to the surface at the fault trace. These springs supported a small bottled water industry in the past. Reservoir Canyon was named from the municipal water reservoir constructed for the Redlands Colony in 1881 (Archer, 1976).

Much of the Yucaipa area was drained through Reservoir Canyon in late Pleistocene times; the drainage was later captured through Live Oak Canyon (Dutcher and Burnham, 1960). Reservoir Canyon was the site of Maria Armenta Bermudez' pioneering farming activities in the area in 1836, when she raised vegetables for the Los Angeles market. Her crops were irrigated by a ditch dug from the zanja near present-day Crafton (Beattie and Beattie, 1951).

7.4 (0.6) Continue on Interstate 10 past the Yucaipa Boulevard exit. We are crossing from the upper plate rocks of gneissic quartz diorite into Pleistocene alluvium.

8.1 (0.7) Cross the Western Heights Fault, cutting Pleistocene alluvium. This fault, which bounds the Crafton Hills on the southeast, is subparallel to the Redlands Fault.

8.6 (0.5) Mount San Jacinto is seen ahead at 12:00; Pisgah Peak (elevation 5,480') is at 10:30. Pisgah Peak is south of the south branch of the San Andreas Fault and consists of upper plate granitic and granitoid gneissic rocks overlying the Vincent Thrust.

9.2 (0.6) Cross Live Oak Canyon Holocene alluvium.

9.8 (0.6) To the right is deep dissection in the Holocene alluvium overlying eroded Quaternary old alluvium of Live Oak Canyon. The dissection of these recent sediments has occurred since the start of agricultural development in the area, no more than 130 years ago (D. Morton, pers. comm. 1986).

11.2 (1.4) At County Line Road off ramp, we have returned to the Pleistocene surface. Fossiliferous Pleistocene sediments of the San Timoteo Formation beneath the Pleistocene surface are located between this off ramp and Calimesa Blvd. off ramp (Dibblee, 1981; Reynolds and Reeder, 1986 and 1991).

12.0 (0.8) Calimesa Boulevard off ramp. Continue on I-10.

12.5 (0.5) Cross tributary canyon of San Timoteo drainage on flat surface of Holocene alluvium. Here and in the next 0.6 mile, note again the depth of incision that has taken place in little more than 100 years.

14.2 (1.7) Terraces to the right are developed on Pleistocene and Holocene alluvium. The badlands topography is developed in the Plio-Pleistocene San Timoteo Formation (Reynolds and Kooser, 1986; Reynolds and Reeder, 1986, 1991). At 10:00 the terraces have been developed at a lower elevation than the badlands topography, and are truncated at their contact with the San Timoteo Formation. The northeast-striking valleys toward the skyline on the left are controlled by a branch of the Mission Creek Fault and the Vincent Thrust. These faults run northeasterly between the south branch of the San Andreas Fault (San Bernardino strand) and the Raywood Flat area on the skyline to the left (Matti and others, 1983).

14.3 (0.1) Cherry Valley offramp.

15.0 (0.8) Return to the Pleistocene surface. At 9:00, notice again how the Pleistocene terraces are truncated at the dissected San Timoteo Formation.

16.8 (1.8) The San Timoteo Canyon Road offramp enters San Timoteo Canyon. The freeway leaves the Pleistocene surface and crosses badlands topography and valley fill, regaining the Pleistocene surface near the junction of Highway 60. A terrace inset along San Timoteo Creek is about 200 years old (Reynolds and Kooser, 1986).

17.8 (1.0) CONTINUE on Interstate 10. Offramp to I-60 West is on the right.

18.9 (1.1) Beaumont Avenue/Highway 79 exit. Continue on I-10 East.

19.7 (0.8) San Gorgonio Pass is the lowest topographic break in southern California through the mountains to the inland deserts, separating Mt. San Gorgonio (11,502') and Mr. San Jacinto (10,786'), the two highest mountains in southern California. The crest of the pass, although broad and ill-defined, is the complicated junction of three major drainage basins: the interior-draining Whitewater River-Salton Trough to the east via Smith Creek; the generally interior-draining San Jacinto basin to the south via Potrero Creek; and the Santa Ana basin to the west via San Timoteo Canyon. The junction of these three basins is on the crest of an alluvial fan complex 2.5 miles north of I-10 between Noble Creek on the west and Smith Creek on the east.

We leave the Santa Ana basin and cross eastward to the San Jacinto basin. Through rapid headward erosion, Potrero Creek (to the right at 3:00) has progressed northward, extending the northward limit of the San Jacinto basin along the crest of the alluvial fan complex essentially to Highland Springs, 2.5 miles north of Interstate 10.

20.5 (0.8) At Highland Springs Avenue we leave the San Jacinto basin and cross eastward into the Whitewater River drainage, to the right.

21.0 (0.5) At 10:00, the Banning Bench is bounded on the south by an unnamed thrust fault, and capped by the Heights Fanglomerate of Allen (1957). The deposit is dominated by deeply weathered clasts of gray migmatitic gneiss and greenschist (Pelona Schist) which is probably derived from the upper San Gorgonio River area near the juncture of the Mission Creek and San Bernardino strands of the San Andreas Fault. *Bison* remains have been recovered from the Heights Fanglomerate (Jefferson, 1986) indicating that it is less than 500,000 ybp (Savage and Russell, 1983). The Heights Fanglomerate unconformably overlies sediments similar in appearance to the San Timoteo Formation, which coarsens north of the Banning Fault.

From this point eastward to Whitewater, we enter an area dominated by compressional features.

24.1 (3.1) Pass the exit for Highway 243 to 8th Street and Idyllwild.

26.4 (2.3) The houses straight ahead are built on a surface cut by dissected thrust fault scarps (Bortugno and Spittler, 1986; Dibblee, 1982). In the hills to the left, the Banning Fault has thrust basement rocks over non-marine sandstones, siltstones, and conglomerates of the Hathaway Formation. In Lion Canyon, the Hathaway Formation is conformably overlain by the marine Imperial Formation which is in turn conformably overlain by the nonmarine Painted Hill Formation. Elsewhere, the Hathaway Formation is directly overlain by the Painted Hill Formation (Allen, 1957). These three formations, Pliocene in age, are caught up between thrust faults along the base of the mountain front from this point to Stubbe Canyon (Allen, 1957; Dibblee, 1982). Allen (1957) divided the Hathaway Formation into two members, a sandstone-dominated lower member and a conglomerate-dominated upper member distinguished by clasts of laser gneiss derived from an area north of the Banning Fault between Cottonwood and San Gorgonio canyons. He also mentioned rare clasts of silicified limestone without speculating upon their possible source.

The San Gorgonio igneous-metamorphic complex in this area is predominantly migmatitic gneiss with intrusions of quartz monzonite (Morton and others, 1980b).

27.0 (0.6) The Cabezon Fanglomerate (lower hills straight ahead) has been anticlinally folded and cut by thrust faults. The Quaternary Cabezon Fanglomerate includes gravels from a variety of sources.

27.6 (0.6) To the left, beneath the water tank, is the most youthful thrust fault scarp in this area related to compression associated with activity along the Banning Fault. At this point, the scarp changes orientation from a northwest strike to a northeast strike.

28.3 (0.7) To the right at 1:00 is the north portal of the San Jacinto Tunnel, a part of the Colorado River Aqueduct system. It cuts through the Paleozoic? metasediments (quartzofeldspathic gneiss and schist, phyllite, quartzite, and marble) intruded by quartz diorite of Mt. San Jacinto (Morton and others, 1980a).

To the left is Millard Canyon; a fault scarp crosses the alluvial fan near the canyon mouth. The debris of the Millard Canyon fan overwhelms debris from Mt. San Jacinto. Drainage to the base of Mt. San Jacinto is thus forced eastward from this point to the Whitewater River.

28.9 (0.6) Cabezon exit. Continue along freeway. To the right, the steep escarpment of the San Jacinto Mountains is interpreted to be the result of uplift on the postulated South Pass Fault (Allen, 1957).

31.2 (2.3) Dinosaurs to the north!!

31.5 (0.3) Good exposures of the Cabezon Fanglomerate are to the left. Hathaway, Imperial and Painted Hill sediments are thrust over the Cabezon Fanglomerate and are in turn overthrust by the San Gabriel igneous-metamorphic complex. Landslides are common at the noses of the ridges.

To the left at 11:00, Lion Canyon is bounded on the east by a large landslide. The upper "boundary" of this landslide is in the Cabezon Fanglomerate and, as shown by Allen (1957), is convex and points to the south. This is contrary to a landslide headscarp and, because pressure ridges are also apparent within the landslide, suggests that the feature is the result of "bulldozing" by a larger mass to the north and not simply a slope failure.

32.4 (0.9) A thrust in the basement rocks to the left at 10:00 at Stubbe Canyon is seen where pink piemontite-bearing rocks are thrust over green epidote-bearing rocks. The distinctive piemontite-bearing gneisses are found as clasts in sediments north and south of the Banning Fault. Since the source area is of limited extent, this has proven useful in estimating fault offset as well as identifying source areas and transport directions (Allen, 1957).

33.2 (0.8) The Banning Fault changes from a low angle fault to a steep angle fault (Reynolds and Kooser, 1986).

34.3 (1.1) Based on geophysical evidence, the ridge of metamorphic rocks (ahead at 12:00 extending from Mt. San Jacinto) continues beneath the alluvium to a point north of the freeway and northward of the southernmost thrusts characteristic of the San Bernardino Mountains' side of the pass. This ridge reduces the energy of the strong winds which are regularly funneled through San Gorgonio Pass, and dune sands are deposited against it.

35.2 (0.9) Whitewater Gravels of the Cabezon Fanglomerate are to the left at 11:00 (Whitewater Hill). The gravels are capped by a Pleistocene soil.

36.0 (0.8) Verbenia exit; continue on Interstate 10.

36.7 (0.7) Highway 111 to Palm Springs passes through the old Whitewater Ranch property. Do not exit. Landslide deposits are to the left.

37.0 (0.3) To the left, the Garnet Hill Fault disrupts alluvium 2/3 of the way from the freeway to the base of the hills. The fault runs across the mouth of Whitewater Canyon where it is visible at 9:00. The fault trace is exposed only west of Whitewater River. Based on trenching between Cottonwood and Whitewater Canyons, there is no evidence for Holocene activity on the Garnet Hill Fault (Reeder, p.c. 1986, cited in Reynolds and Kooser, 1986). The Garnet Hill Fault displaces Pleistocene-age Whitewater gravels of Windmill Hill (Allen, 1957). To the east, its trace is covered by alluvium and the main evidence for its existence within the Coachella Valley is a strong gravity anomaly. Gravity low contours define a trough which is almost as well delineated as the gravity troughs associated with the Banning and Mission Creek faults (Proctor, 1968). Proctor suggests that the Garnet Hill Fault may be an ancestral branch of the San Andreas Fault.

37.6 (0.6) South of the Interstate, large cottonwood trees and scant building ruins mark the site of the Whitewater Ranch headquarters. Pauline Weaver and Isaac Williams were the first Anglos to own land in the San Gorgonio Pass; their San Gorgonio Rancho was granted in 1845 and encompassed the entire pass area. Weaver sold a portion of the rancho to Isaac Smith in 1853; this purchase, which included the land from Beaumont to Palm Springs, was to develop into the Whitewater Ranch. The riparian water rights from the Whitewater River granted in 1850 passed with the ranch to successive owners and allowed ranching to continue. The site was also a regular freight and stage stop along the Butterfield route (Stocker, 1973).

37.8 **(0.2)** Rest area at Whitewater Ranch site.

38.5 **(0.7)** Whitewater Road exit; continue on I-10.

39.0 **(0.5)** Beneath the three buildings at 11:00 (left) is the reverse fault scarp of the Garnet Hill Fault.

39.3 **(0.3)** Cross the Whitewater River.

39.8 **(0.5)** The north side of the freeway runs along the trace of the Garnet Hill Fault next to Whitewater Hill. To the left are Pleistocene fan sediments of the Cabezon Fanglomerate separated from the Imperial and Painted Hill formations (Murphy, 1986) by the Banning Fault. The Cabezon Fanglomerate of Whitewater Hill includes a lens of limestone breccia believed to have been derived from the San Jacinto block (Allen, 1957). Proctor (1968) notes that Whitewater Hill has been uplifted so recently that relict drainages exposed on its surface do not conform to its current topography.

Move to the right lane and prepare to exit.

40.7 **(0.9)** EXIT RIGHT on the Yucca Valley—29 Palms offramp, following Highway 62 northward over the freeway.

41.1 **(0.4)** View southeast down the axis of the Salton Trough. The Garnet Hill Fault trace is on the south side of the low hills (Garnet Hill).

42.2 **(1.1)** Dillon Road. Red exposures at the Whitewater Rock Quarry are visible to the left at 9:00.

42.6 **(0.8)** To the right at 1:00, the trace of the Banning Fault is expressed as shutter ridges between the powerline and windmills. Devers Hill protrudes through the alluvium to the right.

42.9 **(0.3)** Cross the Banning Fault over the next 0.1 mile.

44.8 **(1.9)** Pierson Blvd. Mt. San Gorgonio is viewed to the left at 10:00; to the right at 2:00 are the Little San Bernardino Mountains.

46.4 **(1.6)** Mission Creek Road crosses Highway 62. To the left are dissected Mission Creek alluvial deposits cut by northeast-striking faults with the east side down. To the right at 2:00 is a fault-bounded prism of pinkish sediments against the mountain front which is bounded by the Mission Creek strand of the San Andreas fault system.

47.1 **(0.7)** Cross Mission Creek Wash for the next 0.3 miles.

47.9 **(0.8)** Indian Avenue; continue on Highway 62.

48.1 **(0.2)** Cross the Mission Creek Fault of the San Andreas fault system as you head up Dry Morongo Canyon, entering Mesozoic deformed pluton and Precambrian gneiss. We are leaving the segment of the field trip that is controlled by the right lateral San Andreas fault system and entering the segment of the trip that is influenced by the left lateral Pinto Mountain fault system.

Pinto Mountain Fault Segment

A portion of the Transverse Range Province lies north of the San Andreas fault system and south of the left lateral Pinto Mountain Fault. The Pinto Mountain Fault is a major left-lateral fault which represents the southern structural boundary of the Mojave block (Dibblee, 1992). The Mojave Desert is characterized by a series of active northwest-trending right lateral faults. These faults apparently terminate at or are truncated by the Pinto Mountain Fault. We will be traveling parallel to the left lateral Pinto Mountain Fault until we reach Twentynine Palms.

50.4 **(2.3)** Cross the trace of the Morongo Valley Fault, trending northeast towards Morongo Summit, where it intersects with the Pinto Mountain Fault.

50.7 **(0.3)** To the left is perched alluvium.

51.0 **(0.3)** The highway enters fault-bounded Morongo Valley, with the Pinto Mountain Fault on the north side of the valley and the Morongo Valley Fault on the south side. Morongo Valley drains southward into the Whitewater drainage, which runs through the Coachella Valley and into the Salton Sea.

52.2 **(1.2)** Covington Park and the Big Morongo Wildlife Refuge are to the right via East Drive. The nature reserve is a habitat for more than 240 species of resident and migrant birds as well as a sanctuary for mammals including big horn sheep. Permanent water, brought to the surface at springs along the Morongo Valley Fault, supports a lush riparian community. Continue on Highway 62.

52.7 **(0.4)** A landfill is to the right at 2:00. Note that ridges are terminated by the *en echelon* Morongo Valley Fault east of Big Morongo Canyon. The terrace at the east end of the landfill is capped by a well-developed red soil horizon.

54.0 **(1.3)** The Pinto Mountain Fault runs on the north side of the valley north of the highway. As you look ahead toward the pass, you see the intersection of the Pinto Mountain Fault and the Morongo Valley Fault.

56.3 **(2.3)** Pass Ole Street.

56.5 **(0.2)** Light gray granitic bedrock to the left is separated from overlying brownish granitic bedrock by low angle faults and shears. Note the vegetation growth along the fault contacts. North, at 9:00, the Pinto Mountain Fault crosses near the house (at 11:00) and water tank.

57.0 **(0.5)** Pass Highland Street.

Figure 1. Stop 2. White arkosic sediments capped by basalt north of Pioneertown. *R.E. Reynolds photo.*

57.7 (0.7) Pass Hoopa Road.

58.0 (0.3) Morongo Valley Park. The contact above the shooting range exhibits gray granitic bedrock below overlying reddish bedrock.

58.4 (0.4) The leveled pad at 2:00 on right exposes vertically dipping braided stream deposits.

59.0 (0.6) We are entering the Yucca Valley drainage, which runs eastward along the Pinto Mountain Fault to Copper Basin, and then eastward to Mesquite Lake at Twentynine Palms.

59.5 (0.4) TURN RIGHT off Highway 62 onto Piñon Drive; proceed up hill.

59.7 (0.2) TURN RIGHT on Navajo; proceed to end.

60.0 (0.3) STOP 1. PINTO MOUNTAIN FAULT SEDIMENTS. (see Grimes, this volume). Park at end of cul de sac; do not enter private property. We are near the intersection of the Pinto Mountain Fault and the Morongo Valley Fault. From this vantage point, note the sediments to the north, which contain clasts of basalts with ultramafic inclusions (kaersutite). This fanglomerate overlies and is in fault contact with the quartzite fanglomerate which composes the relatively flat surfaces to the south. The basin between us and these surfaces contains finer-grained arkosic sediments which dip steeply to the north. The arkose is a fault-bounded wedge unconformably overlain by the capping quartzite fanglomerate (Grimes, 1986).

Return to Highway 62, preparing to turn east (right). The Sawtooths are visible against the horizon at 1:00.

60.5 (0.5) TURN RIGHT onto Highway 62 and continue easterly. Look ahead for the flashing yellow traffic lights, where we will be turning left.

62.1 (1.6) TURN LEFT on Pioneertown Road, just past flashing yellow pedestrian crossing lights. To the right, this road is called "Deer Trail". The drainage here runs to Copper Basin.

62.2 (<0.1) Stop sign at Yucca Trail. Proceed ahead on Pioneertown Road.

62.7 (0.5) Cross the most northern suspected trace of the Pinto Mountain Fault as mapped by Dibblee (1967a).

64.5 (1.8) Water Canyon Fault. Notice that the fluviatile sediments past Water Canyon, exposed below the terrace on the left, are undisturbed by the Water Canyon Fault.

65.0 (0.5) During the Tertiary, granitic basement rocks were deeply weathered along joint sets (Oberlander, 1972). Recent weathering has exposed this boulder terrain in the Sawtooths.

65.9 (0.9) Ahead and to the right are the dark Pioneertown Basalts and white patches of Tertiary sediments. The Pioneertown Basalts cover an area of approximately 22 km^2 and may reach a thickness of 60 m. The pile is made up of individual flow units three to seven m thick, each capped by a terminal vesiculated or amygdoloidal top. Eight or nine individual flow units have been observed in the thickest portion of the pile. These basalts are alkali olivine in composition (Neville, 1983); potassium/argon dates for similar flows range from 6.9 to 9.3 Ma (Morton, p.c. 1985, cited in Reynolds and Kooser, 1986; Peterson, 1976; Oberlander, 1972). The basalts overlie and are interbedded with Tertiary arkose deposits and overlie granitic basement. In some places, Tertiary granitic soil horizons are preserved beneath the flows (Oberlander, 1972). The basalts are correlative, in terms of time and petrogenesis, with other alkaline volcanics found throughout the Mojave Desert, such as Cima Dome, Amboy Crater, Dish Hill, and Pisgah Crater (Neville, 1983; Neville and others, 1985, and see Reynolds, this volume; Lawton, this volume; and Hazlett, this volume).

66.3 (0.4) Pioneertown was built as a set for western movies. It was named by Dick Curtis, an actor, on Labor Day 1947 (Gudde, 1974).

66.7 (0.4) Pavement turns right 90 degrees to the northeast; continue along Pioneertown Road. Chaparrosa Springs is in bedrock to the left.

67.0 (0.3) Cross Chaparossa Wash.

67.5 (0.5) Leaving wash, proceed along road to top of terrace; prepare to turn right.

67.8 (0.3) TURN RIGHT onto dirt road marked with rock gate structure and wooden post. Take roads to left, watching for vehicle-size ruts.

68.1 (0.3) STOP 2. TERTIARY PIONEERTOWN SEQUENCE. Park within view of Pioneertown Basalts overlying and interfingered with Tertiary arkose. Tertiary sediments are rare in the eastern San Bernardino Mountains. This section has been referred to as the Old Woman Sandstone by Dibblee (1967b), but the difference in clast lithologies and the age of the overlying basalts indicate that their age and source differ significantly from the Old Woman Sandstone. Similarly, lithology and stratigraphy distinguish this arkose from the Santa Ana Sandstone Fragmentary vertebrate fossils appear to corroborate an age greater than 7 Ma but less than 15 Ma for the lower silty sediments, which suggests they are time correlative with the upper portion of the Crowder Formation in Cajon Pass (Reynolds, this volume).

Walk ahead about 0.2 mi to road cuts exposed faulted sediments. In the cut you can see, from lowest: (1) low energy deposition of brown silty sands and paleosols; (2) higher energy deposition of arkosic sands; (3) vertical faults offsetting the sedimentary section downward, to the east; (4) a possible erosional surface and soil which may have formed prior to basalt flows; and (5) 6.9—9.3 Ma basalts laid down on undulating topography and interfingering with arkose.

Piñon, juniper, joshua tree, scrub oak, manzanita, Mojave yucca, and nolina are members of this handsome plant community (Fig. 1).

RETRACE ROUTE along ruts to pavement.

Figure 2. Stop 3. Pleistocene sediments capped by a dense layer of calcium carbonate along Linn Road. *R.E. Reynolds photo.*

68.4 (0.3) TURN RIGHT onto Pioneertown Road and resume route northwest. The basalts appear to have flowed over a gently undulating surface and are thickest to the southeast. The apparent flow direction was roughly northeast to east. The location of the source vent of these volcanics is unknown (Vaughan, 1922).

70.2 (1.8) TURN RIGHT onto Pipes Canyon Road; proceed northwest.

70.5 (0.3) View ahead of mesas of the Pioneertown basalt flow including Flat Top Mountain to the northeast and Black Hill to its south.

70.8 (0.3) Cross Pipes Wash. Water rises to the surface in Pipes Wash as a result of the shallow bedrock between the volcanic tablelands.

73.3 (2.5) To the right at 1:30, note the crude columnar jointing in basalts overlying a middle Tertiary erosional surface that developed on granitic rocks.

74.3 (1.0) Approximately 8 individual flow units make up the basalt pile to the left.

74.8 (0.5) Coarse Pleistocene gravels are deposited against Tertiary arkose to the south at 2:00 in the bank of Pipes Wash. The dark varnish on the basalt scree at 10:00 indicates that this debris may have been stable since middle Pleistocene times.

76.3 (1.5) Quaternary stream deposits of Pipes Wash are exposed to the left and in road cuts. Pipes Wash and Chaparossa Wash drain northerly and empty into Emerson Lake basin.

76.9 (0.6) Reach Old Woman Springs Road, Highway 247, GO NORTH (LEFT) towards Flamingo Heights.

78.8 (1.9) Pass Chaparral Road to the left in downtown Flamingo Heights.

80.6 (1.4) Hondo Street. Continue on Highway 247.

81.7 (1.1) A deep wash from Bolo Springs cuts through Pleistocene sediments.

82.2 (0.5) Reche Road. Continue north on Highway 247.

82.9 (0.7) New Dixie Mine Road. To the west are mantle xenoliths discussed by Neville (1986).

84.0 (1.9) Lum Lane. Prepare to turn right at Linn Road.

84.5 (0.5) TURN RIGHT (east) onto Linn Road.

84.6 (0.1) Cross the trace of the Johnston Valley Fault. We are between the left lateral Pinto Mountain Fault (the southern margin of the Mojave Desert Province); the northern margin of the province is the left lateral Garlock Fault, 90 miles north. The right lateral San Andreas Fault forms the southwest margin of the Mojave Desert Province.

87.7 (3.1) STOP 3. HOMESTEAD VALLEY FAULT. Park off pavement. Linn Road cuts through Pleistocene sediments: red-brown gravel, gray-brown sands, capped with a 2-foot layer of dense calcium carbonate (Fig. 2). The sharp contact between the sediments and the carbonate layer suggest that the layer is not of pedogenic origin, but may be the result of a groundwater barrier and associated springs. The Homestead Valley Fault is immediately east (see Umbarger, this volume) and Pipe's Wash drainage follows the fault for a short distance.

The Mojave Desert Province is cut by a series of northwest-trending right lateral faults that parallel the San Andreas Fault. The Johnston Valley Fault is the first of these northwest-trending faults that we will cross. We will also cross, in order, the Homestead Valley Fault, the Copper Mountain Fault, the Hidalgo Mountain Fault, the Emerson Fault, and the Mesquite Lake Fault.

Continue east on Linn Road. Goat Mountain is west; Giant Rock is 2 miles east.

87.8 (0.1) TURN RIGHT onto Belfield Blvd, proceed 2 miles south to Reche Road. We are paralleling the projected trace of the northwest-trending Homestead Valley Fault (Hill and others, 1980).

89.9 (2.1) Stop sign. Go east (left) on Reche Road. Cross Pipes Wash drainage which runs north into Emerson Lake basin.

90.5 (0.6) Pass Landers Road.

92.4 (1.9) Hidalgo Mountain, north, is bounded on the north by the West Calico Fault and on the south by the Hidalgo Mountain Fault. Gypsum Ridge, the low ridge at 9:30 southeast of Hidalgo Mountain, consists of Pleistocene lacustrine sediments deformed along the West Calico Fault (Dibblee, 1967; Knauer, 1982).

96.0 (3.6) TURN RIGHT (south) on Border Avenue. Cross the trace of the Emerson Fault near its intersection with Copper Mountain Fault, which runs southeast at 10:30. We are driving over calichified Pleistocene fans, and we have left the Emerson Lake drainage system and have entered the drainage which runs northeast to Deadman Lake and Bouillon Wash. Surprise Spring, which is the locality of a Pleistocene fauna discussed by Jefferson (this volume) is to the east-northeast at 8:00.

100.0 (4.0) TURN LEFT onto La Brisa Drive, a graded dirt road. We have left the Deadman Lake drainage and have entered the drainage that runs parallel to the Pinto Mountain Fault and then to Copper Basin and Mesquite Lake.

102.2 (2.2) VIEWPOINT. Copper Playa is at 2:00, in front of Copper Mountain (Reynolds and Jenkins, 1986), visible at 12:00 and extending to the southeast. The Copper Mountain Fault is on the west side of Copper Mountain and cuts southeasterly through the mountain. The view also includes eroded granitic rocks of the Little San Bernardino Mountains on the distant skyline, part of Joshua Tree National Monument.

103.1 (1.1) TURN RIGHT onto Sunfair, a paved street (except not at the intersection).

104.5 (1.4) Playa sediments (elevation 2380') of Copper Basin are at an elevation 20' higher than the present-day playa surface (which contained water in February 1992).

106.8 (2.3) Pass the High Desert Airport on the outskirts of Sunfair.

107.4 (0.6) TURN LEFT onto Pole Line Road. We are traveling parallel to the trace of the Pinto Mountain Fault. Copper Basin playa, to the north (left), receives water from Yucca Valley and, during the Pleistocene, may have overflowed eastward into Mesquite Lake.

108.4 (1.0) Intersection of Cascade Road. Continue on pole line road.

108.9 (0.5) STOP 4. COPPER BASIN LACUSTRINE SECTION. Green lacustrine sediments capped by caliche on the south side of the Pinto Mountain Fault contain Pleistocene vertebrate fossils (Fig. 3).

109.4 (0.5) Arkosic sediments dip steeply to the southwest on the south side of the Pinto Mountain Fault (Fig. 4). The intersection of the left lateral Pinto Mountain Fault and the right lateral Copper Mountain Fault is approximately 3 miles ahead (east) on the east side of Copper Mountain.

109.6 (0.2) Saddle at elevation 2410+'. If Copper Basin filled during the Pleistocene, it would have drained here. Lacustrine sediments seen at Stop 4 are at the elevation of this saddle. Notice the dissected terraces with well-developed soil on the south side of Copper Mountain.

109.8 (0.2) TURN RIGHT (south).

110.0 (0.2) TURN RIGHT (west).

110.3 (0.3) TURN LEFT (south) at intersection of dirt road and Rotary Way.

111.0 (0.7) TURN LEFT (east) at stop sign at intersection of Rotary Way and Highway 62.

111.9 (0.9) Twenty-nine Palms city limits. *Make certain your vehicle gets fueled in Twentynine Palms; the next reliable gas stop is not until Parker.*

114.3 (3.4) Indian Cove Road. Continue on Highway 62.

Figure 3. Stop 4. Green lacustrine sediments at Copper Basin. *R.E. Reynolds photo.*

114.6 (0.3) To the north at 10:00 is a thick section of Pleistocene sediments between the Hidalgo Mountain Fault, on the east side of Copper Mountain, and the Mesquite Lake Fault. These sediments are cut by the drainage from Copper Basin into Mesquite Lake. The bend in the drainage is on the east side of the Hidalgo Mountain Fault.

118.1 (3.5) Larrea/Manzanita Avenue. Continue on Highway 62. We are driving easterly along the trace of the Pinto Mountain Fault. The southeast branch of the Pinto Mountain Fault runs to the south side of Donnell Hill, south of Hwy 62.

119.1 (1.0) Mesquite Springs Road. We are on the trace of the Pinto Mountain Fault as Highway 62 goes up the mid-Pleistocene alluvium of Donnell Hill.

119.6 (0.5) TURN RIGHT on Boullion Avenue; proceed south.

119.7 (0.1) Stop sign at Cactus Drive.

119.8 (0.1) Stop sign at Old Dale Drive.

119.9 (0.1) Cross the trace of the southeast branch of the Pinto Mountain Fault. Notice scarp on the south side of Donnell Hill.

120.1 (0.2) TURN LEFT (east) at stop sign at intersection with Sullivan Road.

120.6 (0.5) TURN LEFT (north) at stop sign onto Adobe Road.

120.7(>0.1) TURN RIGHT (east) onto Cottonwood Drive; the pavement turns to dirt. Pass the 29 Palms Inn. Groves of native palm trees grow along the trace of the Pinto Mountain Fault (Cornett, 1991).

121.2(<0.4) Stop sign. TURN RIGHT onto National Monument Drive. Go south, then east along the trace of the Pinto Mountain Fault. The palms on the right are on the trace of the Pinto Mountain Fault.

121.8 (0.6) Stop sign at Utah Trail and the Joshua Tree National Monument Visitor Center. TURN LEFT (north) and proceed along Utah Trail.

122.2 (0.4) Stop sign at Twenty-nine Palms Highway. Continue north on Utah Trail.

122.5 (0.3) Utah Trail cuts through calichified sediments below desert pavement. Campbell Hill is to the northeast at 2:00.

123.3 (0.8) Stop sign at Two Mile Road.

124.0 (0.7) TURN RIGHT (east) onto Michaels Road.

Figure 4. Arkosic sediments dip steeply near the pass at Copper Mountain. *R.E. Reynolds photo.*

124.2 (0.2) Cross the flood control drainage. The pink house (ahead) belonged to Elizabeth W.C. and William H. Campbell, famous for their investigative archaeologic work in this area (Campbell and Campbell, 1935).

124.4 (0.2) At the first pole line, cross the trace of the Mesquite Lake Fault.

124.5 (0.1) TURN SHARP RIGHT (south) on dirt track; do not take the road to the Campbell home.

124.7 (0.2) Drive through limonite-stained sediments along the trace of the Mesquite Lake Fault.

124.8 (0.1) STOP 5. CAMPBELL HILL. Late Pleistocene sediments at Campbell Hill (Fig. 5) have been uplifted along the northeast side of the Mesquite Lake Fault (Dibblee, 1968; Jagiello and others, 1992; and see Foster, this volume). Vertebrate remains are typical of the Rancholabrean Land Mammal Age and include ground sloths, dwarf pronghorn, sabertooth, mammoth, horse, and camel (Jefferson, this volume). Bachellor (1978) tentatively identified the Bishop Tuff in the section, which is dated at 0.73 Ma.

The Bullion Mountains are due north; Hidalgo Mountain is N30°W, and Gypsum Ridge the low ridge to the right. The Rodman Mountains are left of Hidalgo Mountain. Copper Mountain is due west and Goat Mountain is N55°W, just to the left of a rise in the Pleistocene sediments which lie between the Emerson/Copper Mountain Fault and the Hidalgo Mountain Fault.

USE CAUTION WHEN TURNING AROUND; it is very sandy. RETRACE route to Utah Trail.

125.7 (1.9) Stop sign. TURN RIGHT (north) on Utah Trail.

125.9 (0.2) Stop; TURN RIGHT on Amboy Road. Shortz Lake (elevation 1795') is one-half mile northwest, and Mesquite Lake is about 2.5 miles further north-northwest. If Dale Lake received water from the Mesquite Lake basin, the basin would have been filled to an elevation of 1800'.

127.0 (1.1) We are between the Mesquite Lake Fault and the north branch of the Mesquite Lake Fault. The drainage one mile north of the road leads to Dale Lake.

130.4 (3.4) Enter calichified Pleistocene sediments. We are going north into Wonder Valley from the Pinto Mountains. Note the thick sequence of sediments uplifted at the base of the Pinto Mountains near the junction of the Bullion Mountain Fault and the combined trace of the Mesquite Lake Fault.

130.6 (0.2) An energy-efficient jojoba farm is on the right.

132.0 (1.4) Cross the approximate trace of the northwest-trending Bullion Mountain Fault.

133.6 (1.6) Wonder Valley Fire Station. We are beginning to enter the sand dunes of the Dale Lake system, visible to the south against the Little San Bernardino Mountains.

136.9 (0.3) "The Palms" watering hole is to the right.

142.7 (5.8) Barnett's Trading Post (no gas). Look right to Dale Lake. The Dale Lake salt works are in the playa surface at 1:00, below Clarks Pass (see Gundry, this volume, for a summary of salt production in the playas we will cross). Dale Lake (elevation 1200') is an internally drained basin that may have received overflow waters from Yucca Valley (3400'), Copper Basin (2400'), and Mesquite Lake (1750') during Pleistocene times. Schroth (this volume) reviews the prehistory of this area; Tchakerian (this volume) discusses the dune field at Clarks Pass.

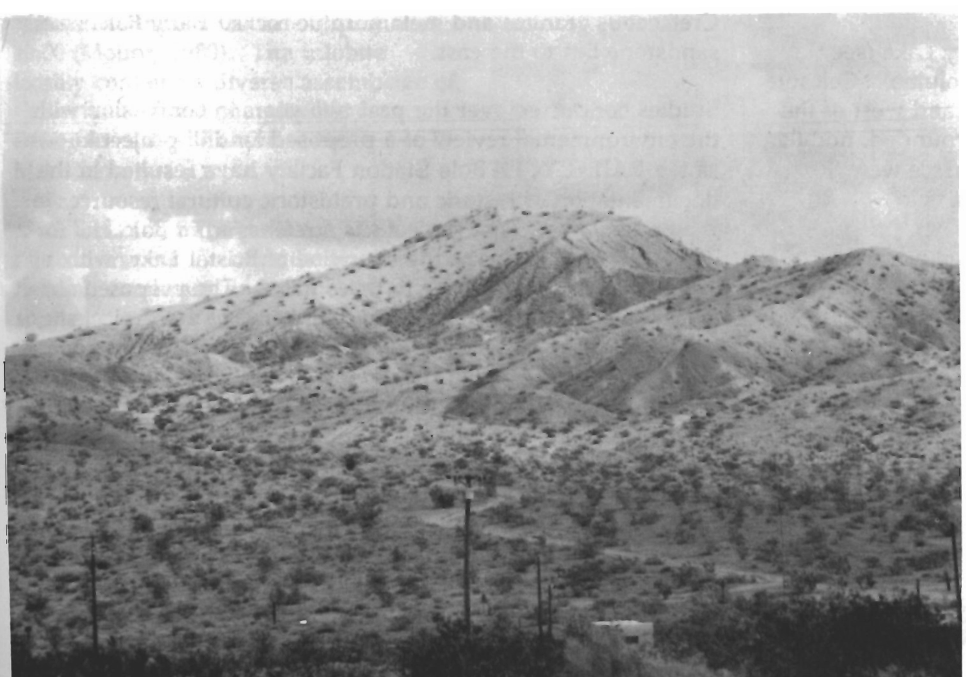

Figure 5. Stop 5, Campbell Hill. *R.E. Reynolds photo.*

143.1 (0.4) Road bends northeast. The Sheep Hole Mountains are at 2:00.

144.1 (1.0) We are crossing the trace of a northwest-trending fault that runs through the eastern Bullion Mountains and projects toward the trace of the Ludlow fault zone.

145.8 (1.7) A northwest-trending fault is mapped as separating the steep granitic escarpment of the Sheephole Mountains from dissected, calichified alluvial fans or debris flows. When we reach Sheep Hole Pass, we will cross a pediment where no alluvial fans are preserved.

Bristol/Danby Trough Segment

148.6 (2.8) Sheep Hole Summit, elevation 2368'. Microwave relay tower

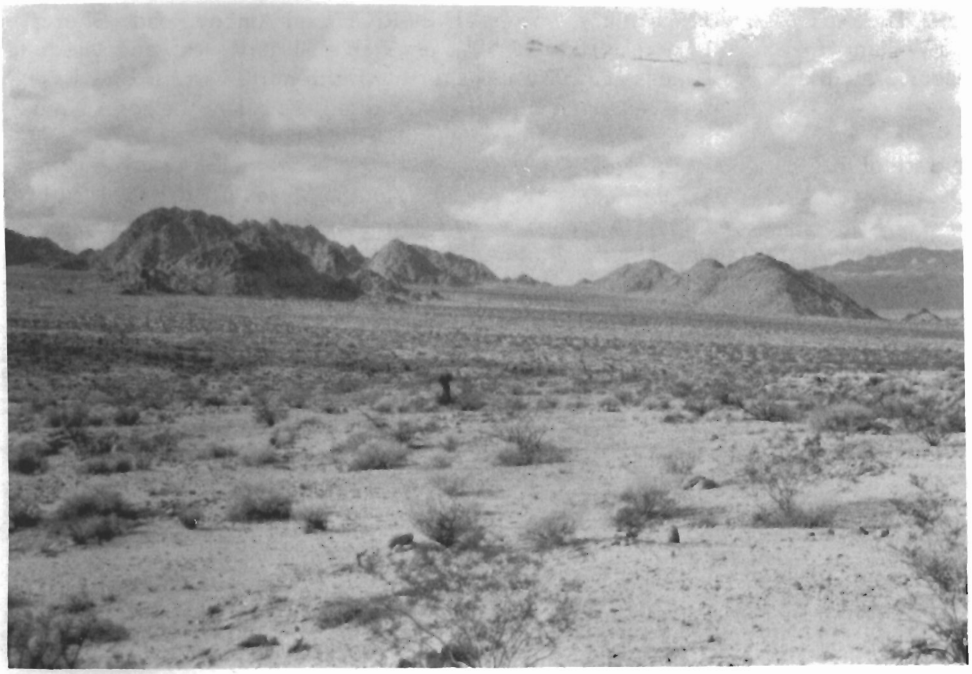

Figure 6. Sheep Hole Mountains Pass, where no fanglomerates are preserved at the base of the peaks on the granitic pediment. *R.E. Reynolds photo.*

is on the right. Look to the west at the eroded granitic pediment; there are no fanglomerates preserved at the base of the peaks, which rise steeply from the pediment (Fig. 6). Further north along the highway we can see the Granite Mountain pediment at the base of the light-colored Granite Mountains (Edinger, 1990).

We are driving downhill into the Bristol Basin (Gardner, 1980; Howard and Miller, 1992; and see Brown and Rosen, this volume). Jachens and Howard describe the structure of the Bristol/Cadiz trough (this volume).

165.2 (16.6) The south margin of Bristol Dry Lake (see Rosen, this volume; Brown and Rosen, this volume). Celestite (strontium sulphate) is found for 3 miles east and west of the highway at the south end of the playa. The rounded, nodular, potato-like masses in muds on or near the surface were exposed by dozing (Gale, 1951; Durrell, 1953).

166.2 (1.0) Salt growth on playa has caused heaving of crust on west side of road (left)

167.0 (0.8) The National Chloride salt plant (Wright and others, 1953; and see Gundry, this volume).

167.4 (0.4) The very rare mineral Antarcticite has been recovered from a trench on the west side of road (Muehle, 1970). This mineral is stable only under very restricted environmental conditions; it is difficult to find, complicated to collect, and nearly impossible to store.

169.4 (2.0) View at 10:00 of Amboy Crater and Amboy basalt flows (see Hazlett, this volume).

170.3 (0.9) View of weathered flows near the road. Brown and Rosen (this volume) note that flows occur 30' below the elevation of the playa surface.

172.4 (2.1) Slow down; sharp bend to the west is 30 mph

173.2 (0.8) STOP at intersection with National Trails Highway. TURN RIGHT onto National Trails Highway, cross over railroad tracks. The western section of the National Old Trails Highway opened between 1911 and 1914; it was the last portion of the highway that crossed the continental United States to open. The general route was originally a Mojave travel and trade trail from the Colorado River to the California coast (Smith and others, 1969).

174.0 (0.8) Amboy. In February, 1992, you could buy gas here during daylight hours.

176.5 (2.5) Pass a road on right that leads to Saltus; continue on National Trails Highway.

179.9 (3.4) STOP 6. BRISTOL BASIN. Kelbaker Road, on left, leads to Interstate 40 (no gas available). Bolo Hill, reached by a gravel road 0.3 miles east of the Kelbaker Road junction with National Old Trails Highway, marks the trace of lineaments discussed by R.E. Reynolds (this volume). The historic route that led to Needles runs to the south of Bolo Hill, to the southerly tip of the Marble Mountains, and then northeasterly to avoid this rugged range. Note volcanic tuff and flows of the Marble Mountains (north) overlying Cretaceous granites and metamorphic rocks. Early Paleozoic sandstone lies to the east.

Studies conducted over the past two years in connection with the environmental review of a proposed landfill project known as the RAIL•CYCLE Bolo Station Facility have resulted in the documentation of historic and prehistoric cultural resources in the project area. That project site stretches from Bolo Hill for a distance of four miles to the shoreline of Bristol Lake, with a width that varies from one to three miles. The proposed landfill will be located on the southern half of the project site, on the opposite side of the railroad. The area between your location and the railroad is to be maintained as a desert preserve. Based on the cultural resources studies conducted for the RAIL•CYCLE project, a brief review of the historic, ethnographic, and prehistoric background of the area is presented by Lerch (this volume).

The Hope-New Method mine is approximately 1 mile north along Kelbaker Road. Collectors have recovered a variety of uranium minerals, including rare fluoborite, at this prospect. The Iron Hat mine, at 10:0, was mined in the 1940s; the iron ore (hematite and magnetite) occurred in small, shallow lenses in Cambrian? limestone (Wright and others, 1953).

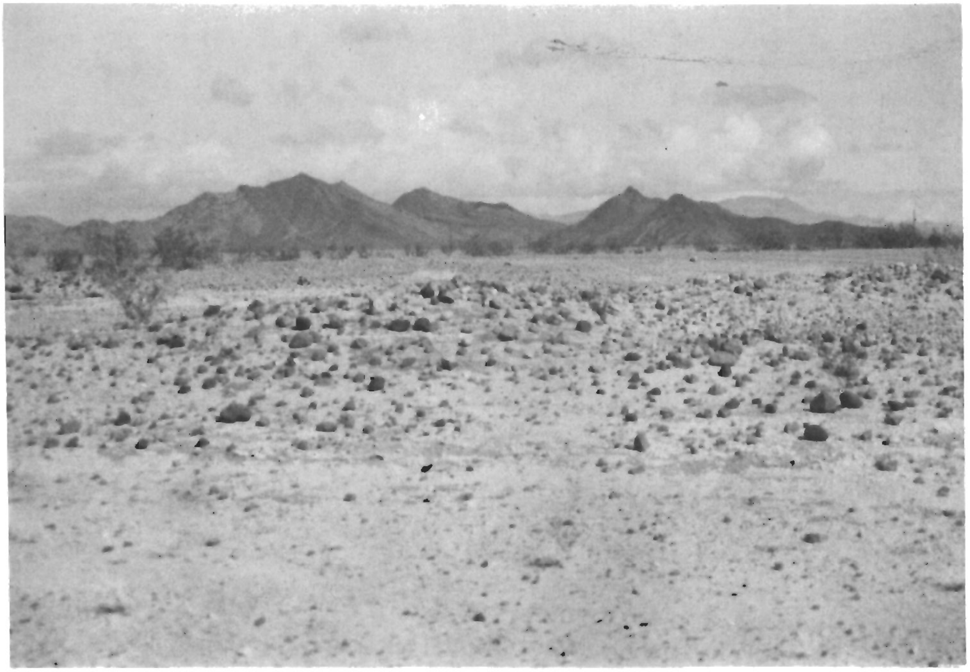

Figure 7. Pleistocene sediments near Stop 7 at Cadiz; Marble Mountains distant. *R.E. Reynolds photo.*

193.0 (0.5) STOP 7. CADIZ PLEISTOCENE SEDIMENTS at Mile Marker 215 (Fig. 7). The pipeline right of way cuts through white and tan calcium carbonate about 200' above the elevation of Bristol playa (Reynolds, 1991; and see Reynolds and others, this volume). These middle Pleistocene sediments contain pedogenic carbonates that may have been deposited as part of a distal drainage system from Lanfair and Fenner valleys. The sediments become finer as they extend westward for about 4 miles, where they end in dissected bluffs. Look northward toward sediments at Bolo Hill on east side of a lineation in eastern Cadiz Valley.

193.6 (0.6) Return to Cadiz Road.

193.8 (0.2) Continue along the south side of the AT&SF railroad tracks, past heat station. This dirt road is not maintained by the county; watch for dips and washouts.

Proceed east to Chambless on National Old Trails Highway.

185.6 (5.5) Chambless. TURN RIGHT ON CADIZ ROAD and proceed to Cadiz.

186.3 (0.7) View to the east of Pre-Cambrian sequence in contact with granitic rocks that date to 165 m.y.a. (Bishop, 1963).

188.9 (2.6) Cadiz railroad siding. Go east 1 mile on pavement. The Marble Mountains trilobite quarry is at 10:00 (Mount, 1980). The trilobite locality contains a diverse assemblage of California's oldest complex life froms, dating to earliest Cambrian times (~570 Ma). Slow to 20 mph.

190.1 (1.2) STOP at the AT&SF railroad tracks. Go over the tracks, heading southeast, and proceed southeast on dirt road to sharp bend.

192.1 (2.0) Sharp bend warns, "Caution 10 MPH".

192.3 (0.2) Cross the tracks again, to the southwest side.

192.5 (0.2) The All-American Pipeline right of way runs northwesterly (right). Proceed northwest on pipeline road to pipeline mile marker 215

194.4 (0.6) Steep-dipping Paleozoic sediments at 9:00 are on the west side of the Ship Mountains, which contain granitic rocks dating to 150 m.y.a. (Bishop, 1963).

198.3 (3.9) STOP 8. ARCHER TUFA LOCALITY. Park at the bend in Cadiz Road. The Archer tufa locality, on the south side of road, consists of clastic fluviatile sediments and root casts with an abrupt transition to massive carbonates (Fig.

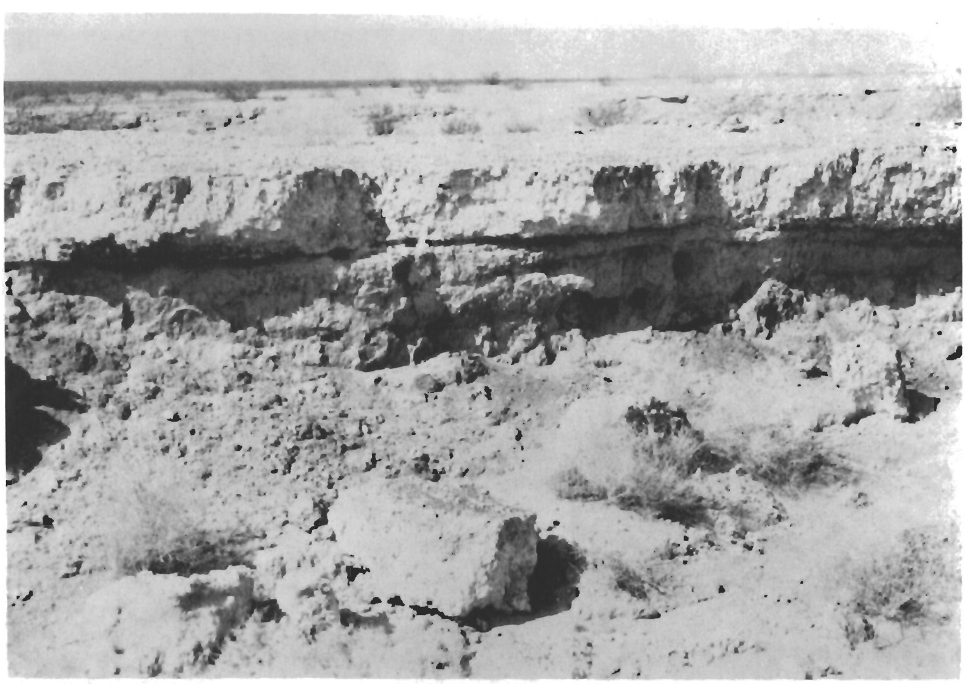

Figure 8. Fluvial sediments containing root casts in sharp contact with columnar to massive pedogenic carbonate at Archer, Stop 8. *R.E. Reynolds photo.*

8). This suggests that the carbonate layer is not pedogenic in origin. The locality is at elevation 750', more than 100' above Cadiz Lake (due south) and 150' above Bristol Lake (west). The Pleistocene fauna recovered here is described by Reynolds and others (this volume). The Cadiz dune field on the east margin of Cadiz Lake is visible to the south at 2:00.

Figure 9. Lime mill at Chubbuck. *R.E. Reynolds photo.*

200.8 (2.5) Site of Archer. Watch for dips; do not take the road heading south. The Old Woman Mountains are visible on the skyline (Howard and others, 1987; Knoll, 1985; Miller and others, 1982).

206.8 (6.0) Do not take the road south to New Frontier mine. Formerly known as the Desert Butte Group, this mine shipped a complex ore from a zone of copper, gold, silver, lead, and zinc in 1914 (Wright and others, 1953).

208.6 (1.8) The road north (do not take) goes to Skeleton Pass and Danby.

209.3 (0.7) Chubbuck. Limestone mines to the south, operated by the Chubbuck Lime Company from 1925-1948 and intermittently since (Fig. 9), explored bodies of metamorphosed limestone in the Kilbeck Hills (Wright and others, 1953).

212.8 (3.5) Fishel. The Little Piute Mountains, at the northeast end of the Old Woman Mountains, contain fossiliferous Tertiary sediments (see Reynolds and Knoll, this volume).

216.8 (4.0) At the south tip of the Old Woman Mountains, we are entering the Danby basin. The Danby and Bristol basins are at similar elevations, ~610'; Cadiz basin, due south, is ~70' lower (see Reynolds and Reynolds, this volume). Ward Valley is to the north, the Turtle Mountains are at 11:00, and the West Riverside Mountains at 12:00. The Arica Mountains lie at 1:00, the Iron Mountains at 3:00. The turnoff south goes to the Standard Salt Company (do not take).

Figure 10. Standard Salt processing plant near Milligan. *R.E. Reynolds photo.*

218.3 (1.5) Site of Milligan.

219.6 (1.3) Standard Salt processing plant (Fig. 10)(see Calzia, this volume). Three roads go east; take the middle route, not northeast or southeast routes.

221.8 (2.2) Tan silts capped with basalt gravels from the Turtle Mountains are 40 feet above the Danby playa surface, at elevation 670'.

222.6 (0.8) Power line road in Ward Valley. The road south to Iron Mountain is closed. I-40 can be reached to the north. Proceed easterly, towards Rice.

224.6 (2.0) Playa sediments here are at elevation 650'.

226.2 (1.6) The Salt Marsh railroad siding is marked by salt cedar trees.

Figure 11. Saltmarsh paleosol, Stop 9. ***R.E. Reynolds photo.***

227.4 (1.2) Mile Post 248 on gas line.

227.5 (0.1) STOP 9. SALT MARSH. The Salt Marsh site, at elevation 630', is within 20' of the current playa surface of Danby Lake. Stabilized dunes have been deposited within dissected pedogenic carbonate horizons developed on silts which contain Pleistocene vertebrate fossils (figs. 11, 12)(see Reynolds and others, this volume).

231.0 (3.5) Sablon Siding.

236.6 (5.6) TURN SOUTH, away from railroad.

237.1 (0.5) Stop at Highway 62. Note that stabilized Pleistocene dunes surround mountains at this point. The Arica Mountains, due south, have yielded gold ore since the late 1800s (Baltz, 1982); the Lum Grey and the Old Priest mines are visible south. Patton's Camp is to the west. Note the carbonate-cemented red soils on the south side of the highway. The surfaces of these deposits appear to dip into Danby basin and are covered by stabilized dunes. We are at the 850' elevation divide between Danby basin to the northwest and Rice Valley, which drains into the Colorado River, to the southeast. TURN LEFT (east) onto Highway 62. The Granite Mountains at 11:00, the Iron Mountains at 3:00.

239.1 (2.0) Danby Basin now drains internally, but subsurface cores in Danby and Cadiz lakes indicate the presence of brackish-water deposits correlated with the Bouse Formation, suggesting a trough during early Pliocene times (see Brown and Rosen, this volume).

242.1 (5.0) Rice. Emergency gas available. Do not take the Rice/Blythe Road heading south. The Turtle Mountains are at 9:00 to the north (see Hazlett, this volume).

243.3 (1.2) A historic marker is placed at the site of Rice Army Air Field. The Camp Rice Desert Training Center was established in 1942. Other camps were Young, Coxcomb, Granite, Iron Mountain, Ibis, Clipper, Pilot Knob, Laguna, Horn, Hyder, and Bouse. The operations involved 13 infantry divisions and 7 armored divisions. Training ended in the spring of 1944. The 5th Armored Division was the first unit trained at Camp Rice, and later spearheaded victories in Europe during World War II.

246.2 (2.9) Note west-dipping Tertiary volcanics to the north at 3:00.

247.7 (1.5) Pass the site of Grommet. We are at the divide between Rice Valley (to the west) and Vidal Valley, at elevation 950' (Carr, 1981).

248.1 (0.4) Cross the railroad tracks. Castle Rock is at 10:00, between the Turtles and the Mopah Range.

Figure 12. Stabilized dunes at Stop 9, Saltmarsh. ***R.E. Reynolds photo.***

Table I. Summary of features of alluvial surfaces and desert varnish in the lower Colorado Region (summarized from Tables 2.4 and 2.13, Bull 1991).

AGES OF ALLUVIAL SURFACES LOWER COLORADO RIVER REGION					DESERT PAVEMENT CHARACTERISTICS SOUTH OF WHIPPLE MOUNTAINS							
Surface	Epoch		Age Range (ka)	Basis for Age Estimate	% Surface in Pavement	% Bare Ground	Varnish	Stream channel preservation	Gravel bar preservation	Median particle size (mm)	Sorting	% >32 mm
Q4b			0	Loci of present streamflows								
Q4a	late	HOLOCENE	0.1-2	No trees in unvarnished abandoned channels	0	n/a	none	excellent to good	excellent	16	poor	22
Q3c			2-4	light brown varnish								
Q3b	middle		4-8	brown varnish; C-14 date	>90	<5	7/5YR 3/4	·none	excellent	14, 15	poor	18-20
Q3a	early		8-12	C-14 dates								
Q2c	late		12-70	C-14 & Th-230/U-234	>80	<2	7/5YR 2/3	none	none to faint	11-15	good	10-16
Q2b			70-200	Th-230/U-234; carbonate in Bk horizon	60-80	4	7/5YR 2/3	none	none to fairly good	13, 14	poor to good	6-18
Q2a	middle	PLEISTOCENE	400-730	Bishop Tuff, basalt K/A date; normal polarity	~30-40	4-20	7/5YR 2.5/2	none	none to faint	12-25	poor	8-12
Q1	early		>1200	Rare basalt; normal & reversed polarity	<1	>20-30	7/5YR 2/2	none	none	18	poor	15

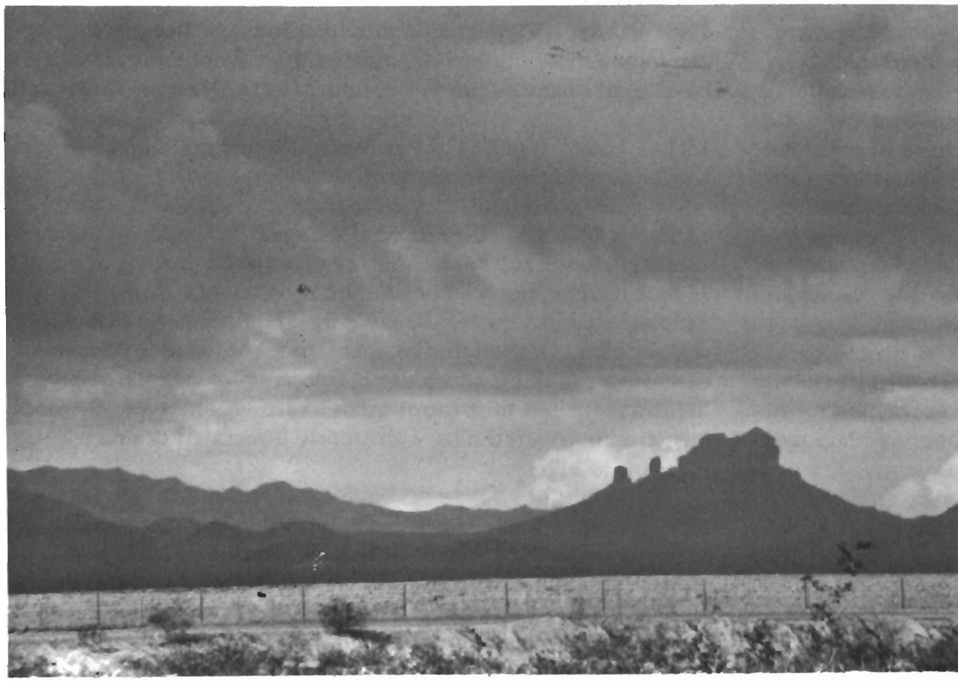

Figure 13. Volcanic buttes of the Mopah Range. *R.E. Reynolds photo.*

255.2 (7.1) Road cuts dissect calichified Pleistocene fanglomerate. Pyramid Butte is at 10:00.

256.9 (1.7) Junction of CA 62 and CA 95 at Vidal Junction. DRIVE EAST on CA 62, towards Parker, AZ. The mountain range to the northwest of the junction is the Mopah Range (Fig. 13), an eastward spur of the Turtle Mountains. The Mopah Range consists of Tertiary volcanic rocks, including the remains of volcanic vents. The neck of one such vent forms the prominent thumb-shaped peak in the middle of the range. These volcanoes mark one of the major volcanic centers active during Miocene extension, and probably was the source of volcanic flows in the western and southern Whipple Mountains. The Turtle Mountains consist of Proterozoic granite and gneiss that is similar to upper-plate basement in the Whipple Mountains. The Mopah Range and Turtle Mountains, both considered to be within the upper plate of the Whipple detachment system, are separated by a set of high-angle normal faults which may be a former headwall splay.

259.9 (1.0) The tan sediments are Quaternary Colorado River sediments.

260.9 (3.0) Savahia Peak is to the north at 8:00. Pass Chambers Well Road.

261.2 (0.3) Mile post 130

261.8 (0.6) A microwave relay station is on the south side of the road. A good overview of the Whipple Mountains can be seen to the north as you drive along this stretch of highway.

266.4 (4.6) Mile post 133

267.4 (1.0) Mile post 134. Check your odometer here; proceed 0.9 miles, just prior to Marker 135. Slow down and prepare to turn left.

268.3 (0.9) Turn left (north) off pavement onto dirt road (the turn is just past a large palo verde tree). Proceed north. This road eventually crosses the Colorado River Aqueduct and then leads to Turk Mine and Chambers Well. PLEASE DO NOT DISTURB THE DESERT PAVEMENT.

Climatic changes during the Quaternary caused profound changes in the mountains, alluvial fans, and soils of the Mojave Desert along the Colorado River (Bull, 1991). The Whipple Mountains piedmont (see Fig. 16) demonstrates many of the general characteristics of alluvial fan morphology, soil development, and chronology (Table I).

– END OF DAY 1 –

DAY 2

Colorado River Extensional Zone Segment

If you camped out at the end of Day 1, return to the large tree immediately north of Highway 62 for start of Day 2. For people joining the field trip on Day 2, go to Vidal Junction, proceed east on Highway 62. DO NOT PARK ON HIGHWAY; PARK NORTH OF HIGHWAY AT THE LARGE TREE on the side of "East Chambers Well Road" (the dirt road).

STOP 1. OVERVIEW. This spot provides an excellent view of the Whipple Mountains. In general, dark-colored rocks are Tertiary sedimentary and volcanic rocks in the upper plate of the Whipple detachment system, and light-colored rocks are mylonitic gneisses in the lower plate. The dark-light contact is the detachment fault. The broad, low dome of the Whipple Mountains, characteristic of metamorphic core complexes, is readily visible.

0.0 (0.0) Cautiously get onto Highway 62 and proceed east (left), towards Parker AZ.

0.9 (0.9) Road cut with pinkish tan sands and brown silts of the Colorado River sediments to the right (south).

3.0 (2.12) Deep gully with dark Tertiary volcanics and red sandstone of the Turk Mine and Twin Lode Mine (?) formations, which predate the Peack Springs Tuff.

3.4 (0.4) Rio Mesa Road. This is the entrance to Big River. Win cash prizes and cadillacs by listening to condo salespeople.

The low, dark hills north of the road are made up of Tertiary andesites. The volcanic flows in the Whipple Mountains, particularly on the southern flank of the range, have been subjected to a type of alteration known as potassium metasomatism. Large volumes of potassium have been added to these rocks, with removal of sodium. K_2O values in typical andesites are commonly less than 1%; K_2O values as high as 16% have been measured from these andesites. Silica and iron were also dumped into both volcanic and sedimentary rocks during this event. The resistant red sandstone ridges seen at mile 21.5 probably reflect this alteration (see Beratan, this volume).

6.5 **(3.1)** Green sediments of the Bouse Formation (Metzger, 1968) on the north side of the road are overlain by pinkish Colorado River sediments.

7.3 **(0.8)** Turn right onto Highway 72 at Earp, CA. This little town (one restaurant/gas station/store, and a post office)is named after Wyatt Earp. He hired out as a guard for stage coaches transporting gold, and eventually stayed on in this area. Turn right (east) to Parker. The white bed in the roadcut on the south side of the road before reaching the bridge is the basal marl of the Bouse Formation (more about this later on). Note the excellent large-scale cross-beds in the gravels below the Bouse. Dugout cabins on right (west) were excavated below marl of the Bouse Formation at intersection with Rio Vista, in 0.3 miles (Fig. 14).

7.7 **(0.4)** Cross Colorado River into Arizona and the metropolis of Parker. Gas up and get foodstuffs for a picnic lunch.

8.6 **(0.9)** Stop, TURN NORTH (left) onto AZ Highway 95.

Figure 14. Shelters dug out under the basal marl of the Bouse Formation. ***R.E. Reynolds photo.***

11.5 **(2.9)** View at 2:00 of white Bouse sediments ringing canyon. The white bed exposed to the east of the road is the basal marl of the Bouse Formation. Cross Osborne Wash.

13.1 **(1.6)** Traffic light at Beacon/Riverside Roads. Continue straight across.

14.1 **(1.0)** Rio Vista Road, Cienega Springs.

14.3 **(0.2)** The hills beside the road contain typical exposures of Proterozoic basement. The basement in this region is highly shattered and sheared, both due to Miocene extensional faulting and older (Cretaceous and Proterozoic) deformation. As a result, it erodes readily, and typically forms low, rubble-covered hills with poorly integrated drainages—not pleasant to hike on.

15.7 **(1.4)** The red-colored ridge straight ahead consists of Miocene sedimentary rocks, predominantly hematite- and quartz-cemented sandstone and conglomerate deposited in alluvial fan-fluvial settings. These redbeds are part of the Copper Basin Formation (Teel and Frost, 1982), which we will drive through.

16.3 **(0.6)** La Paz County Park

18.8 **(2.5)** Another view of the coarse, red Tertiary sediments of the Copper Basin Formation.

19.3 **(0.5)** The light-colored rock exposed in the roadcut is the 18.5 Ma Peach Springs Tuff. This distinctive ignimbrite unit is found throughout the eastern Mojave Desert, and forms an important marker bed. The Whipple Mountains lie at the distal margin of the unit. Exposures are lens-shaped; the hot ash followed topographic lows, and accumulated in valleys. In the Parker Dam area, these lenses are located just below the Gene Canyon-Copper Basin unconformity.

19.7 **(0.4)** Entrance to Buckskin Mountain State Park on left. This is a very nice little campground, with tent spaces along the river (away from RV's) and very friendly rangers. Recommended to those who prefer camp sites with showers.

21.1 **(1.4)** River Island Park. The distinct strata in the high skyline to the east (right) are discussed below (MP 23.2).

22.4 **(1.3)** Volcanic breccias are exposed to the right in road cut.

22.9 **(0.5)** The road goes by a cliff which forms a dramatic exposure of brick-red conglomerates and sandstones of the Copper Basin Formation. These rocks are correlative with those at Stop 3, but are significantly coarser-grained, consistent with a source area to the east-southeast. Clast types include Mesozoic(?) metasedimentary and metavolcanic rocks; such rocks are found in the Buckskin Mountains but are absent from the Whipple Mountains.

23.2 (0.3) The strata underlying the cliff-forming redbeds, part of the Gene Canyon Formation, are dominated by monolithologic breccia beds. Boulder-size material is common. The beds generally are clast-supported, and the matrix consists of the same material as the clasts. The beds are unsorted and disorganized. These breccia beds are interpreted as rock avalanche deposits. The breccia beds on this side of the river are much coarser, thicker, and more abundant than those on the California site, consistent with a source to the east-southeast.

The flat-lying strata that form the high skyline to the east (right) of the road are discussed below (MP 23.2) are dominantly olivine basalts of Late Miocene age that unconformably overlie the Gene Canyon and Copper Basin formations. These volcanic rocks have not experienced the intense alteration that affected the older andesites.

23.8 (0.8) DO NOT TAKE the viewpoint exit. TURN LEFT off Highway 95 onto the road to Parker Dam.

24.6 (0.8) Cross over Parker Dam. (Be careful—the road across the dam is very narrow). Arizona Highway 95 turns into California Highway 62 at the California state line.

24.8 (0.2) STOP 2. GRANITE PORPHYRY OF PARKER DAM. PARK at lot on west side of dam. The roadcut bordering the parking lot on the west side of the dam is the type area for the Granite Porphyry of Parker Dam (better known as "Fred" to geologists working in the Colorado River extensional corridor.) This badly-sheared granite porphyry is part of a distinctive suite of ~1.4 billion year old anorogenic granites. Varieties of this rock are one of the most common rock types within the basement throughout the region.

25.2 (0.4) TURN WEST (right) just before the small community of Parker Dam, before Gene Wash Reservoir Road, onto a road leading to the Metropolitan Water District's Gene Pumping Plant. There is a sign for "Black Meadow Landing Resort" at the turnoff.

25.7 (0.5) STOP 3: PARKER DAM SECTION. Park in the large turnout on the south (left) side of the road. **Watch out for oncoming traffic** — cars travel much too fast on this road. This property belongs to the Metropolitan Water District; permission to visit the exposure can be obtained at the Gene Pumping Plant, about a mile further up this road. The base of the Parker Dam section is just across the road from the turnout. This is one of the most complete and least structurally disrupted Miocene sections anywhere within the Colorado River extensional corridor. Walk upsection along the road.

A basement assemblage dominated by the Granite Porphyry of Parker dam is nonconformably overlain by sandstones, conglomerates, monolithologic breccias, and rare limestones of the Gene Canyon Formation. Note the textural variability of the unit, and the lack of volcanic rocks as clasts. Stratal dips within the Gene Canyon Formation range from approximately 75 degrees near the base to about 40 degrees at the top. A volcanic flow is exposed in the roadcut at the turnoff in the center of the hairpin curve near the top of the hill. The upper member of the Gene Canyon Formation contains some thin andesite flows. This unit is capped by a lens of the 18.5 Ma Peach Springs Tuff, and is unconformably overlain by the Copper Basin Formation.

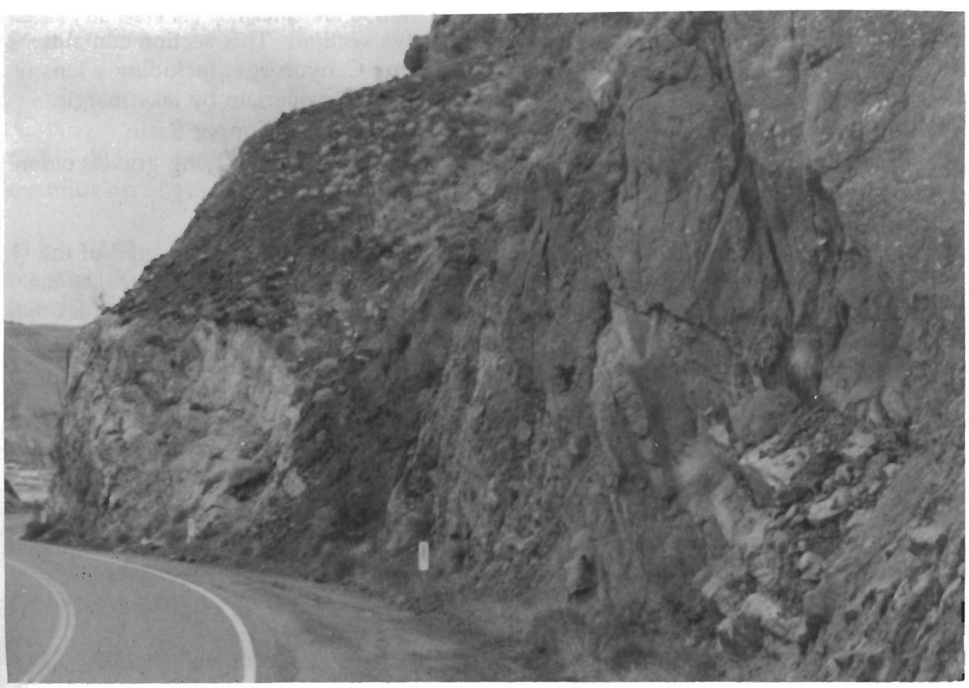

Figure 15. Unconformity between the Gene Canyon and Copper Basin formations. *R.E. Reynolds photo.*

The Gene Canyon—Copper Basin unconformity lies at the top of the hill, just past the hairpin curve. The power line at the top of the ridge sits on a pale pink exposure of the Peach Springs Tuff. The unconformity is well-exposed to the east (left), across the stream. The unconformity where it crosses the road is marked by a complex lens of deformed sedimentary and volcanic rocks. Just beyond the ridge, a pull-off leading to the Gene Reservoir Dam contains a good exposure of typical Copper Basin Formation redbeds.

Return to the vehicles, and proceed to the east on Gene Wash Reservoir Road, back to Highway 62 and the town of Parker Dam.

26.2 (0.5) Stop, TURN SOUTH (right) on CA Hwy 62/AZ 95 and go south.

26.8 (0.6) The coarse-grained deposits viewed at mile 23.3 can be seen eastward across the river. Those rocks are part of the same fault block as the Parker Dam section.

27.3 (0.5) The Gene Canyon—Copper Basin unconformity is well exposed in the cliff on the right (Fig. 15). **Be careful driving along this stretch of road.** People drive stupidly through these curves; head-on collisions and cars in the river are fairly common.

27.5 (0.2) The unconformable relationship between the tilted Gene Canyon and Copper Basin formations with the overlying, flat-lying olivine basalts can be seen eastward (to the left) across the river.

28.9 (1.4) Cable Car day use area.

29.4 (0.5) Quail Hollow day use area.

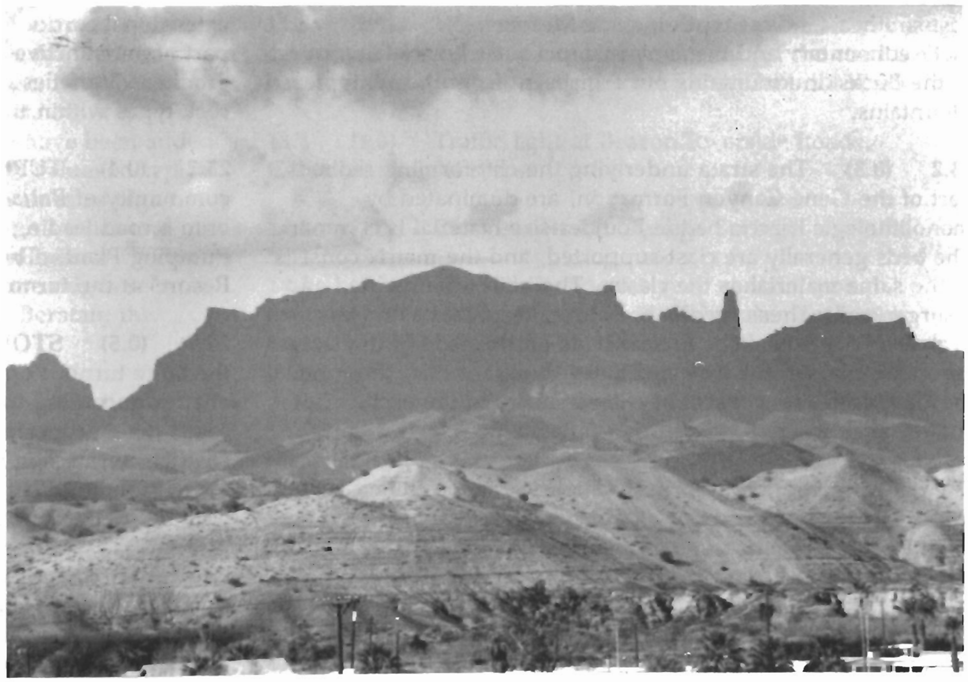

Figure 16. View north of Whipple Mountains. *R.E. Reynolds photo.*

30.7 (1.3) The sand dunes on the right are eolian dunes derived from reworking of underlying sediment. The sediment is probably part of the Bouse Formation, but may also in part be derived from Quaternary sediment deposited by the Colorado River.

30.9 (0.2) Echo Lodge Resort. The exposure on the left, between the road and the river, contains Gene Canyon Formation sedimentary and volcanic rocks unconformably overlain by Copper Basin Formation redbeds. The Whipple Mountains (Fig. 16) are north.

31.3 (0.4) Transition from red volcaniclastic sediments to mostly Proterozoic granite and gneiss; both the sediments and the crystalline rocks are within the upper plate of the Whipple Detachment, and are separated by a high-angle normal fault which presumably soles into the Whipple Detachment Fault.

32.9 (0.7) Bullfrog Day Use Area.

33.2 (1.2) The hills to the north (right) consist of Quaternary gravels deposited by the Colorado River on metamorphic rocks.

34.9 (2.7) Cross Roads Day Use Area and historic monument. The jeep trail that begins here heads up Bowman's Wash to the feather edge of the upper plate. The detachment fault can be easily reached from Bennett Wash, a left fork off of the Bowman's Wash Road. Exposures near the bulldozed rubble pile at the bend in the Bennett's Wash road include limestone and mudstone that were deposited in the lake that formed during the tilting episode that created the Gene Canyon—Copper Basin unconformity.

35.9 (1.0) Bass Point day use area. Metamorphics in the upper plate are visible at 12:00. More Quaternary gravels are to the west (right) at 3:00. **Be careful going around this curve.**

36.7 (0.8) The cliff to the north (right) exposes the unconformity between the tilted mid-Miocene Copper Basin Formation strata and flat-lying strata assigned to the late Miocene Osborne Wash Formation along the next 0.4 miles.

37.3 (1.1) The Bouse Formation can be seen on the north side of road for the next 0.4 miles.

39.0 (1.7) Lake Moovalya section. This section contains strata of middle Miocene Gene Canyon age, including a lens of Peach Springs Tuff, unconformably overlain by lake margin deposits belonging to the mid Miocene Copper Basin Formation. These rocks are overlain by flat-lying gravels older than the Bouse Formation.

40.5 (1.3) Note the distinctive white band of marl of the basal Bouse Formation. The marl is visible from here into the town of Earp. In this area, the marl is underlain by a thick section of flat-bedded gravels.

41.6 (1.1) Earp again. Turn south on Highway 72 and cross the Colorado River into Parker, Arizona.

41.7 (0.1) The road cuts through strata belonging to the Osborne Wash Formation, and the overlying yellow sandstone and the basal marl of the Bouse Formation. The marl is particularly distinctive here, forming a readily recognizable thin white band.

41.8 (0.1) Rio Vista.

42.0 (0.2) We are crossing the Colorado River.

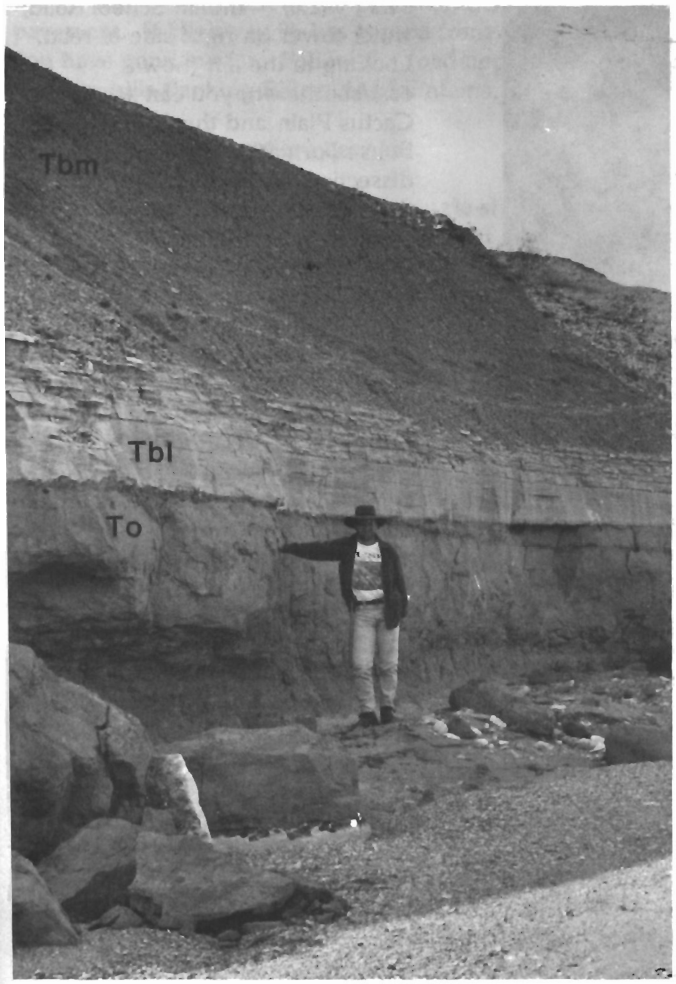

Figure 17. Exposures at Stop 4, Jason Mayfield for scale. To = fanglomerate of "Osborne Wash strata" (Buising, 1990); Tbl = basal white carbonate of Bouse Fm; Tbm = green mudstone of Bouse Fm. Note lag conglomerate at base of limestone. *A. Buising photo.*

42.9 **(0.9)** Stop at Highway 95, but proceed across and continue on Highway 72.

43.9 **(1.0)** Pass a Chevrolet dealership on the left before reaching the turn onto Mohave Road. Continue on highway, but note this turnoff for future use after lunch.

44.5 **(0.6)** TURN LEFT (east) on Shea Road.

44.7 **(0.2)** Stop at railroad tracks; proceed along Shea Road.

45.4 **(0.7)** The low plain to the right (south), the Cactus Plain, is covered by semi-stabilized eolian dunes which overly the Bouse Formation. The sand is fine-grained, and was probably derived by reworking of the Bouse Formation.

47.2 **(1.7)** Curve northeast; Black Peak at 3:00. Note the dark varnished basalt debris covering steep slopes, indicating that the slope surfaces have been stable for a relatively long time.

49.6 **(2.4)** Drop off terrace. Cross, but do not drive up the sandy wash.

50.0 **(0.3)** **STOP 4: OSBORNE WASH.** Park beside the road or on the gravel area north of the wash. The western half of Black Peak is part of the Colorado River Indian Tribes reservation. This mountain is sacred to them—please do not hike on tribal land.

The sedimentary and volcanic rocks exposed in Osborne Wash and up the side of Black Peak were deposited in the Buckskin Basin. This basin was separated from the Whipple Basin by the Billy Mack Mountain Fault and a stable basement high in the Buckskin Mountains. The mid Miocene Black Peak section includes red, coarse-grained sandstones and conglomerates that are time-correlative with the Copper Basin Formation, and which contain large clasts of the Peach Springs Tuff. The redbeds are conformably overlain by olivine basalts which make up the top of Black Peak. The basalt stack contains an angular unconformity within it. A buttress unconformity separates the flat-lying Pliocene Bouse Formation and immediately underlying fanglomerates from the older units.

Walk west along Shea Road to the point where it enters Osborne Wash. Walk southeast along the wash, and revel in the excellent outcrops in the west wall of Osborne Wash and in the sides of smaller tributary drainages entering Osborne Wash from the southwest (Fig. 17). The buff to pale pink unit is an Upper Miocene fanglomerate (Osborne Wash strata of Buising, 1990; Osborne Wash Formation of Davis and others, 1980); this is capped by 1-2 m of yellow sandstone. The yellow sandstone locally interfingers with the bright white, well-bedded carbonate marking the base of the Bouse Formation. The carbonate is in turn overlain by as much as 10 m of poorly exposed green mud of the Bouse Formation basin fill association (Buising, 1988, 1990; Smith, 1960, 1970; Winterer, 1976).

Things to notice in these exposures include:
In the fanglomerate: well-developed reverse and reverse-to-normal grading in gravel and cobble conglomerates; clast assemblage overwhelmingly dominated by basalt clasts derived from flows exposed on Osborne Ridge and Black Peak.

In the yellow sandstone: beautiful medium-scale trough cross-beds (notice especially the contrast in depositional style between the debris flow-dominated fanglomerate and this dilute flow unit!)(Fig. 18); also note the local abundance of ostracodes and cm-scale gastropods.

In the white carbonate: locally abundant ostracodes and cm-scale gastropods; well-developed bedding, locally draped over cobble- and boulder-grade clasts in a basal lag horizon; also note stromatolitic or biohermal structures developed on some of the lag clasts.

50.1 **(0.1)** Return to Shea Road; go west.

55.4 **(5.3)** Stop, cross the railroad tracks.


Figure 18. Trough cross-bedding in yellow sandstone at Stop 4. *A. Buising photo.*

55.6 (0.2) TURN RIGHT (north) on AZ 95.

56.1 (0.5) TURN LEFT (roughly west) onto Mohave Road; a sign indicates Poston, Ehrenburg, Blythe. Continue on Mohave Road.

57.5 (1.4) First Avenue.

57.9 (0.4) TURN LEFT into Colorado River Indian Tribes Museum, before 2nd Avenue.

58.0 (0.1) Return to AZ 95; turn right.

58.7 (0.7) Mohave Road curves to the left (southwest). To the left is Mesquite Mountain, buttressed by low hills of Bouse Formation capped by Tertiary-Quaternary Colorado River gravels and Quaternary alluvial/pediment gravels. Straight ahead (west), across the modern Colorado River floodplain, you can see the Riverside Mountains on the skyline. The low hills surrounding the crystalline massif of the Riversides also consist of Bouse Formation capped by Quaternary gravels.

62.7 (4.0) Mohave Road turns south; continue along it.

63.9 (1.2) Burns Road; continue south on Mohave Road.

64.9 (1.0) Indian School Road; water tower on right side of road. Looking to the left (now east/southeast), you can see the Cactus Plain and the underlying Bouse Formation exposed by deep dissection. Erosion allows us to see the buttressing relationship between the flat-lying Bouse Formation basin fill strata and the crystalline topography at the base of the Plomosa Mountains (behind Mesquite Mountain). Continue on Mohave Road.

66.9 (2.0) TURN RIGHT (west) onto Agnes Wilson Road, marked by a tractor dealership. If you pass the tractor dealership, you have missed the turn.

69.5 (2.6) Cross the Colorado River into California, and go along the Colorado River flood plain.

71.2 (1.7) Cut through Quaternary Colorado River sediments.

71.5 (0.3) Pass a turn out to which we will eventually return.

72.3 (0.8) Agnes Wilson Road stops at California Highway 95. Stop, turn right (north).

74.8 (2.5) The San Bernardino County line.

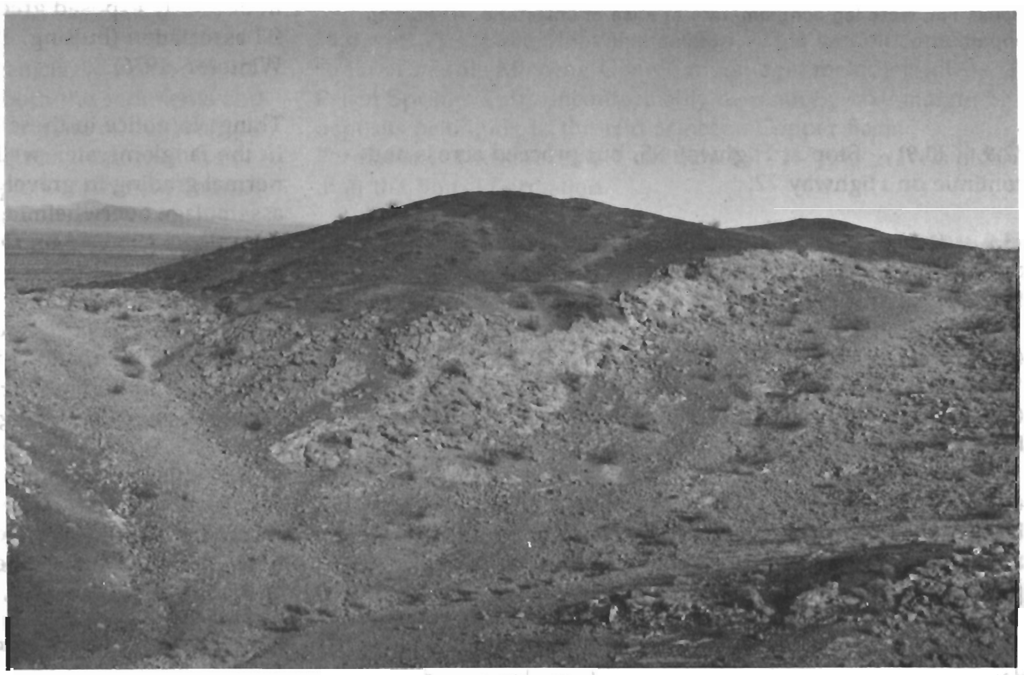
Figure 19. Overview at Stop 5. Bouse Fm. shoreline deposits; tufa (light gray) encursting basement gneisses (dark gray). *A. Buising photo.*

75.3 (0.5) Turn left at the pole line, before a bend in the pavement. If Highway 95 has turned from north to N45W, you have gone too far. Follow the road approximately west into low hills flanking Riverside Mountains.

75.7 (0.4) Intersection; bear left.

76.2 (0.5) This is a complex, triangle-shaped junction of narrow, ungraded dirt roads. Take the left (south-southwest) fork. Follow this road, heading toward the Riversides, up and down across active drainages.

76.3 (0.1) Cross wash.

76.7 (0.4) STOP 5: BOUSE FORMATION SHORELINE TUFA. Park the vehicles to one side of the track as space permits. Please do not drive on desert pavement surfaces!

Hilly outcrops to the left (south) are gneissic basement rocks (black and dark red-brown) with a discontinuous veneer of Bouse Formation shoreline tufa (gray) (Fig. 19). Tufa forms a rind on basement exposures; locally, the base of the tufa includes a monolithologic breccia of angular basement clasts cemented by gray carbonate. More than one layer of tufa is present in some places. If you have time to wander around in this vicinity, exploring the shoreline, you will find exposures of clastic limestone and coarse terrigenous-clastic strata interbedded with the tufa.

The present-day upslope limit (elevation 640'+) of the tufa (apparent Bouse high-water mark) is probably the result of erosion; discontinuous outcrops of tufa upslope from this elevation demonstrate that the shoreline was once significantly higher that it now is.

Turn vehicles around and return to complex, triangle junction.

77.1 (0.4) At complex junction, take the road to the west (left).

78.1 (1.0) Mine road intersection.

78.3 (0.2) STOP 6: OVERVIEW OF BOUSE FORMATION BASIN FILL STRATA. The low, variegated hills to the right are terrigenous-clastic strata of the Bouse basin fill association. Looking to the left, you can see the Bouse lapping up and buttressing against the irregular pre-Bouse topography of the Riversides. At this stop, the Bouse is capped by a 1m+ thick, red-brown soil which contains pedogenic carbonates; the soil is truncated by a well-developed desert pavement surface. Turn the vehicles around here and drive back. Avoid driving on desert pavement surfaces.

Retrace to complex junction and thence to Highway 95.

79.6 (1.2) Complex junction; bear left.

80.5 (0.9) At Highway 95, stop, turn south to Agnes Wilson Road.

83.5 (3.0) TURN RIGHT (west) off Highway 95 onto Wilson Road (2 lane graded dirt).

83.8 (0.3) Road divides; take the left fork.

83.9 (0.1) TURN LEFT on small, faint, ungraded trail.

84.6 (0.7) STOP 7. BOUSE FORMATION SHORELINE COMPLEX; GRAVELS OF AMBIGUOUS AFFINITY. Take left fork to triangle turn-around; park as convenient, but please stay off desert pavement surfaces. Walk to edge of large, roughly east-west trending drainage south of turn-around.

The basement rock north of the drainage is the same gneiss as that at Stop 5; south of the drainage, the basement is Paleozoic carbonate. From your vantage point at the edge of the drainage, you should be able to look across (roughly south) and trace the Bouse shoreline (tufa + coarse terrigenous clastic strata) around the drainage. Notice that the layered units dip markedly into the drainage. This is original (depositional) down-to-basin dip. Looking down, you will see erosional remnants of green and buff Bouse basin fill strata in place in the drainage. This drainage clearly predates Bouse deposition: the Riversides, like Mesquite Mountain and other nearby ranges, were deeply dissected prior to Bouse time.

Follow the road down into the drainage and – carefully! – across the old wooden bridge to the Calzona mine.

The steep narrow part of this drainage is cut into sandstone (± calcarenite) and conglomerate of the Bouse Formation basin fill association. After crossing the bridge, walk uphill (following the path) until you can peer down into the drainage without falling in. If you walk far enough, you will be able to see the coarse terrigenous-clastic units overlying the tufa in the bottom of the drainage. You can climb down into the drainage and take a closer look at the sandstones and conglomerates. Notice that the conglomerate clast assemblage includes locally derived quartzofeldspathic gneiss and carbonate as well as tufa clasts.

Walk back out the way you came in. One of the most intriguing things about the outcrops in this drainage is the difficulty in establishing the stratigraphic affinity (affinities?) of the various gravels/conglomerates exposed here. Gravels overlie tufa; these are probably Bouse. Gravels are overlain by tufa; these may be Bouse Formation or Osborne Wash. Less consolidated gravels overlie the Bouse gravels of the "narrows" – are these post-Bouse, or just a younger generation of Bouse gravels??

When you have finished here, drive out the same way you came in, and rejoin CA 95.

85.6 (1.0) TURN NORTH on Highway 95 to Vidal Junction. (Or turn south, if you want to go to Blythe, but you'll miss the last stop).

86.3 (0.8) STOP 8. QUATERNARY COLORADO RIVER SEDIMENTS. Johnson and Miller (1980) provide a sequence of deposition that describes episodes of post-Bouse Formation deposition, downcutting, and backfilling during the Pleistocene. Agenbroad and Reynolds (this volume) discuss Irvingtonian LMA mammoth fossils which indicate that some of these sediments were deposited in middle Pleistocene times.

Retrace your route to Highway 95.

87.1 (0.8) Junction of Highway 95. Now you can turn south to Blythe without missing anything.

89.6 (2.5) San Bernardino/Riverside county line.

93.2 (3.6) Railroad tracks at Vidal.

99.4 (6.2) Vidal Junction and intersection of CA Highway 95 and CA highway 62. Fill up with gas.

Highway 95 leads north to Needles and I-40. Highway 62 leads west along our Day 1 route to Twentynine Palms. At Granite Pass, Highway 171 turns south and intersects with Interstate 10 to Indio.

-- END DAY 2 --

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Sediments in Yucca Valley Adjacent to the Pinto Mountain Fault

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ABSTRACT

Three sedimentary units in Western Yucca Valley were deposited in late Miocene, Pliocene, and Pleistocene times. The age of these sediments, although poorly constrained, suggest timing and rates of movement along the Pinto Mountain Fault.

INTRODUCTION

The Pinto Mountain Fault is the northernmost of the east-trending faults which characterize the eastern Transverse Ranges in southern California. Dibblee (1967, 1975) demonstrated that Mesozoic plutonic contacts previously crossing the central portion of the Pinto Mountain Fault are offset by about 10 miles (16 km) of left-slip. These displacements are suspected to be the cause or result of deformations of the San Andreas Fault in the San Gorgonio Pass area (Matti and others, 1985).

The Morongo Valley Fault is a normal fault diverging from the western portion of the Pinto Mountain Fault. This fault bounds the southern margin of Morongo Valley, which is a graben structure between the Morongo Valley and Pinto Mountain faults (Fig. 1).

This paper addresses late Cenozoic sedimentary rocks near the eastern intersection of the Pinto Mountain and Morongo Valley faults. These sedimentary rocks record the local depositional environments created by initial displacements on Pinto Mountain Fault and subsequent uplift of the San Bernardino Mountains.

GEOMORPHOLOGY

Morongo Valley and Yucca Valley are interconnected, northeasterly-trending troughs along the Pinto Mountain Fault which have been partially filled with sediment. The two valleys are separated by a narrow drainage divide, south of which the geomorphology is characterized by a nearly horizontal, deeply incised surface on late Cenozoic sedimentary rocks.

The geomorphic surfaces on the late Cenozoic sedimentary rocks have a moderately developed desert pavement composed predominantly of quartzite clasts and basalt clasts. The ages of these geomorphic surfaces are estimated to be from late Pliocene to early Pleistocene for the quartzite clast surface; and middle Pleistocene for the basalt clast surface.

The deeply-incised geomorphic surface is at a constant elevation throughout the central portion of the study area. A similar geomorphic surface has developed on quartzite clast fanglomerate in Morongo Valley, but it is about 800 feet lower in elevation than the surface in Yucca Valley due to normal displacements on the Morongo Valley Fault.

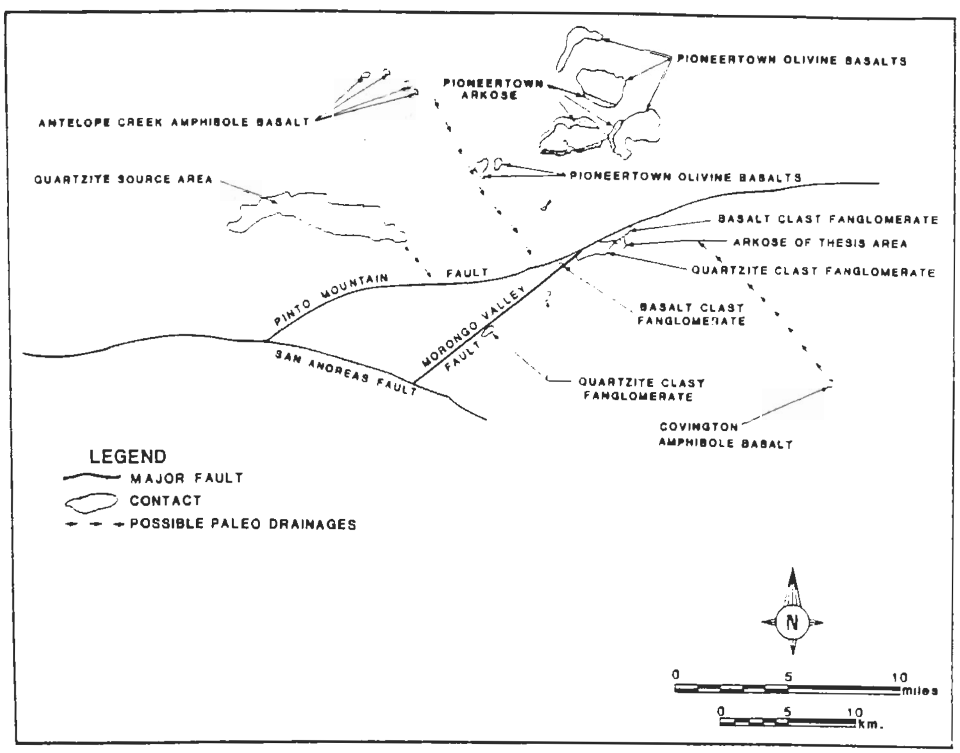

Figure 1. Faults and rock units in the Yucca Valley vicinity.

LATE CENOZOIC SEDIMENTARY ROCKS

Late Cenozoic sedimentary rocks are exposed south of Twentynine Palms Highway between Yucca Valley and Morongo Valley. The exposed sediments are separated into three distinct formations based on differences in lithology, structure, and clast types. The oldest formation is a silty arkosic sandstone (Arkose). It is unconformably overlain by a fanglomerate characterized by quartzite clasts (Quartzite Clast

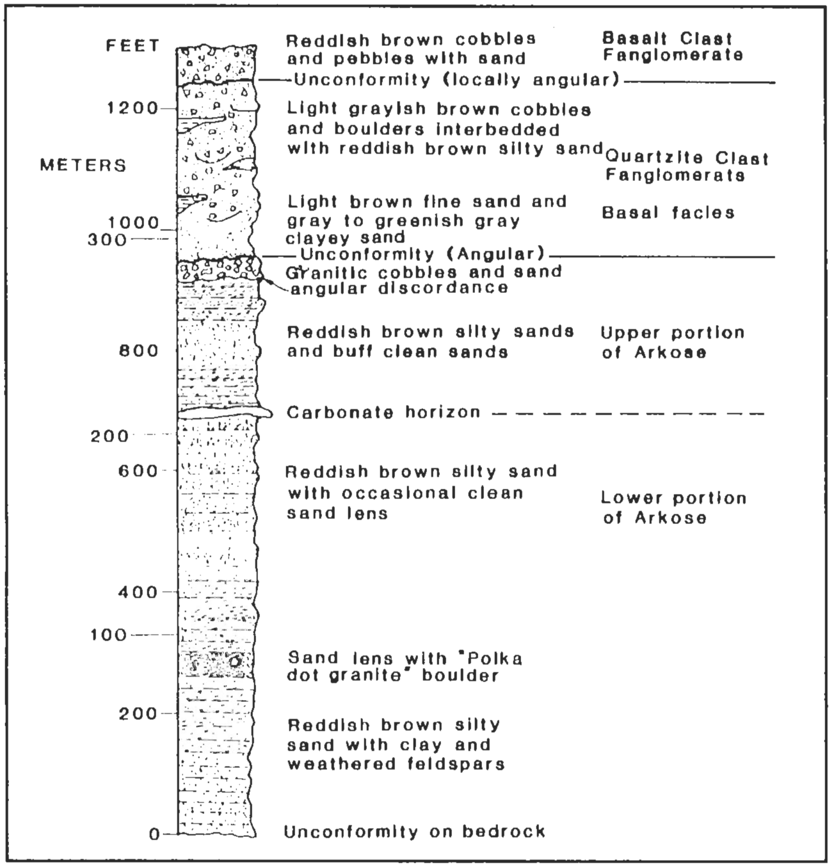

Figure 2. Generalized composite stratigraphic column of late Cenozoic sedimentary rocks.

Fanglomerate). The youngest Late Cenozoic deposit is a fanglomerate characterized by basalt clasts (Basalt Clast Fanglomerate) which unconformably overlies the arkose and a distal or lower facies of the Quartzite Clast Fanglomerate. All unconformable contacts are angular; the upper beds dip more gently than the underlying beds. Refer to Figure 2 for a generalized composite section of these units.

Arkose

The Arkose consists of lower silty sandstones and upper cleaner sandstone beds separated by a pedogenic carbonate horizon. The maximum exposed stratigraphic thickness is estimated to be 300 meters.

The lower ~2/3 of the Arkose is composed of well indurated, reddish-brown to olive-brown, poorly sorted silty sandstones that contain minor pebbles and biotite flakes. Beds are from 0.5 to 5 m thick and are not distinct, giving a massive appearance to most exposures. A two foot thick pedogenic carbonate horizon marks a change from the poorly sorted lower portion of the Arkose to the better sorted upper portion of the Arkose.

The upper ~1/3 of the section, above the carbonate horizon, is similar in composition to the lower portion, but it is sorted into well defined beds. These light brown to buff, clean sandstones and olive-brown, silty sandstones have bedding thicknesses from 10 cm to 2 m, respectively.

A massive conglomerate is slightly unconformable with underlying beds at the top of the arkosic section. This deposit is in a tight fold with a core of metamorphic rock immediately below near horizontal Basalt Clast Fanglomerate. The clast composition appears to be locally derived from metamorphic terrain with the exception of subangular granitic cobbles suspected to be from a source area approximately 11 miles west of the study area.

The Arkose in the study area is equivalent to the eastern facies of the Santa Ana Sandstone (Lowman, 1989; Sadler, 1982; and see Reynolds, this volume) exposed in Pioneertown. The arkose in Pioneertown is capped by and interbedded with olivine basalt flows. Peterson (1976) has dated the basal portion of these flows by K-Ar methods as being 7.3 my old.

Quartzite Clast Fanglomerate

The Quartzite Clast Fanglomerate grades upward from thick bedded combinations of greenish-gray claystone, siltstone and fine-grained sandstone (basal facies) to a reddish-brown arkosic sandstone, cobble, and boulder fanglomerate. Bedding characteristics are variable and abrupt, with most beds ranging from 1 to 3 meters in thickness. The estimated thickness of this well consolidated section is 90 meters.

Although quartzite clasts are distinctive in this formation, granitic and gneissic cobbles and boulders predominate (estimated 70% of clast types). The quartzite clasts are white to gray and vitreous with thin opaque laminae or alternating dark gray and light gray banding from 2 to 5 cm thick.

Basalt Clast Fanglomerate

The Basalt Clast Fanglomerate is composed predominantly of cobbles and pebbles with a reddish brown matrix of sand, silt, and clay. There is generally clast-to-clast support and bedding is poorly developed. The formation is poorly to moderately cemented. The maximum exposed stratigraphic thickness is 9 to 12 meters.

Basalt cobbles containing subhedral black glassy megacrysts up to 12mm in length and olivine basalt clasts distinguish this formation. Other clast types are gneissic cobbles and granitic boulders. The morphology and color of the basalt clasts is distinctive. Basalt clasts containing black glassy megacrysts have a 2 to 3 mm thick medium gray rind on subrounded, slightly flattened clasts with a pitted surface. Olivine basalt clasts are dark gray and rounded with smooth or vesicular surfaces. The age of the extrusive source areas for the basalt clasts is 6 to 9 Ma (Grimes, 1986; Neville and others, 1983).

DISCUSSION

The age of the initial displacements on the Pinto Mountain Fault is constrained by the Tertiary arkose south of the fault. The Arkose beds are conformable in the exposed section until coarse grained sediments at the top of the section indicate a significant change in the depositional environment which I attribute to initial displacements on the Pinto Mountain Fault. The age of initial displacements on the Pinto Mountain fault is likely near the age of the basal basalt flows which are interbedded with the upper portion of the arkose in the Pioneertown area (about 7 mya). Latest Miocene and early Pliocene displacements can not be ruled out due to an imprecise age for the uppermost Arkose in the study area.

Significant uplift of the San Bernardino Mountains resulted in deposition of the Quartzite Clast Fanglomerate. The uplift appears to have been active during the Pliocene.

The Basalt Clast Fanglomerate appears to indicate that source areas within the central San Bernardino Mountains had been displaced prior to the middle Pleistocene.

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The Tertiary Pioneertown Sequence

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ABSTRACT

An isolated section of Tertiary sediments is located north of Pioneertown in the eastern San Bernardino Mountains. The age is constrained by Clarendonian LMA (11.5 Ma) fossil vertebrates in the lower half of the section and by basalts as old as 9.3±0.7 Ma in its upper part. These age constraints help demonstrate that the Pioneertown Sequence is older than and separable from the 7-4 Ma Santa Ana Sandstone to the west and the 3 Ma Old Woman Sandstone to the northwest.

BACKGROUND

Tertiary sediments exposed north of Pioneertown are the only Tertiary sediments in the far eastern San Bernardino Mountains and in the Little San Bernardino Mountains. The Pioneertown locality is in the Transverse Range Province, 12 miles north of the Pinto Mountain Fault.

Tertiary sediments in and around the San Bernardino Mountains record periods of uplift and deposition. In the western San Bernardino Mountains at Cajon Pass (Woodburne and Golz, 1972; Woodburne, 1991a), the age of the late middle Miocene Cajon Formation (Meisling and Weldon, 1989) is well constrained, as is the span of time represented by the Crowder Formation (Reynolds, 1991a). Grimes (1986 and this volume) notes that sediments south of the Morongo Valley Fault in western Yucca Valley are similar in appearance to the sediments at Pioneertown. Dibblee (1967) referred the sediments at Pioneertown to the Old Woman Sandstone, 20 miles northwest of Pioneertown and south of Lucerne Valley. May and Repenning (1982), however, describe late Pliocene, Blancan Land Mammal Age (LMA) vertebrate fossils from the Old Woman Sandstone, suggesting that it is 3.2-2.5 m.y. in age and thus much younger than the youngest date on the Pioneertown basalts (6.86±-.25 Ma, Lowman, 1989).

Sadler (1982, 1985), Strathouse (1982), Neville (1983), and Lowman (1989) recognized this age discrepancy and referred the sediments at Pioneertown to the Santa Ana Sandstone. The type section of the Santa Ana Sandstone is 12 miles to the west at Barton Flats, along the Santa Ana River drainage.

Lowman (1989) divides the sediments at Pioneertown into a lower brown member containing fragments of gneiss, and an upper white arkosic member. This division is similar to the division made by Grimes (this volume). The vertebrate fossils reported herein were collected at San Bernardino County Museum localities SBCM 1.94.6-1.94.9, from the middle of the exposed portion of the brown member at the base of Black Hill, north of Pioneertown. A composite faunal list is given in Table I.

Table I. Pioneertown Composite Fauna

Hypolagus sp.	rabbit
Gomphotheriidae	elephant
Antilocapridae (lg)	large pronghorn
Camelidae (sm)	small camel
Equidae	horse
Sciuridae	squirrel
Geomyinae	geomyid rodent
Perognathus furlongi	pocket mouse
Cupidinimus n. sp. (sm) nr. *C. lindsayi*	kangaroo rat
Cupidinimus. sp. cf. *C. avawatensis*	kangaroo rat
Copemys sp.	cricetid rodent

PIONEERTOWN FAUNA

Hypolagus sp., the camel, and the sciurid are known throughout the middle and late Miocene (Savage and Russell, 1983). *Copemys* species are known from North America from approximately 16.4 Ma (Lindsay, 1991; Woodburne, 1991b; Reynolds, 1991b) to latest Hemphillian LMA (Czaplewski 1990). *Perognathus furlongi* spans an equally long period of time (Reynolds, 1991b; Lindsay, 1972).

The gomphothere family of primitive elephants first appeared in North American about 15 m.y.a. (Woodburne, 1991b) and persists through the Miocene.

The geomyid at Pioneertown is similar in measurements and morphology to *Parapliosaccomys* sp. from the Clarendonian LMA portion of the Crowder Formation (Reynolds, 1991a) and to an unnamed geomyid from the early Clarendonian Avawatz Formation (SBCM collections).

The equid from Pioneertown consists of an upper molar fragment. Although worn, the remaining portion is as tall as unworn molars of *Merychippus* sp. from the late Barstovian LMA. The antilocaprid horn core is significantly larger in diameter and in height than *Merycodus* sp. from the Barstovian LMA. Both these specimens suggest a LMA younger than Barstovian.

Cupidinimus n. sp. (small) from Pioneertown is smaller than *C. lindsayi* from the Barstovian LMA (Barnosky, 1986) and compares very well with the small species from the early Clarendonian LMA of Avawatz. The simple morphology of the P^4 suggests that this species was derived from *C. lindsayi*. *Cupidinimus* sp. cf. *C. avawatensis* is a large dipodomyinae and compares well morphometrically with *C. avawatensis* from Avawatz. The early Clarendonian LMA Avawatz fauna is dated at 10.9 Ma (Evernden and others, 1974, recalculated after Dalrymple, 1979).

DISCUSSION

The fauna from sediments north of Pioneertown appears to be referable to the early Clarendonian LMA. Two similar taxa of *Cupidinimus* occur at Avawatz with radiometric dates of 10.9 Ma (Evernden and others, 1964). The Clarendonian LMA spans a period of time from approximately 11.5-8.5 Ma (Woodburne, 1987, 1991b). Thus the age of the lower member of the Tertiary sediments north of Pioneertown may be as old as about 11.5 Ma, while the upper member interfingers with basalts dated between 9.3 ± 0.7 Ma and 6.86 ± 0.25 Ma.

The Pioneertown sequence was deposited on a middle Tertiary erosional surface developed on granitic rocks (Oberlander, 1972). Judging by the northwestward thinning of the section (Lowman, 1989), the sediments may have been deposited in a basin of limited extent, perhaps developed along local structures.

In contrast to the Pioneertown sequence, the Santa Ana Sandstone (Sadler, 1985; Strathouse, 1983) was deposited between 7-4 Ma and includes 6.2 Ma basalts (Woodburne, 1975) low in the sequence. Sadler (1985) recognizes basal arkosic sediments that contain middle Miocene (Hemingfordian? LMA) mammals. He notes that this supports the presence of a nonconformity between basal sediments in the Santa Ana drainage and the overlying thick sequence of sediments that comprises the Santa Ana Sandstone, as suggested by Jacobs (1982).

All but the lowermost Santa Ana Sandstone is younger than all but the very uppermost of the Tertiary sedimentary sequence north of Pioneertown. The uppermost sediments north of Pioneertown may overlap in time the oldest, quartzite facies of the Santa Ana Sandstone, but they are lithologically distinct. The formational status of the sediments north of Pioneertown can not be evaluated until possibly similar sediments (such as those recognized by Grimes in Yucca Valley) are studied. Until that time, the sediments north of Pioneertown that include early Clarendonian LMA vertebrate fossils and which interfinger with dated basalts may be referred to as the "Pioneertown Sequence" to indicate their temporal and geographic distinction from the Santa Ana Sandstone.

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Quaternary Geology of the Spy Mountain Region, Landers Quadrangle, San Bernardino County, California

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ABSTRACT

The Spy Mountain region, north of Yucca Valley and east of the Bighorn Mountain range, is near Landers, in the southcentral Mojave Desert, California. Exposed bedrock consists of gneiss and migmatite that has been intruded by at least two generations of Mesozoic plutonic rocks. Quaternary sedimentary deposits nonconformably overlie Precambrian and Mesozoic metamorphic and igneous rocks. These deposits include fluvial, alluvial, aeolian and lacustrine sediments, in part of Pleistocene (?) age. Two landslides and several rock falls formed on steep slopes that characterize the bedrock exposures. Deeply incised canyons due to uplift and erosion have contributed to the thick, alluvial fill at the base of the mountains. The northwest-striking Homestead Valley fault has a high-angle reverse to vertical, right lateral sense of movement. It has had historic activity as recently as 1979. Only minimal displacement of the fault since its inception is suggested by similar rock types at Spy Mountain and Homestead Mountain, which are separated by this fault.

LOCATION AND ACCESSIBILITY

The study area is in the south central part of San Bernardino County, California near the southern edge of the Mojave Desert Province, east of the San Bernardino Mountains and north of the Little San Bernardino Mountains. State Highway 247 provides access from the north and west (Fig. 1). Graded dirt roads and jeep trails provide good access to most parts of the area.

PHYSIOGRAPHIC SETTING

The Spy Mountain region is part of a broad, alluvial plain which is characterized by northwest-trending faults and bedrock exposures. With the exception of Pipe's Wash, drainages slope gently to the northeast. Pipes Wash flows to the northwest until it reaches the northern extreme of Spy Mountain, where it changes course abruptly in response to regional drainage patterns. Elevation ranges from approximately 829 m above mean sea level (MSL) at the valley floor to a high of 1054 m above MSL atop Spy Mountain (Fig. 1).

The area is about three miles east of the northeast corner of the San Bernardino Mountains. These mountains achieve elevations of greater than 2743 m MSL. To the south, the

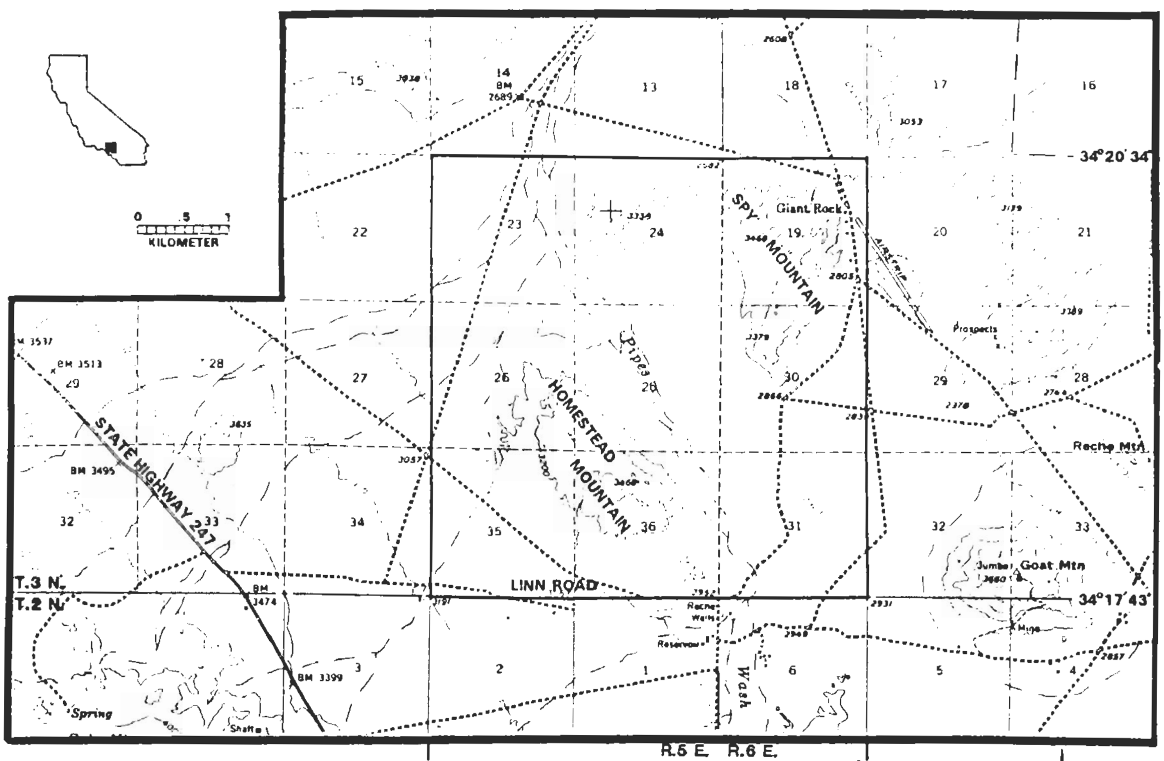

Figure 1. Location map for study area showing topography. Reference: U.S.G.S. 15' Emerson Lake Topographic Quadrangle, 1972, Scale 1:24,000.

Little San Bernardino Mountains are approximately 1768 m above MSL. The closest mountain ranges to the north and east are the Rodman and Bullion Mountains, respectively.

The area is named (Umbarger, 1992) for its most distinctive geomorphic feature, Spy Mountain, which lies in Homestead Valley just south of Johnson Valley. An unnamed, large exposure of bedrock situated to the west of Spy Mountain is herein referred to as "Homestead Mountain".

PREVIOUS INVESTIGATIONS

Hewett (1954) mapped faults in the Homestead Valley region as part of his fault map of the Mojave Desert region. Further geologic mapping of the area was conducted by Dibblee (1967). In 1979, Hawkins and McNey conducted a geologic study of the area after a swarm of earthquakes occurred on the Homestead Valley fault.

Several authors performed further seismic and geophysical studies at Homestead Valley in 1980. Their work was published in *California Geology* in a series of seven articles. Researchers included McJunkin and others of the California Division of Mines and Geology (CDMG); Stierman and others, of the University of California, Riverside; Parker and Schick, from Geothermal Surveys, Inc.; Williams and McWhirter of San Bernardino Valley College; Hill and others, of the California Institute of Technology (CIT) and CDMG; Ebel, Hill and Pechmann from CIT,; and Hutton and others, from CIT and the USGS.

Manson (1986) performed an investigation of the Homestead Valley fault and the Johnson Valley fault for the California Division of Mines and Geology. This study was conducted for a fault evaluation program to determine if these fault traces were "sufficiently active and well-defined" to be included for zoning under the Alquist-Priolo Special Studies Zones Act. Miller and Morton (1980) conducted field studies and sampling to determine potassium-argon apparent ages on minerals from crystalline rocks in the eastern Transverse Ranges and southern Mojave Desert, including the Spy Mountain region. Lewis (1972) conducted a ground-water resources study of the Yucca Valley-Joshua Tree area, including Landers.

QUATERNARY DEPOSITS

General Description and Age Relationships

Quaternary sedimentary deposits nonconformably overlie Precambrian and Mesozoic bedrock. These sediments include fluvial, alluvial, lacustrine, debris flow, landslide and aeolian deposits.

Age relationships, as determined in the field, are too vague to enable accurate stratigraphic positions, thus differentiation between the Quaternary deposits is based upon topographic position and geomorphic expression.

The Pleistocene is represented by older fanglomerate and lacustrine deposits, as well as debris flow, alluvial and landslide deposits. Older alluvium constitutes a major portion of the desert floor in and around the study area.

Holocene sedimentary deposits include aeolian sand that overlies the Pipes Wash older alluvium. Aeolian sands also interfinger with alluvial deposits that nonconformably overlie Mesozoic granite (Mgr) comprising the pediment in the core of Spy Mountain. Some of the aeolian sand is sufficiently stabilized so that it supports substantial vegetation.

Fanglomerate

Old, extensively dissected alluvial fan deposits are exposed on the western margin of Homestead Mountain. The fan deposits are composed of sand and gravel derived from gneisses and granitoid rocks that make up Homestead Mountain. These fans have been elevated and currently are being incised.

Old Caliche Deposits

Low-densite, carbonate-rich caliche deposits are porous, light beige-yellow to greenish-white, and consist of clayey silts rich in mica. These deposits crop out at the southern end of Homestead Mountain and cross over Linn Road south of the mapped area for approximately 0.5 km, although in this region they are mostly buried by alluvium. They are crudely bedded and horizontal and easily identified on the Landsat satellite thematic mapper image (JPL, 1984) by their reflection.

Based on visual and microscopic examination, these deposits are presumed to be pond or lake deposits. A regional study is necessary in order to determine the full extent of these outcrops and present a more complete picture of their origin. However, the location of the deposits is on the uplifted side of the Homestead Valley fault zone immediately west of and adjacent to Pipes Wash. It is feasible that there was a pond or marshy area that drained into Pipes Wash from the Transverse Ranges highlands during the Pleistocene, and that the ponded area later was uplifted due to activity on the fault and its eastern portion subsequently removed by erosion.

Old Alluvium

Old alluvium comprises a large part of the desert floor in and around the study area. Of probable Pleistocene age, the coarsest beds in the Spy Mountain region are exposed at the southern end of Pipes Wash. They ultimately pinch out at the north end of Pipes Wash.

Within the old alluvium, approximately 10 m of fluvial channel deposits are well-exposed in the banks of modern-day Pipes Wash. These beds are poorly- to moderately-sorted, subrounded to subangular, weakly-cemented silty sands that are interbedded with poorly-sorted, pebble to boulder gravel. Clasts up to 0.6 m in diameter are angular to well-rounded.

The channel deposits exhibit essentially horizontal bedding. Bedding is multiply-graded from coarse gravels to fine sands and silts.

Upstream (the southern end), dissected channel deposits consisting of subrounded to subangular gravel beds range from 15 to 60 cm thick in a sand-supported matrix. These deposits contain pebbly sands and lenticular gravels that may represent stream flood deposits of low viscosity (Tucker, 1981). Subsequently, they were buried by younger alluvium and re-exposed by on-going stream downcutting.

Clasts of Precambrian to Tertiary age derived from local uplands include vesicular basalt, granitic rock fragments (granodiorite, basalt, latite, quartz monzonite, granite, aplite, gabbro and diorite), and metamorphic rock fragments consisting of metavolcanic rocks, quartzite, gneiss and schist.

Old alluvium supports more vegetation than younger alluvium. This probably reflects the development of argillic, pedogenic soils.

Debris Flows

Northeast of Homestead Mountain is an apparent debris flow deposit of presumed Pleistocene age and unknown source. The deposit is well-indurated. The debris flow deposit consists of a notable reddish-brown, aplite-rich sand with coarse gravels and a poorly-sorted, sand-supported matrix consisting of carbonate-cemented, clayey sand. Clasts consist dominantly of garnet aplite, and also include representative clasts of the bedrock of Homestead Mountain.

Exposed in low, moderate- to gently-sloped ridges, it is tilted to the northeast and comprised of thin, internally-massive beds up to 0.75 m thick that trend west-northwest and dip an average of N70W/16NE. Source of the tilt may be derived from drag due to movement along the Homestead Valley fault. Conversely, the tilt may be derived from primary dip from Homestead Mountain.

This debris flow deposit is in fault contact with the Cretaceous (?) quartz monzonite of Homestead Mountain. Because of faulting, this deposit has altered to finely laminated and desiccated clay at the fault contact.

Landslides

The area contains two landslides and several piles of rock talus. Both landslides are on the eastern and northeastern side of Homestead Mountain, and have failed from fairly steep bedrock slopes of approximately 22°.

At the middle-eastern margin of Homestead Mountain, the toe of one of the larger bedrock slides has been laterally eroded by Pipes Wash, leaving it approximately 12 m above the floor of the channel. The arcuate nature of the failure suggests a type of slope movement classified by Varnes (1978) as a rotational rock slump.

The deposit has a muddy matrix with poorly-sorted, angular clasts, including boulders up to 1.8 m in diameter. Calcium carbonate coatings cover the surface. Extensive clay alteration of clasts and the inner matrix surface has occurred, observed in a small cavity within the toe of the landslide. Alteration is from ground water that has infiltrated the landslide. The interior surface is damp to the touch.

Within this same cavity, the landslide deposits are in direct contact with Precambrian amphibolite that is heavily chloritized. It is postulated that Precambrian amphibolite rocks were intruded by Mesozoic plutonic rocks, which later failed on the weaker amphibolite. Cause of the landslide is partly attributable to saturation during the Pleistocene epoch. In addition to saturation, ultimate failure may have been aided by activity on the Homestead Valley fault, although field evidence for this mechanism is sparse.

The landslide at the northeastern end of Homestead Mountain is a large rock block slide, as per Varnes' (1978) classification. Below its head scarp is a small, enclosed basin approximately 38 m in diameter. Field evidence, including fractures and angular to subangular clasts, indicates that quartz monzonite boulders ruptured from pre-existing discontinuities, possibly joints, and formed rock piles at slope bottom in the shape of fans or aprons. This slide has a rotational aspect to it, as indicated by a concave-upward rupture surface.

Based on the nature of the slide material (bedrock), and the elevation from which the slide occurred, rate of movement probably was very rapid. Rubble from this landslide overlies the trace of the Homestead Valley fault.

Colluvium

Colluvium consists of granitic and metamorphic residual soil derived from the San Bernardino-Joshua Tree (?) source terrane. Large, sub-angular to sub-rounded potassium feldspar cobbles up to 6 cm in diameter are residual augens from Precambrian augen gneiss. Old, varnished surfaces are intermixed with fresh, broken surfaces from recent movement, presumably triggered by the recent earthquake activity on the Homestead Valley fault.

Soils within the colluvium are loosely cemented, permeable, light brown to yellow colored, arkosic sands. Calcrete is prominent near contacts and in shear zones. The contact between the colluvium and underlying lithologies is gradational, except where it is a fault contact.

Young Alluvium

Recent alluvium occurs in active washes and stream channels and consists of arkosic sand and gravel that is predominantly feldspathic, poorly-sorted, and poorly-bedded. It contains granitic, sedimentary and metamorphic clasts of clay to boulder size derived from Pipes Wash, the San Bernardino Mountains, and the Bighorn Mountains. Sediments are medium- to coarse-grained, sub-angular to sub-rounded, uncemented, poorly sorted sands, gravels, and cobbles. Large boulders, some up to 4 m in diameter, are present within the deposit.

A review of several water well drillers' reports (Department of Water Resources, 1960-1985) indicates that depth of alluvium ranges anywhere from 28 to 135 m in Johnston Valley, just west of the Homestead Mountain. Well data is sparse in Pipes Wash and at Spy Mountain.

Depth of sedimentary deposits, as measured by the bouguer gravity anomaly map of the Twentynine Palms Marine Corps Base and Vicinity (U. S. Geological Survey, 1984), is as much as 120 to 150 m in a study area immediately to the east and north of Landers. Beyond the easternmost boundary of Spy Mountain, sedimentary deposits consisting of both older and younger alluvium, range from approximately 150 to 305 m thick.

Aeolian Sand Deposits

Holocene aeolian sand is deposited on the surface of the eastern bank of Pipes Wash, where it interfingers with young alluvium. It overlies Mesozoic bedrock and older alluvium in Pipes Wash. The stabilized portions of the deposit supports vegetation, while newly arrived sands remain uncemented.

Aeolian sand is fine-grained and quartzofeldspathic. A pedogenic B-horizon that would be expected at the stratigraphic top of the older alluvium and beneath the A-horizon is missing, possibly blown away by southwesterly winds or eroded by streams.

FAULTING

Regionally, a complex interaction of structural elements of the continental crust is expressed by the right-lateral San Andreas fault system, the left-lateral Garlock fault and the northwest-striking, right-lateral faults of the Mojave Desert Province. Movement on these faults is dated at Late Miocene to Recent, most of which occurred from the Pliocene to Recent (Brose, 1985; Garfunkel, 1974). The study area lies between two of these major northwest-striking faults: the Johnson Valley fault and the Camprock fault. Within the study area is the Homestead Valley fault.

The Homestead Valley fault is well-exposed at the northeast end of Homestead Mountain, where the sense of movement appears to be right oblique slip. The fault has displaced the upper block of Mesozoic crystalline bedrock (Mqm) over Quaternary debris-flows (Qdf). One fault plane measured in the field showed the fault to be moderately-dipping at N30W55SW. Slickenslides in the clay gouge of the exposed fault plane trend N12E and plunge 42SW.

A second measurement taken approximately 100 m northwest of the initial exposure reveals the fault as a vertical plane striking N45W (photo 25), with the Homestead Mountain block uplifted relative to Pipes Wash. Between the two stations, a third reading was N33W60SW.

Along the southeast end of Homestead Mountain at the western boundary of Pipes Wash, the trace of the Homestead Valley fault has little to no surficial expression, other than what may be sublinear, eroded old fault scarps approximately 12 m above the base level of Pipes Wash. The trend of these linear features is N25W. These features could be caused by lateral stream erosion; therefore, the fault is mapped in this area as inferred. However, a lineament is recognizable at this point when analyzing air photographs and satellite imagery of the area (JPL, 1985), as evidenced by changes in texture, lithologic signature and reflectivity.

These possible old Homestead Valley fault scarps consist of a loosely consolidated soil matrix with gravel and debris. Several talus slopes (colluvium) overlie a possibly uplifted old stream terrace consisting of older alluvium (Qoal) that includes clasts of vesicular basalt. The basalt clasts, believed to have a Pioneertown Basalt provenance (Lowman, 1989), were transported to the area via Pipes Wash. Existence of the basalt clasts 12 m above stream base elevation could be evidence for relative vertical uplift in this area.

The Homestead Valley fault apparently blocks the flow of ground water from west to east. Two local residents, whose home sits on the western, uplifted block of the Homestead Valley fault, drilled to a 95 m depth in order to reach ground water. On the downdropped block in Pipes Wash (at the southeast corner of the intersection of Linn Road and Belfield Road), residents reportedly only drilled 10 m to reach groundwater.

SEISMICITY

Seismograph studies conducted by Stierman and others, (1980) subsequent to the 1979 earthquakes revealed that the Homestead Valley earthquake occurred at relatively shallow focal depths (less than 5 km). Accelerographs indicated that ground accelerations were rapidly attenuated with increasing epicentral distance (McJunkin, 1980). Right-lateral strike-slip movement was documented with at least 10 cm near the northern end of the fracture zone, along a 3.25 km section of the Homestead Valley fault north of the study area. Ebel and others, (1980) calculated seismic moment to be 7 x 10^{23} dyne-cm, and thus a calculated average displacement of 17 cm.

DISCUSSION

The eastern boundary of Pipes Wash is distinguished by a >12 meter high levee consisting of Quaternary sedimentary deposits. A major lineament, easily observed on aerial photographs and satellite images, trends along this levee and extends northward out of the Spy Mountain region. The parallelism of this lineament with the Homestead Valley fault opposite it could indicate the presence of an unmapped fault, however, there is insufficient field evidence to verify this.

Studies were conducted (Hill and others, 1980) of surface cracks that appeared after the 1979 earthquakes. During the course of these studies, two sets of extension cracks along the axes of Pipes Wash were shown to have left-lateral, right stepping *en echelon* fractures, whereas cracks on the western boundary of the stream channel indicated relative oblique slip movement with a right-lateral and nearly vertical component. The resultant sense of movement, according to Hill and others (1980), could indicate a downdropped, south-moving block relative to the block east of these fractures.

In the same study, Hill and others (1980) observed within the extensive slumping of sediments that occurred along the levee, a large crack greater than 9 m long, with vertical offset, east side up, of up to 3.2 cm. This particular crack was not typical of the other post-earthquake cracks. Hill and others (1980) contend that this crack may have been the result of vertical movement along an unmapped fault. The authors acknowledged, however, that the crack could merely be an incipient slump crack.

The linearity just discussed stands out starkly on the aerial photographs, satellite image and ortho-photo quads. Contrasting textures of the ground surface in these photos is indicative of a subsurface discontinuity. Additionally, a slight curvature of the lineament, concave west, can be seen in the orthophoto quads.

Geomorphology

The Spy Mountain region is a desert landscape with bouldery outcrops and a low relief profile. Homestead Mountain and Spy Mountain are two northwest-trending remnants of Precambrian basement and Mesozoic plutonism that may have been connected at one time. Presently, they are separated by a narrow drainage (Pipes Wash) that flows through the area.

Precambrian, metamorphic terrain comprises deeply "V"-shaped canyons and rounded crests. The less eroded, rugged topography rendered by plutonic rock suites is typical of domed inselbergs (Oberlander, 1972), including boulder mantled slopes and pediments with grus in between them. Cuboidal jointing patterns characteristic of granitic terrane is prominent.

Old, depositional surfaces are reflected by the presence of older, dissected fanglomerates currently undergoing degradation. They are lower in elevation than the bedrock canyons. Using some of the criteria established by Christenson and Purcell (1985), including drainage pattern, depth of incision, fan surface morphology, desert pavement and varnish, the fanglomerates may represent early Pleistocene deposition. The old fans have stream incisions greater than 10 m and have been cut off from their original source areas by modern drainages.

The older caliche deposits represent an old depositional basin at somewhat lower elevations than the older fanglomerates. Their heavy degree of dissection and induration indicate that they may be as old as 500,000 BP (Reynolds, 1989).

Younger, steeply dissected canyons have developed at the margins of Homestead Mountain and Spy Mountain, aided, at least in Homestead Mountain, by tectonic uplift along the Homestead Valley fault.

The slopes of Homestead Mountain consist mainly of slope-forming colluvium and slope wash. Few outcrops remain on these steep slopes due to the nature of the bedrock, extent of fracturing, and weathering.

On the east face of Homestead Mountain, colluvium stops abruptly at what is either a meander scarp cut by Pipes Wash, or the buried trace of the Homestead Valley fault as mapped by Dibblee (1967). The poorly expressed fault in this section of Homestead Mountain has elevated older alluvial deposits, which subsequently have been undercut, and which grade into the slightly incised, young alluvium of Pipes Wash. To the northwest, the trend of the fault is physiographically well-expressed.

The horseshoe shape of Spy Mountain is an anomaly. One possible explanation is the younger, pediment-forming granite (Mgr) in the concave core of Spy Mountain. The granite was intruded into the inner and central margins of the ridge-forming leucogranite. This granite slopes gently westward at about 3-5° into older alluvium and aeolian sand deposits. It is phaneritic and has a tendency to flake; being coarser-grained than the older granite, it is less resistant and has eroded so substantially as to give the center of Spy Mountain its present configuration.

There is no evidence to indicate that Spy Mountain is a remnant of a collapsed caldera or ring dike structure. However, shapes of plutons tend to be controlled by magma-induced fracture patterns. It is possible that at one time, given their similar lithologies and relative distributions, Spy Mountain was connected with Homestead Mountain on the opposite side of Pipes Wash. If Spy Mountain is part of an uplifted fault block relative to Pipes Wash, then the apparent linear truncation of the western section of Spy Mountain can be explained by faulting.

The Spy Mountain region is considered by Dibblee (1982) to be part of the north block of the San Bernardino Mountains. This block includes basement rocks characteristic of the Mojave block, and the Joshua Tree terrane of Powell (1982). If thus interpreted, the Spy Mountain region represents the northeastern extreme of an elevated surface plateau sloping gradually eastward into the Mojave Desert plain.

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Fissuring near Twentynine Palms, California

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ABSTRACT

A 1 km long ground crack occurs north of the Mesquite Lake Fault in the playa basin. This crack trends due north and was thought by earlier investigators to be fault related. However, the discovery of other parallel healed cracks and subsurface detailed studies have shown these features to be desiccation related. Results of drilling show highly plastic clay over most of the playa basin at relatively shallow depth and perched groundwater at about 70 feet below the surface. With groundwater present at about 20 meters in clay and the potential for a very large capillary fringe, possible as much as 10 to 12 meters, then the actual control of the ground cracking is likely desiccation of the clays at the 8-10 meter depth. Where overlain by sandy alluvial playa filling deposits, these may sustain cracks to the surface above the desiccated clay-layer fracture.

INTRODUCTION

This paper presents the results of ongoing work begun in 1982. The studies have been conducted in and around Mesquite Lake Playa, north of Twentynine Palms, California (Fig. 1). This area is a part of the southern Mojave Desert and lies just north of the eastern end of the Transverse Ranges and Joshua Tree National Monument. The geology and structure are typical of the Mojave Desert, that is, rugged granitic mountains shaped by northwest-striking faults with mostly right-lateral strike-slip movement. The Mesquite Lake Playa area is unique to the Mojave Desert in that the northwest striking Mesquite Lake Fault terminates or merges with east-west striking left-lateral strike-slip Pinto Mountain fault zone south of Campbell Hill and east of the town of Twentynine Palms. Mesquite Lake Playa is a low basin between a relatively flat high plains area to the west and the western spur of the Bullion Mountains to the east. Most of the playa and study area is within the boundary of the Marine Crops Air Ground Combat Center at Twentynine Palms.

GEOLOGY OF THE LAKE BASIN

The playa basin stratigraphy may be similar to that described by Wells and others (1989), based on cores drilled in Silver Lake. These Silver Lake cores reveal finely bedded to laminated sequences of sandy and clayey layers, and thick sequences of greenish clay. If Mesquite basin is a similar basin, it may have dried up after the pluvial period between 18,400 and 11,400 years ago (Oxygen Isotope Stage 2) into the playa environment of today. This drying is continuing today as groundwater declines regionally, creating large desiccation cracks in many of the playa sediments from the shrinking of the expansive playa clays.

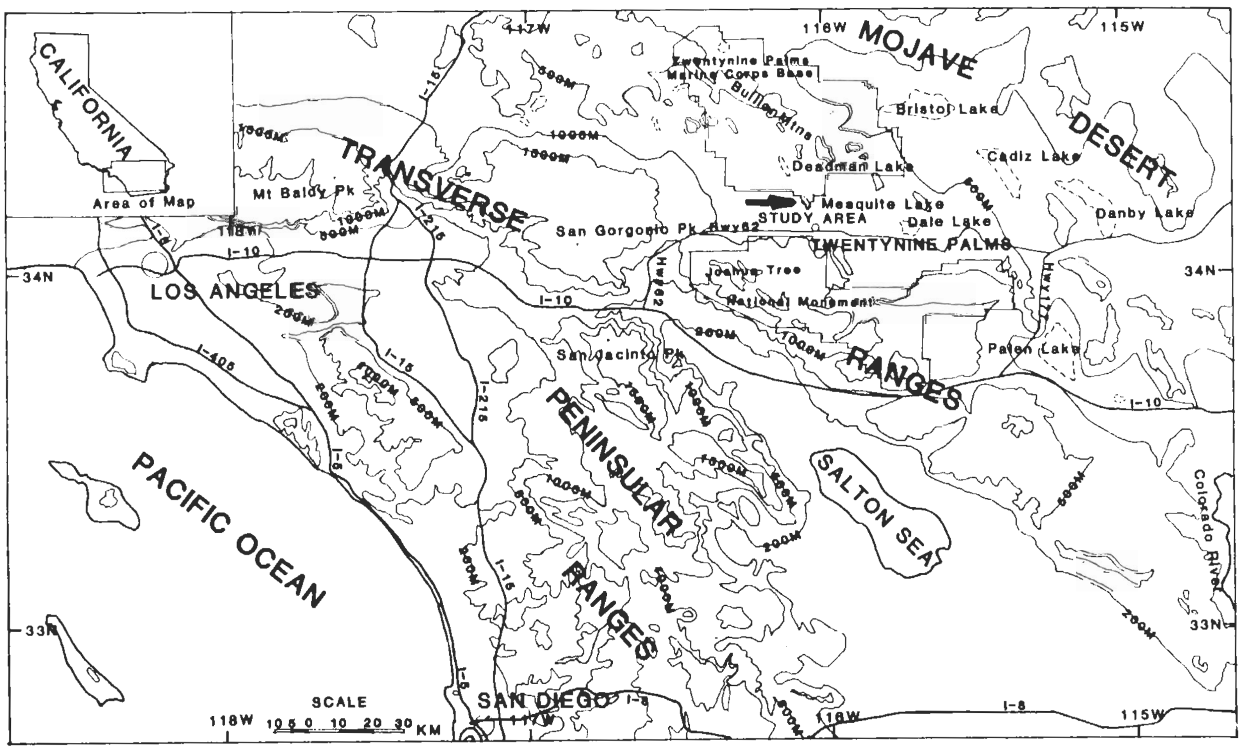

Figure 1. General contour index map of S. Calif., STUDY AREA marked by arrow near upper middle. Contours in meters at 500 m intervals.

GROUNDWATER

Perched groundwater was noted by Wahler Associates (1984) at a depth of about 21 meters in clay and sand near the south end of the playa, and at about 29 meters near the north end of the playa. Therefore, the water table of at least a regional perched groundwater body exists at about an elevation of 1687 feet (514 meters) above mean sea level.

HISTORY OF GROUND CRACKING

Ground cracks are present in Mesquite Lake Playa trending northwards near the southeast end of the lake bed. These *en echelon* fissures step to the right in four one-half kilometer segments and were referred to by Fife (1978) as the Airfield Fault (see Fig. 2). The cracks are easily visible from the ground and air as narrow filled depressions with abundant vegetation in an otherwise barren environment (see Fig. 3).

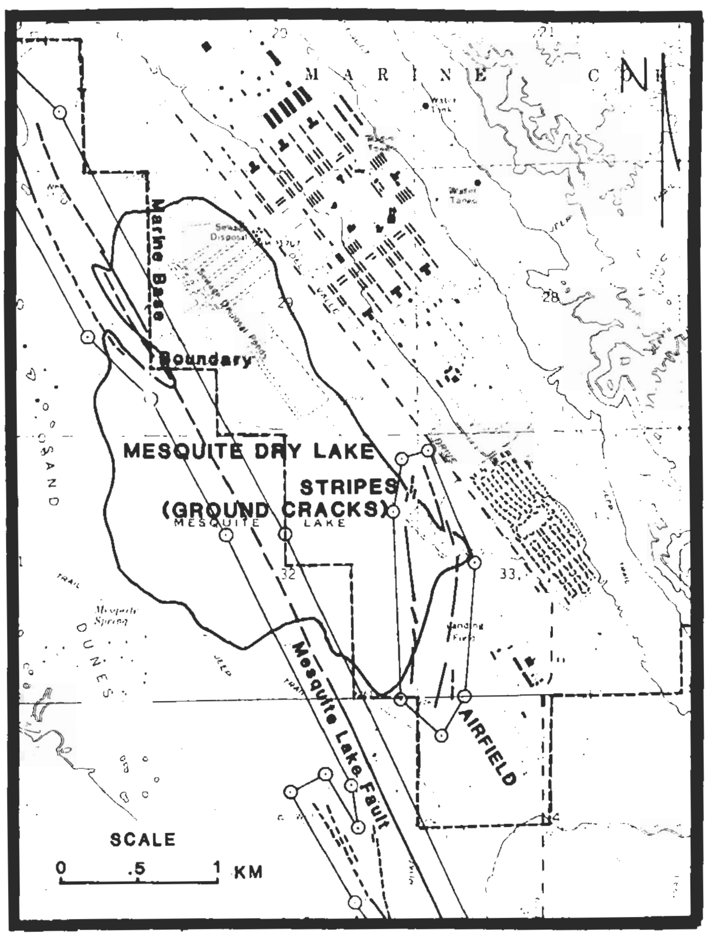

Figure 2. Portion of the Twentynine Palms Special Studies Zone Map showing ground cracks (stripes) and fault zonation. Mesquite Lake fault trends NW, west of the stripes.

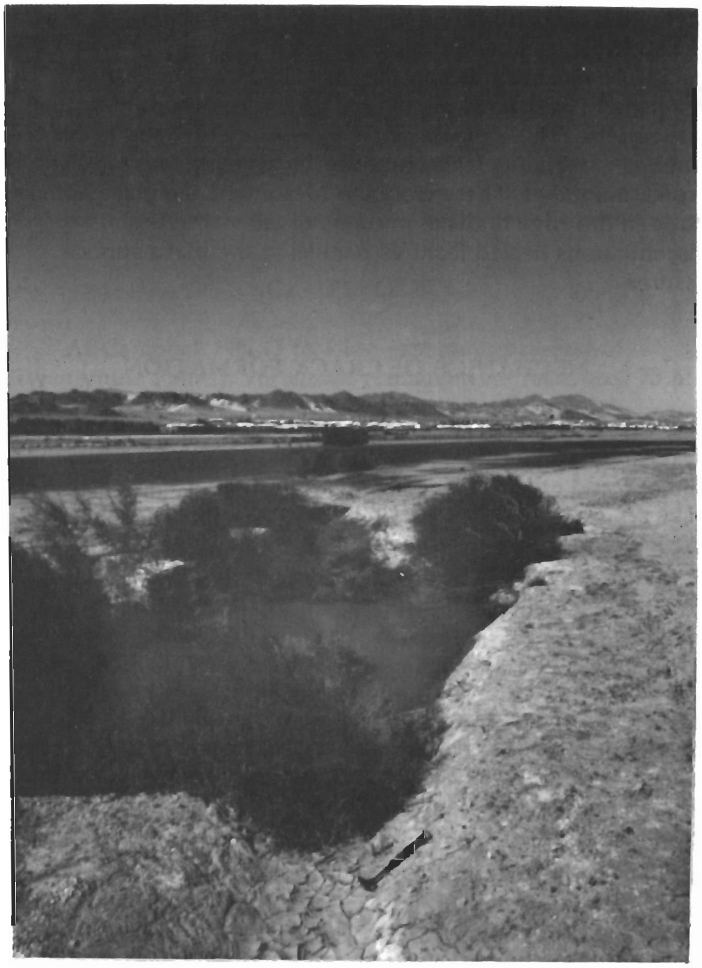

Figure 3. Vegetated, linear depression formed by infilling of a ground grack formed in 1976. 45cm-long geologic pick for scale.

Trenching of the fissures by Wahler Associates (1984) showed that these features are formed by tension with open linear cavities below the filled depressions as well as large failed blocks in a funnel-shaped downward-narrowing system of cracks. The zone is bounded by single breaks on either side of the main feature which merge at a depth of about 15 feet to a zone no more than a few inches wide. The layered clays and silty sands through which the cracks have propagated are thinly bedded horizontal strata which are not offset across the fissures. A previous study by Fife (1978) described similar subsurface features just after the cracks originated following heavy rainfall in the summer of 1976. The *en echelon* nature of the ground crack suggested to Fife (1978, 1980) that these cracks were essentially the surface expression of a left-lateral fault. However, the subsurface cavities and linear voids suggest tension as the stress mechanism, not compression and shear, and it appears the fault model is untenable. Neal and others (1968) have shown that discontinuous *en echelon* "stripes" like those in

Mesquite Lake Playa are one of several types normally found in desert playas due to desiccation.

More recent studies in the playa by Schaefer Dixon Associates (1988) have discovered ground cracks outside the playa basin. These cracks are narrow funnel-shaped linear features continuous with and filled by an overlying reddish brown mudflow. These cracks in Mesquite Lake playa appear through the edge of distal portions of alluvial fans and are discontinuous healed features parallel to the playa surface fissures.

MECHANICS OF CRACK FORMATION

Crack formation in Mesquite Playa is caused by long term desiccation of the underlying clayey playa and lake sediments. As Neal and others (1968) and numerous other investigators have reported, desiccation, whether from long term climatic change or groundwater mining, causes fissuring. Historical records of wells from years back show groundwater to be at a depth of about 80 meters, as it is today. However, long term desiccation of the desert areas following the pluvial period of the last glaciation has had a major impact on numerous desert basins as reported by Neal and others (1968). Although Fife (1978, 1980) concluded that these ground cracks are fault related, their appearance and subsurface geometry are identical to features of ground cracks in other desert basins shown by Neal and others (1968) and are best explained as being related to long term desiccation.

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Pleistocene Fossil Vertebrates from Twentynine Palms, California

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INTRODUCTION

Eroded exposures of alluvial and fluvial sediments in the Twentynine Palms area yield a diverse, Rancholabrean Age assemblage of primarily large mammals (Jefferson, 1991a). Other vertebrate paleontologic sites of comparable age and geologic setting within the eastern Mojave Desert include Pinto Basin (Campbell and Campbell, 1935; Scharf, 1935; Jefferson, 1986, 1991a) and Piute Valley, California (Jefferson, 1991a), and Las Vegas Valley, Nevada (Reynolds and others, 1991). Vertebrate fossils from the Twentynine Palms area were first reported by L. F. Noble in 1918 (Bassett and Kupfer, 1964), and Bachellor (1978) has described the Quaternary sediments of the area and identified fossiliferous horizons within the stratigraphic sequence.

Table I. Faunal List of Assemblages from Campbell Hill, Surprise Springs, and 29 Palms Gravel Pit Localities

Campbell Hill
Institutional Locality Numbers: LACM 4281-4283
SBCM 1.86.4, 1.86.9, 1.86.11-1.86.13
Taxa:
- *Gopherus* sp.
- *Thomomys* sp.
- *Taxidea* sp.
- *Felis concolor*
- *Smilodon* sp. cf. *S. fatalis*
- *Mammuthus* sp.
- *Equus* sp. (large size)
- *Equus* sp. (small size)
- *Camelops* sp.
- *Hemiauchenia* sp.
- *Odocoileus* sp. cf. *O. virginianus*
- *Capromeryx* sp.
- Antilocapridae
- *Ovis* sp. cf. *O. canadensis*

Surprise Springs
Institutional locality number: LACM 3350
Taxa:
- *Equus* sp.
- *Camelops* sp.
- *Hemiauchenia* sp.
- *Bison* sp.

Twentynine Palms Gravel Pit
Institutional locality number: SBCM 1.86.1-1.86.3
Taxa:
- *Equus* sp. (large size)
- *Camelops* sp.
- Antilocapridae or Cervidae
- *Ovis* sp. cf. *O. canadensis*

CAMPBELL HILL

At Campbell Hill (3 to 4 km east and northeast of Twentynine Palms), vertebrate remains were recovered 73 m below the eroded top of an approximately 358 m thick stratigraphic section of fluviatile sandstones and gravels. The deposit is informally known as the Campbell Hill Formation (CHF) (Bachellor 1978), and was mapped as Quaternary older gravels (Qog) by Dibblee (1968). This unit contains lithologically distinctive sedimentary clasts derived from the San Bernardino Mountains about 45 km west of Twentynine Palms (Dibblee, 1968; Bachellor, 1978). The beds have been uplifted along the northeast side of the northwest-southeast trending Mesquite Lake Fault, are gently folded, and dip about 30 degrees to the northeast.

TWENTYNINE PALMS GRAVEL PIT

The Twentynine Palms Gravel Pit area (4 km east of Twentynine Palms and south of Highway 62) lies immediately west of the main trace of the Mesquite Lake Fault in an area crossed by parallel branches of the Mesquite Lake Fault near its intersection with the Pinto Mountain Fault to the south. Although Dibblee (1968) mapped this area as Qoa (Quaternary older alluvium), it is clear from the detailed work of Bachellor (1978) that the eroded exposures of sandy gravels and fine-grained conglomeratic alluvial sediments can be assigned to either the CHF (= in part to Qog of Dibblee, 1968) or the Twentynine Palms Formation (TPF) (informally named by Bachellor, 1978). Here, <40 m of CHF dips gently to the northeast, is in fault contact with, and/or is overlain by members of the TPF. Precise stratigraphic correlation of these exposures of CHF with the Campbell Hill section has not been made. Although there is no reason to believe that the vertebrate remains here were recovered from local exposures of the TPF, the TPF is laterally equivalent and interfingers with the CHF (Bachellor, 1978). Distinctive sedimentary clasts in the TPF are derived from the Little San Bernardino and Pinto Mountains a few km south of Twentynine Palms.

SURPRISE SPRINGS

The vertebrate remains from Surprise Springs (approximately 20 km northwest of the Twentynine Palms Gravel Pit and 8.5 km west of Deadman Lake) were also recovered from sediments mapped as CHF (Bachellor, 1978) (= in part to Qoa of Dibblee, 1967). Near Surprise Springs, the flat-lying, medium-grained sandy alluvial strata are over 235 m thick. Exposures of these deposits crop out west of the northwest-southeast trending Mesquite Lake Fault and north of the Pinto Mountain Fault in low hills over a broad, 15 by 35 km area from south of Twentynine Palms to northwest of Deadman Lake.

VERTEBRATE REMAINS

Age and Correlation

The vertebrate remains from Campbell Hill, Surprise Springs, and the Twentynine Palms Gravel Pit localities were recovered from fluvial and alluvial deposits that are in part laterally correlative, and are probably of comparable, later mid-Pleistocene age (Morrison, 1991). The Bishop Tuff, dated at 730 ka BP (Sarna-Wojcicki and others, 1984), has been tentatively identified 90 m below the top of the Greenish Sand Member of the TPF (Bachellor, 1978). This ash is low in the TPF, and is presumably correlative with the lowest exposures of the CHF at Campbell Hill and Surprise Springs. Although the presence of *Bison* sp. has not been confirmed from Campbell Hill or the Twentynine Palms Gravel Pit localities, *Bison* sp. from the CHF at Surprise Springs may approach the age of the earliest record in the region (about 280 ka BP at Lake Manix).

Campbell Hill (San Bernardino County Museum, SBCM 1.86.1, 1.86.2, 1.86.4, 1.86.9; Natural History Museum of Los Angeles County, LACM 4281—4283), Surprise Springs (LACM 3350), and Twentynine Palms Gravel Pit localities have produced remarkable Rancholabrean assemblages. Many of the taxa are regionally very rare (Jefferson, 1991b). Although the collections are not large, a total of fourteen mammals have been identified (Table 1). Of the three represented carnivore species, *Taxidea* sp. has been reported from two localities (Kokoweef Cave, Schuiling Cave), *Felis concolor* from three localities (Lake Manix, Mitchell Caverns, Schuiling Cave), and *Smilodon* sp. cf. *S. fatalis* from only one other locality (Lake China) in the Mojave Desert region (Table 2). In addition, the extinct pronghorn *Capromeryx* sp. is known from only one other desert locality (Schuiling Cave), and the occurrence of *Odocoileus* sp. cf. *O. virginianus* is the only record for the region.

The diversity and abundance of browsing forms and/or animals that are interpreted as having preferred more upland habitats within the desert — such as *Felis concolor*, *Capromeryx* sp., *Odocoileus* sp. cf. *O. virginianus*, and *Ovis* sp. cf. *O. canadensis* — probably reflects the proximity of the Twentynine Palms sites to the Little San Bernardino and Pinto Mountains (Jefferson, 1986). In contrast, most lowland assemblages are typified by the presence of large and small *Equus* sp., *Hemiauchenia* sp., and an abundance of *Camelops* sp. (Table 3). Lowland localities like those from the Daggett/Yermo (Reynolds and Reynolds, 1985) area and Lake Manix in the central Mojave Desert, usually yield few *Odocoileus* sp., *Antilocapra* sp. or *Ovis* sp. (Jefferson, 1991b).

Within the Mojave Desert region, *Odocoileus* sp. cf. *O. virginianus* is represented by a single, nearly complete left antler from the Twentynine Palms Gravel

Table II. Late Pleistocene vertebrate assemblages from Lake China, Lake Manix and the eastern Mojave Desert

Only genera of medium to large-sized mammals are listed. Localities follow from west (left) to east (right). Abbreviations: a = Antelope Cave, Kokoweef Cave, & Mitchell Caverns; b = Pinto Basin; c = Lake China localities; d = Daggett/Yermo localities; H = Campbell Hill; m = Lake Manix localities; p = Piute Valley; s = Schuiling Cave; S = Surprise Springs; T = 29 Palms Gravel Pit; v = Las Vegas Valley localities.

TAXA	l	d	m	s	H	S	T	b	a	p	v
Megalonyx			*								*
Nothrotheriops			*						*		*
Glossotherium			*								
Canis	*		*								
Arctodus			*								
Ursus			*								
Homotherium			*								
Smilodon	*		*		*						
Felis			*	*	*						*
Panthera			*							*	*
Mammuthus	*	*	*		*						*
Equus (large)		*	*	*	*		*		*		*
Equus (small)			*	*	*	*			*	*	*
Equus sp.	*				*		*				
Camelops	*	*	*	*	*	*	*		*	*	*
Hemiauchenia	*	*	*	*	*	*	*				
? *Navajoceros*									*		
Odocoileus	*				*				*		*
Capromeryx				*	*						
? *Antilocapra*			*		*				*		
? *Tetramerys*											*
Bison (large)	*									*	
Bison (small)	*		*			*		*		*	*
Ovis		*	*	*	*				*		

Table III. Relative abundance of large mammalian herbivores in Mojave Desert assemblates. Abundance in given as a percentage of the total number of identified specimens, NISP (Jefferson 1988).

Localities follow from west (left) to east (right). Abbreviations: c = Lake China localities (NISP = 583); d = Dove Springs localities (NISP = 14), m = Lake Manix localities (NISP = 766); CST = Twentynine Palms area localities (NISP = 47); v = Tule Springs localities (NISP = 563), Las Vegas Valley.

TAXA	c	d	m	CST	v
Edentata	<1	.	<1	.	1
Mammuthus sp.	5	36	4	2	6
Equus sp. (large)	26	.	9	38	10
Equus sp. (small)	<1	.	7	2	8
Equus sp.	26	14	16	57	18
Camelops sp.	56	21	64	23	61
Hemiauchenia sp.	1	.	15	11	1
Bison sp.	11	29	<1	6	12

Pit locality, SBCM 1.86.3. It is an extralocal taxon that does not presently occupy California, the Great Basin, or Colorado Plateau, but occurs in the Sonoran Desert. In this respect, its geographic distribution is similar to other extralocal species known from the eastern Mojave Desert region, such as the packrat *Neotoma albigula* (Jefferson, 1991a).

Bison sp. is known from only six of 47 (13%) late Pleistocene localities that yield large mammals in the greater Mojave Desert region (Table 2). These are Lake China (Fortsch, 1978), Dove Springs Wash (Whistler and others, 1991), Lake Manix, Pinto Basin, Piute Valley, and Surprise Springs (Jefferson, 1991b). This distribution may reflect the presence of seasonally favorable higher elevation habitats near localities where *Bison* sp. remains are found. The taxon is present (represented by two specimens) in only one lowland, mid-desert assemblage (Lake Manix) (Table 3). It is relatively well represented in assemblages from the southeastern Mojave Desert (Pinto Basin, Piute Valley, and Surprise Springs). *Bison* sp. is also abundant at sites located adjacent to the eastern flank of the Sierra Nevada (Dove Springs and Lake China) (Table 3). Assemblages from Las Vegas Valley (Tule Springs, Table 3) in the southernmost Great Basin also yield *Bison* sp. remains. *Bison* sp. is about twice as abundant throughout the Great Basin than in the Mojave Desert, and is present in 11 out of 41 (27%) large mammal assemblages from Nevada.

Most of the other large mammals from localities in the Twentynine Palms area occur throughout the region (Jefferson, 1991b). *Camelops hesternus*, *Hemiauchenia macrocephala*, *Equus* sp. (lg), and *Equus* sp. (small) are common in eastern California desert assemblages. However, the abundance of *Equus* sp. (large) from assemblages in the Twentynine Palms area appears similar to that in the Lake China fauna (Table 3).

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Æolian Geomorphology of the Dale Lake Sand Sheet

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INTRODUCTION

The Dale Lake Sand Sheet, located in the southern Mojave Desert, consists of climbing sand sheets extending about 5 km eastward from Dale Lake Playa (Fig. 1). The sand sheet terminates in a series of sand ramps which climb up the southwest flanks of the Sheephole Mountains and the adjoining Pinto Mountains (Fig. 2). Sand ramps are formed primarily by episodic deposition of æolian sediments blown against mountains, subsequently reworked by hillslope runoff, mantled by rock talus, and generally entrenched by fluvial activity, the latter leading to the formation of dune wadis. In the Dale Lake Sand Sheet, several dune wadis expose the underlying sediments (Fig. 3). The surface of the sand sheet is mostly stabilized by vegetation and veneered with rock talus from the adjoining mountains. Exposed sections in æolian deposits are rare in the southwestern deserts, and therefore these exposures are important for analyzing the paleoenvironmental conditions of æolian deposition and paleosol formation. The study site is part of a series of interconnecting desert basins in the southern Mojave Desert. The basins are believed to have acted as major pathways for æolian transport and deposition (Williams and others, 1991).

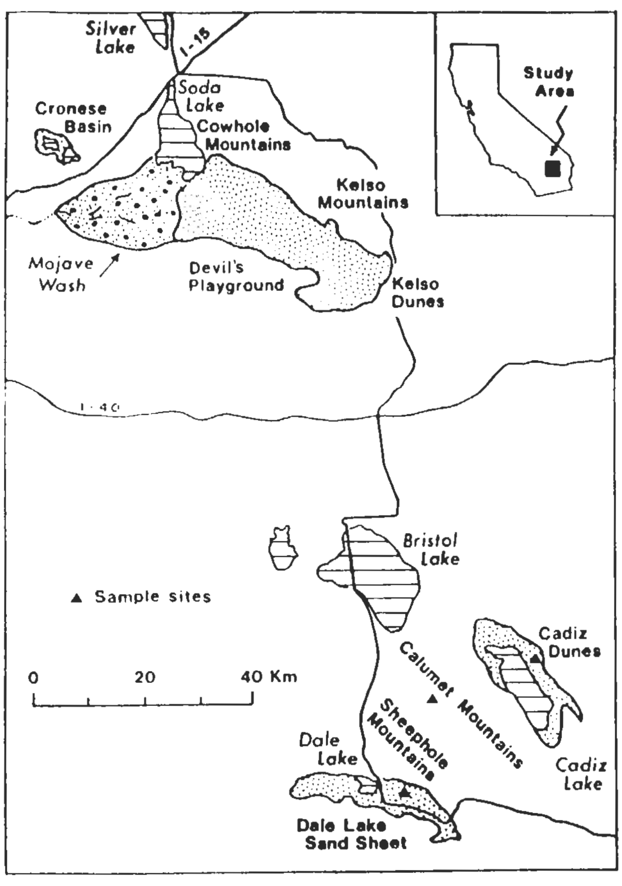


Figure 1. Location of the Dale Lake Sand Sheet and nearby dune fields.

GEOMORPHOLOGY OF THE SAND SHEET

At the Dale Lake Sand Sheet and surrounding areas, up to six æolian depositional phases separated by paleosols have been identified, on the basis of geomorphic, soil-stratigraphic, and scanning electron microscope (SEM) analyses of quartz-grain surface microtextures (Tchakerian, 1989, 1991, and Fig. 4). The following discussion is a brief summary of the geomorphological and sedimentological characteristics of the sand sheet. For more detailed analysis, the interested reader is referred to Tchakerian (1991).

Table 1 presents the results of the textural analysis. The sediments with the most fines are found in the stabilized climbing dunes of the Calumet Mountains, while the coarsest sands are found in Qe3. The most well sorted sediments are found in the crests of active transverse and complex dunes in Cadiz Valley, while the most poorly sorted sediments are from Qe3. Highest silt-clay percentages (4.8%) are found in the stabilized climbing dunes of the Calumet Mountains, followed by Qe2 (3.1%) and Qe1 (2.9%).

High silt and clay contents are related to a number of post-depositional weathering processes, including dust infiltration. This is largely accomplished through grain translocation and mineral disaggregation over time, the latter particularly effective on feldspars which, upon weathering, release various clay minerals and clay-size particles (Tchakerian, 1989). Additional sources for fines include airborne evaporites, such as calcium carbonate and sodium carbonate, owing to the fact that most dune fields are near to and downwind from plays, potential sources for both deflation and groundwater solutes. Airborne evaporitic materials also enhance the Ph conditions of the available moisture, thereby facilitating silica dissolution and reprecipitation (Tchakerian, 1989; Pye & Tsoar, 1990). Vegetated linear dunes in Rice Valley, California, and in the Cactus Plain in southeastern Arizona (east of the Colorado River), contain up to 6% silt and clay (V. Tchakerian, unpublished data).

Table 2 presents a summary of the SEM analysis. The quartz grains from Qe1 and Qe2 show high percentages of solution etchings, large hollows and pits (ranging in diameter from 5 to over 10 microns), blocky and nodular silica precipitation, and adhered particles and clay platelets. Silica precipitation, abundant on the surfaces of the quartz grains from the stabilized climbing dunes of the Calumet Mountains and from Qe1 and Qe2, is formed largely as a result of chemical weathering processes owing to the presence of vadose water and alkaline solutions (including desert dew),

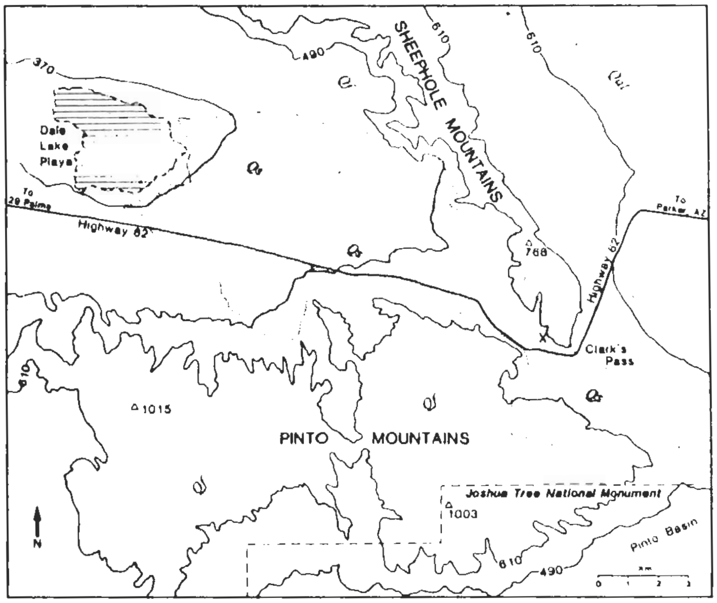


Figure 2. Close-up map of Dale Lake Sand Sheet (dotted land labeled Qs) terminating in sand ramps enveloping Sheephole and Pinto Mountains. X denotes dune wadi seen in Fig. 3.

which dissolve quartz (and silica minerals) and then promote reprecipitation during evaporative periods (Pye and Tsoar, 1990). The quartz grains from the stabilized climbing dunes on the Calumet Mountains show more pronounced chemical weathering microfeatures than those from Qe1 and Qe2, indicating perhaps a longer stable period, and thus could represent older deposits (Table 2). Energy dispersive x-ray analysis (EDAX) of the surface coatings found on the grains from the Calumet Mountains, Qe1, and Qe2 show the presence of iron oxide/oxyhydroxide, calcium carbonate, clay platelets, and other evaporitic materials, with the highest concentrations on the surfaces of grains from the Calumet Mountains (Tchakerian, 1991). The quartz grains from Qe4 to Qe6 show similar frequencies of mechanically and chemically derived microfeatures, with about 45% of the grains exhibiting upturned plates, and about 25% showing microfeatures indicative of largely chemical weathering processes (Table 2).

A principal component analysis was performed on the frequencies of quartz-grain microfeatures, to ascertain whether the different æolian deposits can be distinguished. The following micromorphological characteristics associated with mechanical and chemical weathering processes as well as with gross surface morphologies were used for the analysis: breakage blocks, conchoidal fractures, mechanical depressions, randomly oriented grooves, upturned plates, rolled topography, solution etchings, pitting, silica precipitation, fractured surfaces, angular outline, rounded outline, low relief, medium relief, and high relief. With the exception of Qe4, Qe5, and Qe6, all the æolian units were clustered separately (Fig. 5). The details of the statistical analysis can be found in Tchakerian (1989, 1991).

PALEOENVIRONMENTAL IMPLICATIONS

The data from the Dale Lake Sand Sheet (and from other æolian depositional sites in the east-central Mojave Desert) indicate significant oscillations in æolian activity during late Quaternary time (Tchakerian, 1989, 1991). The æolian deposits are believed to have accumulated largely in response to lowering of lake levels and a consequent increase in fine sediment availability in those desiccating lake basins, possibly accompanied by winds that were stronger and more persistent than today (Wells and others, 1987; Tchakerian, 1989).

Six stages of æolian deposition are indicated based on the relative weathering data from quartz grain microfeatures and soil-stratigraphic relationships (Table 2). Stages I to III most likely represent deposits that predate the last Wisconsinan glacial maximum around 18 ka (Tchakerian, 1989). Stage IV sediments are represented by Qe3. However, Qe3 shows evidence of significant fluvial activity in the form of cut-and-fill structures. It contains the highest proportion of coarse sediments with an average mean grain size of 1.76 phi. It is probable that Qe3 is a fluvially redistributed dune sand or wind-winnowed fluvial sand, most likely representing a

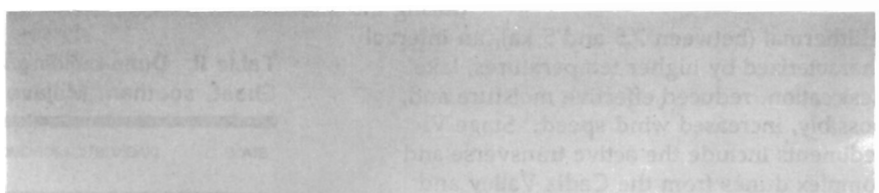


Figure 3. Sand ramp and dune wadi marked X in Fig. 2. Sheephole Mountains in background. Exposed samd ramp here is about 20 m in depth.

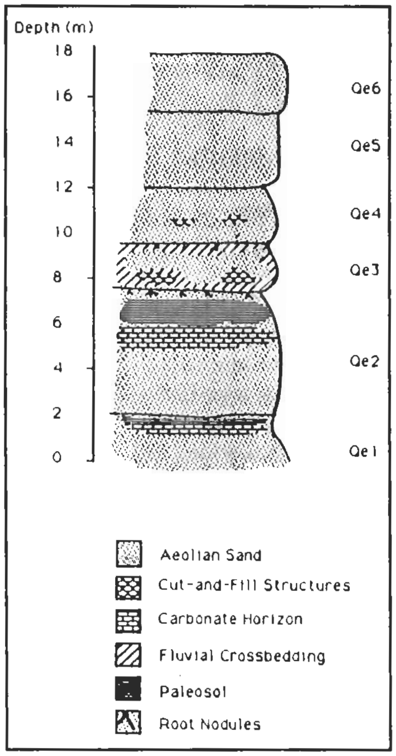


Figure 4. Composite stratigraphic column compiled from 4 detailed sections along dune wadi (Fig. 3).

transitional episode between Qe2 and Qe4. The æolian units in Stage V (Qe4, Qe5, Qe6) were most likely deposited during the Holocene. The uppermost unit, Qe6, yielded a cation-ratio date from rock varnish on the talus mantling it of about 5 ka (Dorn and others, 1986). Additionally, as seen in Fig. 5, there is considerable overlap between Qe4, Qe5, and Qe6, indicating their similarities with respect to quartz grain microfeatures and, thus, most likely represent pulses within a major depositional event. Based on regional evidence, it is likely that Stage V sediments represent a major Holocene depositional episode that took place during the Altithermal (between 7.5 and 5 ka), an interval characterized by higher temperatures, lake desiccation, reduced effective moisture and, possibly, increased wind speed. Stage VI sediments include the active transverse and complex dunes from the Cadiz Valley and represent renewed aeolian activity over the last few thousand years. The sediments consist of fine sands, are very well sorted, and show negligible amounts of silt and clay. Furthermore, the quartz grains from the crests of the transverse and complex dunes in Cadiz Valley exhibit surface micromorphologies indicative of active æolian processes, such as high percentages of upturned plates and meandering grooves and ridges (Tchakerian, 1989). The sand dunes contain well preserved primary sedimentary structures and show no soil or calcium carbonate development.

A preliminary study of the æolian stratigraphy in a sand ramp at the eastern edges of Iron Mountain (about 30 km east of the Dale Lake Sand Sheet) also indicates multiple episodes of æolian deposition (V. Tchakerian, unpublished field notes). Up to three æolian depositional pulses separated by paleosols have been tentatively identified. The sand ramp is veneered by rock talus from the mountains and entrenched by several ephemeral streams, and contains dune wadis similar to those found in the Dale Lake Sand Sheet.

A tentative evaluation of the absolute dates of the æolian sediments at the Dale Lake Sand Sheet

Table I. Mean values of grain size and sorting (a) and percentage sand (b) for Dale Lake Sand Sheet sand samples.

(a)

Aeolian Unit	Mean	Standard Deviation	Skewness	Kurtosis	Percent Silt and Clay
Qe1	2.24	0.83	0.10	1.17	2.90
Qe2	1.91	0.91	0.14	1.15	3.10
Qe3	1.76	1.11	0.05	1.06	1.55
Qe4	2.04	0.77	-.22	1.05	2.55
Qe5	2.23	0.78	-.07	0.93	2.45
Qe6	2.08	0.75	-.15	0.95	2.40
Calumet Mtn	2.91	0.97	0.22	1.29	4.83
Cadiz Dunes	2.35	0.29	-.03	0.89	

(b)

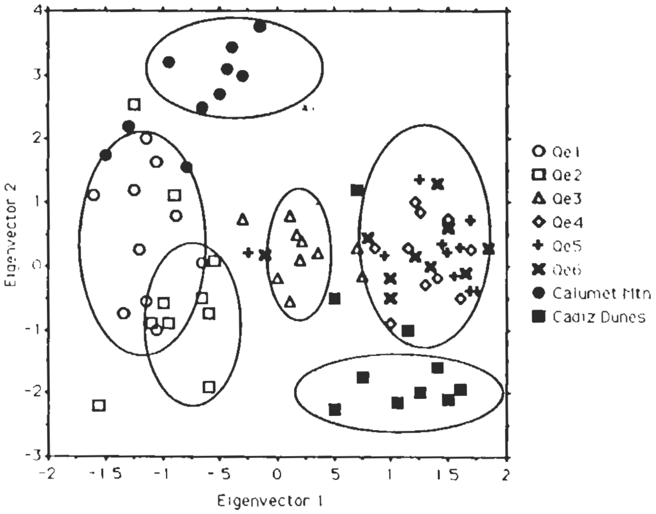
Aeolian Unit	Very Coarse	Coarse	Medium	Fine	Very Fine
Qe1	01.46	08.14	41.75	40.80	08.03
Qe2	04.46	09.13	35.45	41.90	07.06
Qe3	11.31	26.95	32.86	22.14	05.50
Qe4		17.97	47.50	29.35	04.26
Qe5		04.50	47.95	30.35	06.78
Qe6		09.16	48.70	37.64	03.63
Calumet Mtn	01.55	08.80	22.50	48.15	15.23
Cadiz Dunes			35.45	63.29	02.67

Table II. Dune-building episodes from SEM analysis, Dale Lake Sand Sheet, southern Mojave Desert, California.

STAGE	DIAGNOSTIC MICROMORPHOLOGIC CHARACTERISTICS	GEOMORPHOLOGICAL CHARACTERISTICS
VI	Upturned plates dominant; mechanical features predominate; low frequencies of chemical weathering microfeatures; angular outline; low relief	Active transverse and complex dunes, Cadiz Valley
V	Chemical features more abundant than mechanical; upturned plates abundant (45%); rounded outline; medium relief	Units Qe4, Qe5 and Qe6; sand sheet surface veneered with rock talus and entrenched by ephemeral streams (dune wadis)
IV	Chemical features dominant, especially solution etchings and silica plastering; upturned plates low to medium (25-30%); rounded outline; medium to high relief	Qe3? most likely representing a transitional episode between Qe2 and Qe4
III	Chemical features dominant, especially adhered particles, pits, solution etchings and silica plastering; upturned plates low (20%); rounded outline; high relief	Qe2, paleosol on top of Qe2; calcium carbonate horizon (stage II-III) below the paleosol layer with prominent root nodules
II	Chemical features dominant, especially solution etchings, pitted surfaces and silica plastering; upturned plates less than 20%; rounded outline; high relief	Qe1, paleosol on top of Qe1; Stage II carbonate horizon below paleosol layer
I	Chemical features predominant; large pits and sutures; silica plastering; upturned plates about 15%; rounded outline (50%); angular (45%); high relief	Stabilized climbing and falling dunes from the Calumet Mountains; highest silt and clay contents; Fe coating on grains; located on high mountains downwind from playas

Figure 5. Scatterplot of eigenvectors I & II from the principal component analysis. Over 50% of the variance can be explained by the first two eigenvectors.

Note the clustering of Qe4, Qe5, & Qe6 as one group. Clustering of the remaining aeolian units is similarly striking.



by thermoluminescence (TL) and infra-red stimulated luminescence (IRSL) techniques and cation-ration (CR) and accelerator mass spectrometry (AMS) radiocarbon dates from rock varnish from stabilized talus deposits, suggests that the sequence was deposited between 60 and 5.5 ka (Dorn and others, 1989; Lancaster and others, 1992). Current work by the author in collaboration with N. Lancaster (Desert Research Institute), A. Wintle (University of Wales), and H. Rendell (University of Sussex), using primarily luminescence dating methods, aims for establishing an absolute chronology of æolian deposition in the Mojave Desert.

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Cultural Resources at Dale Dry Lake

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INTRODUCTION

Dale Dry Lake is located in the south-central portion of the Mojave Desert almost immediately north of the boundary between the Mojave and Colorado deserts. The Transverse Ranges, which mark the border between the two deserts in this area, are directly south of the valley in which Dale Lake lies.

The lake is in a high desert valley surrounded by the Sheep Hole Mountains to the north and east, the Bullion Mountains to the north, the Pinto Mountains to the south, and the Little San Bernardino Mountains to the south and west (Fig. 1). It is separated from the west half of the valley and the Mesquite Lake–Twentynine Palms area by Valley Mountain, rising 2,311 feet above sea level and about 1,000 feet above the valley floor. The lowest point of Dale Lake is 1,164 feet above sea level; the beach is about 1,180 feet above sea level.

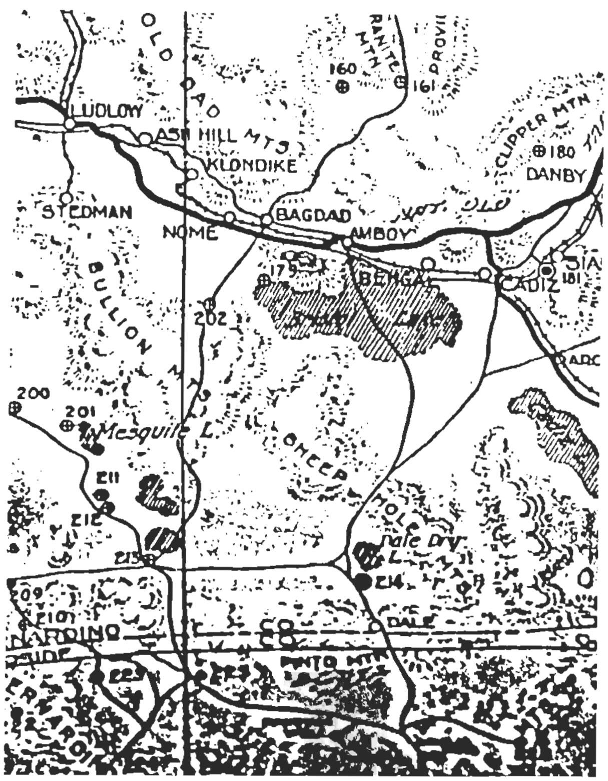


Figure 1. Location of Dale Dry Lake, from AAA (1914).

The location of the lake with respect to the Pinto Basin area and Joshua Tree National Monument make this a potentially rich area for archaeological resources. Very little is known about the prehistory of the area north of Joshua Tree National Monument and its connection with surrounding areas. It can be assumed that the prehistoric natives used the area when the lake contained water; however, there are no studies indicating when that may have occurred.

Environment

The present vegetation generally follows the old beach line. The lake is surrounded by active dune systems, and the shore line is covered by more or less level, loose sand exhibiting numerous small washes. The area is subject to flash floods and, along the northern perimeter, numerous trenches have been dug to channel the rain water to the lake bed and around structures. This modern earth-moving activity has greatly disturbed the natural contours and the surface of that portion of the Dale Dry Lake shoreline, and any prehistoric sites in the general area were probably destroyed by that activity.

Vegetation on the lake playa is sparse and is restricted to a few bushes next to modern wells. Around the lake perimeter, the Alkali Sink Scrub and the Creosote Bush Scrub Plant Communities are both present (Ornduff 1974). The vegetation of the Alkali Sink Scrub Community occurs predominately in areas where the loose sand has blown or washed away, leaving an alluvial deposit with a hard crust. The Creosote Bush Plant Community occurs predominately in those areas where sandy deposits remain and where small, sandy swells occur. Along the northern shoreline, some Palo Verde and other hardy trees occur along the perimeter of an old pond and its run-off channel. A desert tortoise was seen in this same area. Palo Verde is also present in a few of the larger, natural channels along the southern perimeter of the lake. To the west away from the lake shore, sand dunes occur with mesquite trees interspersed among them. Animal life observed in the area was restricted to jackrabbits and various lizards, but bighorn sheep and deer are probably present in the surrounding mountainous areas.

In its present condition, the lake perimeter has little to offer in terms of subsistence for prehistoric populations; however, with water in the lake, lake shore plants would have abounded and the lake would have provided fresh water clams and other invertebrates, possibly brine shrimp, and a haven for migrating birds. In addition, more and varied terrestrial mammals would have been present. The mesquite beans, from the trees in the western sand dunes, would be the most important resource in the area, with or without water in the lake (Schroth 1987). The water table is fairly high and fresh (but brackish) water can still be found by digging in sand dunes away from the lake bed proper. Thus, the area would have provided many of the resources needed to subsist in the desert environment.

PREHISTORY OF THE REGION

The earliest arrival of humans in San Bernardino County has not been determined. Some archaeologists advocate a very early occupation (before 20,000 B.P.), but the evidence for this is sparse and unconvincing (cf. Bada et al. 1974; Minshall 1976; Simpson 1980; Childers and Minshall 1980). It is generally accepted, however, that by the late Pleistocene-early Holocene (ca. 10,000-8,000 B.C.), the desert area was inhabited. Basically, the prehistory of the region can be divided into four temporal spans, based on the primary hunting weapon and the points used for the weapon. During the first span, hunting was accomplished using the thrusting spear with Lake Mojave, Silver Lake, and fluted points. Although thrusting spears continued to be used throughout prehistory, the points were replaced with those of the Humboldt series. The introduction of the dart and atlatl began the Archaic, which lasted about 8,000 years. Points used on the dart include those from the Pinto and Elko series, Gypsum, and some Humboldt. This period was followed by an Intermediate Period, during which points became smaller with the introduction of the bow and arrow. During the last prehistoric period, the Late Period, the bow and arrow had completely replaced the dart and atlatl and the projectile points were small, leaf-shaped, Desert Side-notched, and Cottonwood Triangular points.

The Big Game Hunters and Thrusting Spears

The prehistoric complex related to this inhabitation was given the name "Lake Mojave" by Warren (1984), "Western Lithic Co-tradition" by Davis (1969), "San Dieguito" by Rogers (1939), along with various other names. Sites from this complex consist almost completely of surface scatters of flake and core tools and lack tools for seed preparation. The artifact assemblage related to the tradition includes large ovate scrapers and bifaces, Lake Mojave and Silver Lake points, and large lithic flakes. The few fluted points found in the Mojave Desert may be related to this complex. Diagnostic artifacts, along with the points, include crescents and eccentric crescentrics (Rogers 1966). Because of the lack of tools for seed processing, the related cultures are believed to have relied primarily on hunting for subsistence, with the thrusting spear as the primary weapon. The reliance on hunting alone as a means of subsistence is being questioned and more evidence for the use of seed-grinding equipment during this period is being amassed (Warren, personal communication 1991).

The Archaic Period

The Pinto Period as defined by Warren (1984) derives its name from points from the Pinto series as described by Amsden (1935), Rogers (1939), and Harrington (1957). Sites related to this period are small, such as described for the Pinto Basin by Campbell and Campbell (1935), having little or no midden, with the exception of the Pinto site at Little Lake, Inyo County (Harrington 1957). The artifact assemblage consists of large scrapers (called by Warren [1984] "heavy-keeled scrapers", Rogers' [1939] "pulping planes", and Hunt's [1960] Death Valley II "scraper planes"), knives, scrapers, and choppers. The few slab milling stones found at these sites have been postulated to be pulping platforms, although some manos have also been recovered.

Warren (1984) has postulated that the Pinto complex evolved from the earlier hunting complex. The assemblage was seen by Warren as an adjustment to the more arid conditions that mark the end of the Pleistocene. Along with technological changes, Warren suggested that the population moved to the periphery of the deserts and to oases. According to Warren (1984), wetter conditions around 4,500 B.C. led to a reoccupation of the desert region by small groups of highly mobile hunting/gathering peoples. Around 3,500 B.C., a dry period began which necessitated another withdrawal to the desert periphery and to the few active oases. Warren (1984) postulated that from that time until 2,000 B.C., the desert was basically uninhabited.

Warren (1984) delineated the ensuing period, the Gypsum (ca. 2,000 B.C. - A.D. 500), by the occurrence of Humboldt concave base, Gypsum Cave, Elko-eared, and Elko Corner-notched points. The assemblage included leaf-shaped points, rectangular-based knives, flake scrapers, T-shaped drills, large scraper-planes, choppers, hammerstones, manos, and milling stones, with the mortar and pestle introduced during the latter part of the period. In addition, shaft straighteners, incised slate, sandstone tablets and pendants, bone awls, and *Haliotis* and *Olivella* ornaments were included in the assemblage. From Newberry Cave, perishable objects not found at open sites were recovered. These included atlatl parts, feathered plumes, sandals, S-twist cordage, tortoise shell bowls, and split twig figurines (Smith et al 1957; Davis 1981).

Warren (1984) believed that this complex derived from the Pinto complex since many of the tools were similar. He saw it as a complete adaptation to the desert environment involving technological change, religious constraints, and an expanding trade network. The use of the mortar and pestle was equated with the use of mesquite.

This dependence on projectile points to determine temporal periods within the Archaic of the Desert West has been questioned by Aikens (1970) and by Wilke and Flenniken (1990). Radiocarbon dates in association with Elko series points and Pinto series points do not substantiate a differentiation of two periods within the Archaic (ca. 8,000 B.P. to A.D. 500). Instead, a functional approach to the classification of points has been suggested (Flenniken and Wilke 1989). The point types used to assign sites to either the Pinto or Gypsum periods have been found to co-occur in many sites, particularly along the eastern portion of Nevada. Based on this information, and replicative studies by several people, Flenniken and Wilke (1989) postulated that all of the points of the Archaic were interrelated, with Pinto series points, Gypsum points, Humboldt concave-based points, and some Elko points the result of rejuvenation and reworking of two proto-types: The Elko Side-notched point and the Northern Side-notched point. Thus, the points can be viewed as part of the hunting system, rather than as temporal designation. The increase in milling stones and manos reflects a greater dependence on hard seeds.

In addition, recent environmental data suggests that the periods of dryness and high temperatures postulated by Antevs (1938) were less drastic than previously believed. According to Mehringer (1977), the Holocene climate never reached such extremes that the desert had to be vacated. Thus, the two periods postulated by Warren, the Pinto and the Gypsum, can be subsumed under one long (ca. 6,000 B.C. to A.D. 500) Archaic Period in which the dart and atlatl were the primary hunting weapons.

The Intermediate Period

Following the Archaic, point types began to reflect the introduction of the bow and arrow and became progressively smaller. Sites from this period are termed the Saratoga Springs complex (ca. A.D. 500 - 1,200) and are defined by the occurrence of Rose Spring and Eastgate points although some Elko and Humboldt series points may still be present (Warren 1984). Again, no firm dating of this period is known and the span given by Warren must be treated with caution (Yohe, personal communication 1992). These sites generally reflect the influence of Southwestern complexes, primarily defined by the pottery types present at the sites. During this period, the Anasazi from northeastern Arizona and southeast Nevada mined turquoise in eastern San Bernardino County. In the Colorado Desert, sites often display the influence of the Hatakaya from western Arizona. In addition, more and varied coastal shell artifacts began to appear. This period, according to Warren, was marked by an expanded trade network and regional developments; however, the subsistence and settlement practices remained basically the same as in the previous period.

Late Prehistoric

The last prehistoric period (ca. A.D. 1,200 to Historic) saw a continuation of the regional development and a continuation of the influence from the Southwest (Warren 1984). Although the influence from the Anasazi declined, that of the Hatakaya increased and expanded, possibly reaching as far north as the central Mojave Desert. The bow and arrow had completely replaced the atlatl, with Desert Side-notched and Cottonwood Triangular points predominating. Pottery from these sites consists primarily of buff and brown wares. During this period, the regional affiliations of the native Americans, as defined by early ethnographies, became finalized.

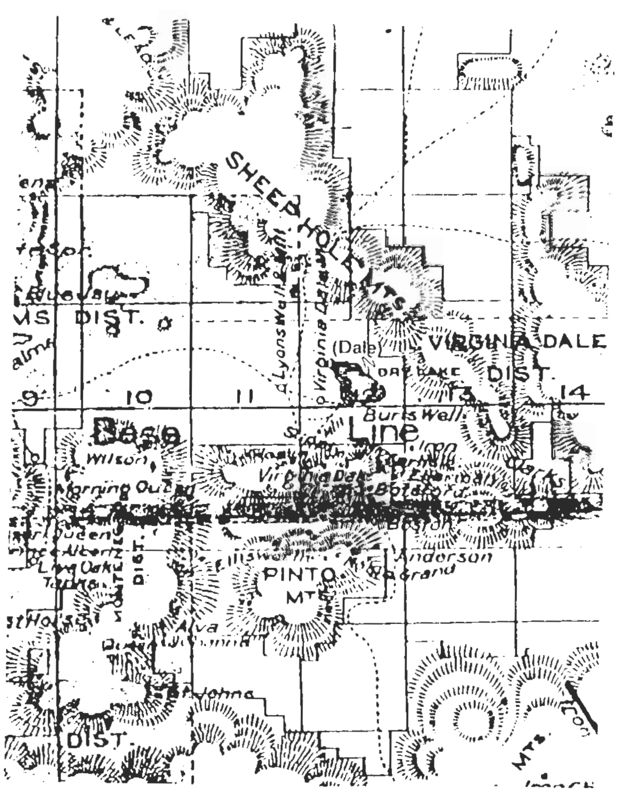


Figure 2. 1896 map (from Levet) showing location of wells around Dale Dry Lake.

Protohistoric Occupation

The native American group in the immediate project area during the protohistoric and historic periods was the Serrano (Bean and Smith 1978). European influence on the Serrano was negligible until 1834 when most of the natives were moved to an asistencia near Redlands. Prior to that time, according to Strong (1929), they were organized into local family linages. Each local linage retained a small, local territory but the Serrano has a whole did not occupy a tribal territory. They subsisted primarily by hunting and gathering, with settlements generally along the foothills and near permanent or semi-permanent water supplies. Early accounts of the European settlement of nearby Twentynine Palms describe meetings with small bands of natives who used the oasis as a stopping/watering place in their seasonal round.

HISTORY OF DALE DRY LAKE

According to C. Erickson (personal communication, 1986) of the Twentynine Palms Historical Society, no historical structures are known to be located immediately adjacent to Dale Dry Lake. The Plat Map from the 1855-1856 survey for the Department of Interior does not show any structures in the area of Dale Dry Lake.

Historically, the valley within which Dale Lake is located remained range land until the early 1900s, when homesteaders began to enter the valley. Two adobe foundations, presently located in the area noted as Bush along the northeastern margin of the lake, may belong to this historical period. The gold mining districts in the mountains to the south were of more importance. Although the first American discoveries of gold in that area were in 1873 (Vredenburgh, Shumway, and Hartill 1981), the Spaniards had previously mined some gold from the mountains south of Twentynine Palms (O'Neal 1957).

The area of Dale Dry Lake did not become important until the need for water for the gold mines developed. In 1892 or 1893, John Burt and F. L. Botsford located several claims in the Pinto Mountains. Although a 1896 map does not name the lake, it shows a well, Burts Well, along the southern edge of the lake, which probably relates to these early mining activities (Fig. 2). The mining claims to the south were sold several times until bought by the Brooklyn Mining Company in 1901, who renovated the well at Burt's Dry Lake, later known as Dale Dry Lake (O'Neal 1957:64). A pipeline was laid from Dale Lake to the Brooklyn Mine which used about 2,000 gallons of water for every ton of processed ore (Vredenburgh, Shumway, and Hartill 1981:141). A pump station to the south of Dale Dry Lake, shown on 1914 and 1925 maps (Fig 3), probably was in operation during this period.

The town of Dale was located about four miles west of Dale Dry Lake. Sometime between 1896 and 1914, New Dale was built about five miles south of Dale Dry Lake (Fig. 3). New Dale was probably built to accommodate the miners in the mountains to the south.

A chemical company, located along the eastern edge of Dale Dry Lake, was formed in 1920 by Irvin Bush and

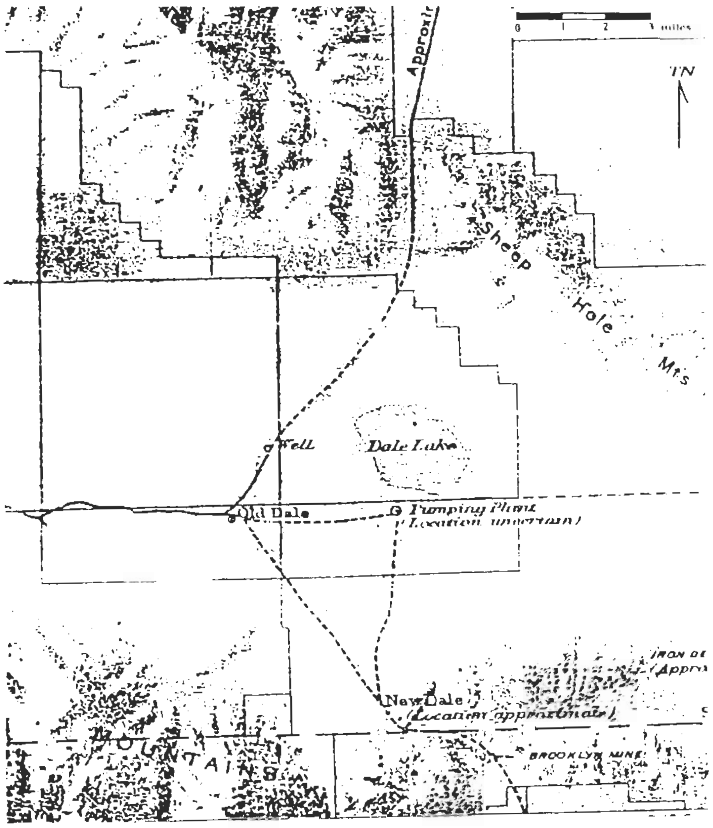


Figure 3. 1916 map showing locations of Old Dale and New Dale, from Thompson (1921).

continued in operation until 1949. The slabs and works from this chemical company remain in place (Fig. 4). This area has been recorded as the historical archaeological site, CA-SBr-5681H. It includes two adobe foundations and an old well that may date to the late 1890s. The cement slabs are probably related to the chemical plant as is an old pond located nearby.

An old water line consisting of an underground cement pipe runs for about 2.5 miles along the access road from the town of Bush to Amboy Road. It was last improved in the 1940s, at the same time that the access road was improved. For the other 0.5 miles, the water line is an above-ground metal pipe of undeterminable age.

HISTORY OF INVESTIGATIONS

Artifacts from the Campbell Collection

In the 1930s, William and Elizabeth Campbell began a long term research project on early man in California, centering their attention on the dry lakes (Campbell and Campbell 1937). They planned to visit every dry lake in California, collect diagnostic artifacts (primarily points), and write a report describing and comparing the assemblages. As part of that research, Dale Dry Lake was investigated.[1]

Interest in Dale Dry Lake began with the procurement of a collection from Mr. Stonecipher in 1930. This was a general collection described under Accession Number 498G125 on as "Dale Dry Lake, approx. 21 miles east of Twenty-Nine Palms" (page 125). The collection included "192 Miscellaneous pieces (flints, arrowpoints, knives, chips, etc.) Possibly Chemehuevi. Found on Surface encircling old Lake Shore" (page 125). A second entry for the same general area included "26+ arrowpoints (?) . . . Crudely, and well-shaped stones, intended for arrowpoints; some finished, some probable rejects" (page 125).

In 1931, E. F. Walker and his wife turned over collections from sites 428 and 429, located on the southwest shore of the lake. Site 429 was described as the larger site, located among the sand dunes. The lake was then visited by the Campbells who located Site 430, located beside a small sand dune approximately 1/8 of a mile from the northern shoreline. "Firestones" (fire-cracked rock) and ashes were also found at site 430.

In 1935, two new sites were found on the southwest corner of the lake by the Campbells: sites 748 and 750. Site 748 was located "near a mud hummock of self-rising ground about head high" (page 2691). Site 750 was described as a large site that "extended for quite a distance, had no potsherds or arrowpoints, and appeared to be a fairly old camp, judging from the weathering of the stone" (page 2697). It was located on spits extending out into the lake playa, with one metate found on the edge of the lake. That same year, site 749 was found by Walter Berg on the southeast corner of the lake. The latter site was described as a "camp on a low spit that extends out into the lake at this point" (page 2692).

The largest site, No. 431, was found by a group composed of the Campbells, the Walkers, and Craig Grover. Site 431 was thought, at first, not to be a shoreline site, but rather was located west of the lake in the area recognized today as an ACEC by the Bureau of Land Management. It was described

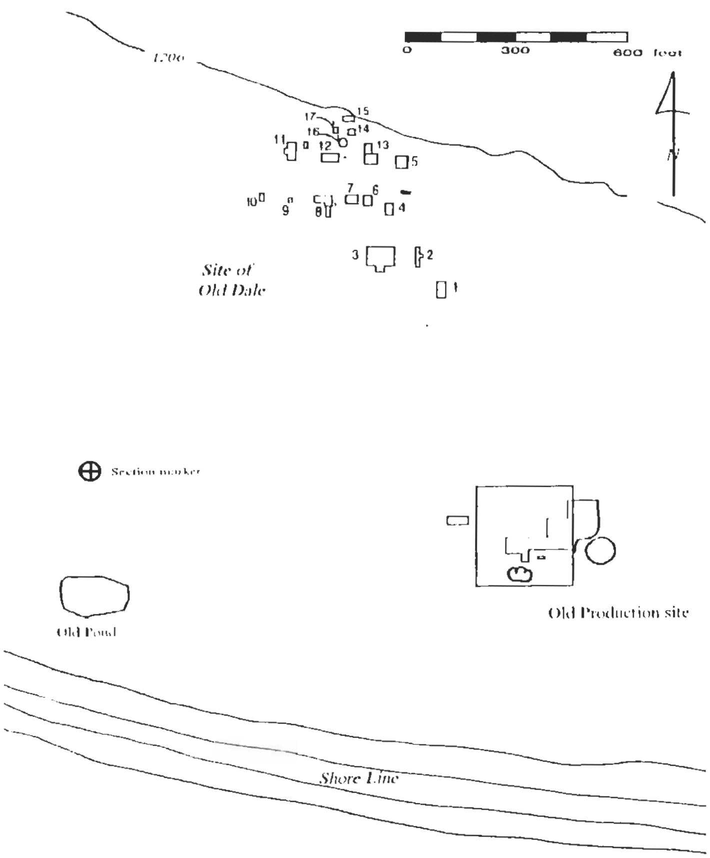
Table I. Summary of cultural material collected from Dale Dry Lake

	Campbell Site Number								
	75	428	429	430	431	748	749	750	Totals
Arrowpoints	26	--	--	2	7	--	--	--	35
Atlatl points	--	--	--	--	26	--	--	--	26
Bone Awl Point	--	--	--	--	1	--	--	--	1
Chips	--	--	--	--	10	--	7	23	40
Core	--	--	1	--	--	1	--	--	2
Crystals	--	--	--	--	1	--	--	--	1
Drill	--	--	--	--	--	--	--	1	1
Hammerstones	--	--	10	--	14	--	--	--	24
Knife	--	--	1	--	1	--	1	--	3
Manos	--	1	5	4	--	--	1	1	12
Metates	--	--	2	1	--	1	--	1	5
Misc. pieces	125	--	--	X[a]	X	--	--	--	125
Net sinker	--	--	--	1	--	--	--	--	1
Other points	--	--	--	--	1	--	--	--	1
Polishing stone	--	--	1	--	--	--	--	--	1
Potsherds	--	--	--	TX[a]	X	--	T	--	0
Scrapers	--	--	3	--	14	--	--	2	19
Turtle-back	--	--	1	--	--	--	--	--	1
Totals	151	1	24	8	75	2	9	28	298

[a] X - In catalog but no total given. T = Noted as present on the site, but not collected.

Figure 4. Historic district of Bush, adapted from Mine Mill Plot Plan, 1986, D. Grott.

#1: part of old salt works; #2: salt works trench; #3: salt works slab w/ trench; #4: graded away; #5-7, 10-15: cement slabs; #8: cement slab w/ storm shelter; #9: adobe location (no slab); #16: well location; #17: cement slab w/ adobe.



as several campsite scattered among mesquite-covered sand dunes that the Campbells felt could mark a shoreline of the old lake. This site was visited again in 1933 and in 1935 by the Campbells and finally in 1936 by a group known as the Campbell Expedition. Numerous items were collected during each visit. The Campbells notes indicate that some of the campsites contained potsherds, but many did not. The site was later described as extending from the shoreline to about one mile west of the shoreline, with the camps dispersed throughout the mesquite-covered dunes.

The assemblages collected from the sites and curated at the Southwest Museum and the Visitors Headquarters, Joshua Tree National Monument, are summarized in Table 1. At this time, no attempt was made to complete an in-depth analysis of the material; the artifact categories are those given by Elizabeth Campbell. In all, over 300 items were collected.

The majority of the collected items are points and "chips" or pieces of debitage. Lithic materials include obsidian, jasper, chalcedony, chert, basalt, jasp-agate, rhyolite, and metamorphosed rhyolite; however, the vast majority are vein quartz or crystalline quartz. Point types range from Pinto series points to small Desert Side-notched and Cottonwood Triangular points. One point may be a variant of Silver Lake. Many of the "knives" discussed by the Campbells appear to be biface cores and preforms. The "crude points" (arrow and dart) include numerous preforms and rejects. In addition to the biface cores, single platform, unidirectional cores are present. Core tools in the collection include hammerstones and scraper planes. One "thumb-nail" scraper of jasper was collected as were several other small scrapers and a few drills.

Unusual artifacts collected from Dale Dry Lake include a net-sinker and quartz crystals. One small metate that may have been used for grinding medicinal plants, tobacco, or ochre was collected. Some of the manos are simple used cobbles; others are shaped and smoothed on all sides and edges. None of the larger metates are still in the collections.

Recent Archaeological Investigations

One prehistoric site was located along the northern shoreline by R. Reynolds in 1970. Its location was given as between the northern edge of the Dale Lake salt evaporators and the access road. Its description included artifacts and projectile points. This site could be the Campbells' site 750.

Two surveys were conducted within 1.5 miles of the lake. One, by Sutton (1983), involved only the intersection of Amboy road with the road leading to Dale Dry Lake. The one site located during this survey was not recorded in depth, but appears to be a Late Prehistoric camp site. During the second survey, four sites were located, but none are close to the dry lake. These sites consist of lithic scatters and have been defined as small camps. They are located to the west of the lake in an area denoted as the Dale Lake ACEC, according to the Bureau of Land Management Desert Plan, and have special archaeological significance (US Department of the Interior 1980). As noted before, in this area the Campbells collected numerous artifacts and noted the occurrence of several campsites scattered throughout the dunes (site number 431).

As a result of the 1986 field survey by the author, two historic locales and three prehistoric sites were located. One prehistoric site consists of a widely dispersed scatter of lithic flakes, bone fragments, shell fragments, one potsherd, and projectile points. It is located directly south of the lake. Lithic flake material include jasper, chalcedony, quartzite, and metavolcanics. Included in the bone is the partial mandible of a jackrabbit (*Lepus californicus*); all other fragments were too small or lacked identifiable characteristics. None of the bone appeared to be human and none of it appeared to have been burnt. The shell was fragmented to the extent that exact identification was not possible; however, it appeared, from field observation, to be fresh water clam (*Anodonta* sp.). Only one potsherd, probably Tizon brown ware, was observed. The projectile points included one Desert Side-notched point made from chert and a Cottonwood Triangular point base made from quartzite. This site appears to be Campbells' site 429.

Another prehistoric site, located on the southwest corner of the lake, was found in an area devoid of the loose sand. Instead, it was located on an alluvial deposit such as found in the playa but lacking the salty deposit on the top. It consists of three loci, each ranging in diameter from 5 to 10 m. The loci appear to be well-defined, concentrated lithic stations and each contained numerous flakes and one to three cores. Lithic materials included metavolcanics, volcanics, and quartzites. No cryptocrystalline quartzes were present. This may be the Campbells' site 748, found in an area of "mud."

The third prehistoric site was another concentrated, small site which also probably represents a lithic flaking station. It was located approximately 90 m south of the shoreline. It was also on an alluvial deposit rather than loose sand and contained the same types of lithic flakes as the previous site; however, no cores were present. This appears to be the Campbells' site 428.

Historic Resources

Previous markers from the 1920s-1940s include rock cairns at the corners of 20 acre parcels. These cairns are not to be confused with similar prehistoric features. Most of the cairns are on the lake playa but some exist above the beach line and can be recognized as historic property boundary markers by their position in a true N-S or E-W line with other historic plot markers.

The town of Bush has been official recorded as an historic site. It covers an area of approximately 10 acres and is located along the northeastern edge of Dale Dry Lake. There are 17 mapped cement slabs or structure platforms which includes some of the slabs for the salt mining operation (Fig. 5). One structure is a deteriorating cement slab covering a large, stone-lined well (Fig. 5; Number 16). According to D. Grott (personal communication, 1986), two of the structures were made from adobe, and the rest had cement blocks for walls. The most probable locations of the adobe structures were located in the field (Fig. 5; Numbers 9 & 17).

A second historic locus was located to the southwest of Dale Lake in an area of sand dunes. It appears to be connected with the gold mine district of Dale in that an old bottle, dating to between 1900 and 1920, was found at the site (Lorrain 1968). Other historic artifacts include round nails, wood, rusted metal pipe, and wire. This may be the site of the original well, Burts Well, discussed above.

SUMMARY

Dale Dry Lake has been important throughout the occupation of the area by humans, prehistorically as a water source and a subsistence base, and, historically, as a water source and as a chemical plant. The point types found by the Campbells suggests that whenever there was sufficient water in the lake to maintain a population, people used the resources. The presence of *Anodonta* shell, found in the 1986 survey, implies that at least once there was water in the lake that remained long enough to allow for a shellfish population to survive, and that prehistoric natives used that resource. The "sinker" found at site 430 may indicate the use of nets in a lake. It was found with Desert Side-notched points and ceramics.

The numerous camps found by the Campbells to the west indicate a long tradition of the use of the mesquite. Ceramics were found at some of the sites along with small points, but some Pinto points also were found in campsites lacking ceramics and arrowpoints.

After the Anglo settlement in the region, the lake became an important water source and later a chemical plant site (sodium). The water was used in the mining operations in the Pinto Mountains for almost 50 years, and the lake bed is still being used for extraction of sodium.

ACKNOWLEDGEMENTS

This paper was first written as an archaeological assessment conducted by the Archaeological Research Unit (ARU), University of California, Riverside, on 1,840 acres located at Dale Dry Lake and on a three mile stretch of access road leading from Dale Dry Lake to Amboy Road in San Bernardino County (Schroth 1986). The survey was undertaken at the request of Dave Grott for the Western America Ore Co. and was performed by the author under the direction of Dr. P. J. Wilke (UCRARU Project No. 847). The maps used herein were reproduced using the Archaeological Research Unit computer system.

Lester Ross, of the San Bernardino Archaeological Information Center, provided additional information and the copies of the historic maps used in this report. His support was invaluable. Rosie Pepito, Curator at the Visitors Headquarters, Joshua Tree National Monument, allowed access to that portion of the collection by the Campbells. I thank her for her support and the time she spent tracing the information. George Kritzman, Curator of the Southwest Museum, found and shared the portion of the Campbell collection at that museum, and his help was greatly appreciated. Richard Cerreto read a draft of the paper and his comments were greatly appreciated.

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Bristol Lake Basin—A Deep Sedimentary Basin along the Bristol-Danby Trough, Mojave Desert

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ABSTRACT

Bristol Lake basin, midway along the 200-km-long Bristol-Danby trough, contains the deepest sedimentary fill anywhere along the trough. Northwest-striking faults near the basin show latest motion in the early or middle Pleistocene, whereas faults farther west moved as recently as the Holocene. A method of gravity modeling that separates the gravity field into "basement" and "cover thickness" components was used to define the shape of the sedimentary fill in the basin and estimate its depth. The gravity model indicates the basin is elongate northwest and is sharply bounded from adjacent basins.

STRUCTURAL AND SEDIMENTARY SETTING

The Bristol-Danby trough (or Barstow-Bristol trough) is an enigmatic structural depression about 200 km long that crosses the Mojave Desert along a west-northwest trend (Thompson, 1929; Bassett and Kupfer, 1964; Gardner, 1980; Glazner, 1981). It truncates northwest-trending ranges on either side. The origin of the trough remains unknown, but many of the neighboring ranges are known to be bordered by late Cenozoic strike-slip faults related to the San Andreas transform fault system (Dibblee, 1961; Dokka and Travis, 1990).

Several of these northwest-striking faults show evidence of Quaternary movement in the region surrounding Bristol Lake. Faults that last moved in the early or middle Pleistocene are present in ranges close to Bristol Lake (Fig. 1). The age of the youngest faulting decreases to the west, so that, for example, the Valley Mountain fault system cuts alluvium assigned to the Holocene (Howard and Miller, 1992).

Lowlands along the trough include several closed depressions occupied by playas. One of the most conspicuous is the Bristol Lake playa at an elevation of 180 m, midway along the trough. The nearby Cadiz Lake playa lies slightly lower, at 165 m, and represents the regionally lowest outcrop of the groundwater table. Both basins contain thick upper Neogene and Quaternary deposits extending below sea level (Bassett and others, 1959; Rosen, 1989). The presence of these young basin fills and the closed topographic depressions indicate to us that the basins formed by youthful tectonism. Ranges surrounding the basins expose largely pre-Cenozoic plutonic and metamorphic rocks, and lesser amounts of Miocene volcanic and sedimentary rocks (Bishop, 1964; Miller and others, 1982; Howard and John, 1984).

Drill cores dated by tephrochronologic techniques revealed dipping beds and strong evidence for a local surge in sedimentation rate at about 2 Ma in deposits beneath Bristol Lake, compared to a thinner section with a near-constant sedimentation rate under the west margin of the basin; the dipping beds and sedimentation surge were attributed to

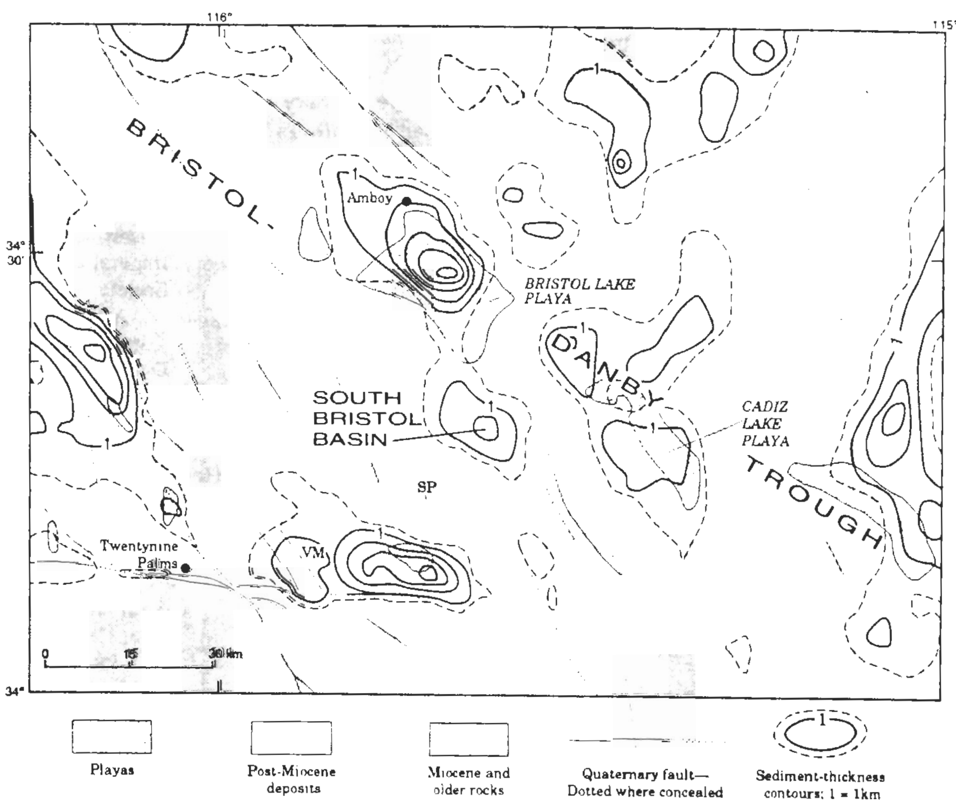

Figure 1. Geologic sketch map of the Bristol Lake area showing contours of Cenozoic deposit thickness as derived from gravity analysis (see text). VM = Valley Mountain faults; SP = Sheep Hole Pass.

deepening of the basin interior by faulting (Rosen, 1989). Rosen found that 270 m of cored sediments at the margin of Bristol Lake underlie an ash bed at 200 m depth correlated with a 3.7-Ma tephra. Extrapolated sedimentation rates below the ash bed suggested to Rosen that the deepest cored sediments may exceed 6-10 Ma in age and deeper undrilled sediments require an even older age for initiation of the basin. Persistent playa environments and lack of long-lived lakes since at least 3.7 Ma (Rosen, 1989), and incision and modern capture of streams in the Bristol Mountains (Miller, submitted), suggest that subsidence of the basin began by the Pliocene or earlier and continues on at present.

GRAVITY MODEL

An earlier analysis of gravity and aeromagnetic anomalies in the area indicated that sedimentary fill in the Bristol Lake basin may reach 1.6 km in thickness under the southeast margin of Bristol Lake (Simpson and others, 1984). Here we present a new analysis of gravity data which models the thickness of sedimentary fill in the Bristol Lake basin and surrounding area in three dimensions.

Figure 1 shows our map of thickness of sedimentary deposits as derived from the isostatic residual gravity in the Bristol Lake area. The map was prepared using an iterative procedure described by Jachens and Moring (1990), in which the gravity data, a knowledge of the surface geology, and an estimate of the densities of Cenozoic deposits were used to separate the isostatic residual gravity field into two components, a "basement" component due to density variations within pre-Tertiary basement, and a "cover thickness" component caused by low density Cenozoic deposits. The gravity anomaly caused by the young deposits was then inverted to yield a map of the depth to pre-Tertiary basement. The basic gravity data are from Snyder and others (1982), Simpson and others (1984), and Mariano and others (1986).

A layered density model was applied in which the model densities of Cenozoic deposits increased from about 2.15 g/cm^3 near the surface to 2.42 g/cm^3 at a depth of 1.2 km. In practice, density contrasts of -0.40 g/cm^3 (0-200 m), -0.35 g/cm^3 (200-600 m), -0.30 g/cm^3 (600-1,200 m), and -0.25 g/cm^3 (below 1,200 m) compared to a nominal basement density of 2.67 g/cm^3 were used in all calculations. This density structure was chosen for the interior Mojave Desert basins on the basis of limited subsurface density information and a comparison of the inferred depths to basement resulting from this model and the actual depths to basement found by drilling in about 20 widely distributed locations. Drill holes that penetrate more than 1,200 m below the surface are rare in the central Mojave Desert and are not adequate to define the variation of density with depth in the deeper parts of the basins. Therefore, a constant density contrast of -0.25 g/cm^3 was assumed below 1.2 km. Because of the resulting uncertainties in the deeper parts of the basins, figure 1 shows a light depth contour at 0.5 km, a labelled depth contour at 1.0 km, and unlabeled deeper contours that represent 1 km intervals at a constant density contrast of -0.25 g/cm^3. These deeper contours can be viewed as the gravity counterpart to a seismic-reflection time-section in which the contours present the correct basin geometry even though the actual depths may be different from those shown.

The derived map (Fig. 1) indicates that modeled sedimentary fill thickens toward a locus under the southeast margin of Bristol Lake, and that the sedimentary basin is elongate northwest-southeast. The map suggests that abrupt changes in basin geometry occur between the basin beneath Bristol Lake and adjacent shallower sedimentary basins to the south (south Bristol basin) and southeast (Cadiz Lake basin). Structural boundaries evidently segment the basement surface under the Bristol-Danby trough in this area. The structures may be related to right-lateral faulting or to pull-aparts between blocks affected by right-slip (Miller and others, 1982; Dokka and Travis, 1990; Richard and Dokka, 1992). An alternate possibility is that the Bristol-Danby trough reflects a compressional buckling in the crust (Howard and Miller, 1992). Seismic-reflection profiles across the Bristol, south Bristol, and Cadiz Lake basins are presently under analysis by D. Okaya and others and may help to resolve the specific structures that bound the basins. Along the approximately 200-km length of the Bristol-Danby trough, the gravity modeling identifies the sedimentary basin under Bristol Lake as conspicuously the deepest.

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The Depositional Environment and Evolution of Bristol Lake Basin, Eastern Mojave Desert, California

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INTRODUCTION

Bristol Dry Lake is situated in the Mojave Desert region of southeastern San Bernardino County near Amboy, California (Fig. 1). It is the largest (155 km^2) in a system of three northwest-southeast trending dry lakes (playas) located in a structural trough between the Bristol Mountains to the north and the Bullion and Sheephole mountains to the south. This location is one of the most arid places in the United States. Bristol Dry Lake is filled with over 500 m of sediment in the basin center, of which 260 m is almost pure halite.

GEOLOGIC SETTING

Bristol Dry Lake is separated from Cadiz Lake by the coalesced alluvial fans of the Calumet Mountains to the south and the Marble Mountains to the north (Fig. 1). At present, both basins have completely separate internal drainage.

To the west, a third basin, Alkali Dry Lake (Fig. 1), is separated from Bristol Dry Lake by Amboy Crater, a relatively young cinder cone and basalt flow which is thought by Parker (1963) to be less than 6,000 years old. Recent work by the U.S.G.S. suggests that it is much older than this and may be on the order of 50,000 years old. Before the basalt flow blocked off the northern end of the drainage area, creating Alkali Dry Lake, this area drained into Bristol Dry Lake. Therefore, for most of the history of the Bristol Dry Lake basin, the total drainage area into the playa was significantly greater than it is today (about 4000 km^2). The present total drainage area into Bristol Dry Lake, excluding drainage from Cadiz Dry Lake and Alkali Dry Lake, is just over 2000 km^2.

Thompson (1929) proposed that during Pleistocene time a large lake occupied all of the Bristol and Cadiz basins. However, no evidence for the existence of such a large lake has been documented (Bassett and others, 1959). More recently, the Bristol Dry Lake basin has been interpreted to have alternated between subaerially exposed periods and times when shallow ephemeral water bodies covered the playa surface (Handford, 1982a). Overall, even during periods of regional high rainfall, evaporitic conditions seem to have been dominant. Modern meteorological conditions for the central Mojave Desert and Bristol Dry Lake specifically indicate a mean annual rainfall of less than 100 mm. Thompson (1929) noted that there are periods of 2-3 years when no precipitation has been recorded.

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DEPOSITIONAL ENVIRONMENTS

Handford (1982a, b) recognized four broad depositional subenvironments in Bristol Dry Lake. They are, from alluvial fan to basin center, 1) the alluvial fan, 2) the playa-margin sand-flat and wadi system, 3) the saline mud flat, and 4) the salt pan at the basin center. These designations differ slightly from the Hardie and others (1978) subenvironments, but are useful for overall descriptions of the playa geometry. The four subenvironments recognized by Handford (1982a) are used in this paper in order to be consistent with his work (Fig. 2). However, detailed field work indicates that these broad subenvironments are more diverse than previously described. The following descriptions detail further subdivision of the four broad depositional environments outlined by Handford (1982a) into more useful sedimentological units.

Alluvial Fan

The alluvial fan can be divided into three gradational subfacies based on the sediment grain size. The terms used here are proximal, mid, and distal fan.

Proximal Fan

The **proximal fan** includes the portion of the fan closest to the mountain sorce. At present, this area is dominated by channelized flow through arroyos incised through the older proximal and mid fan deposits. The older fan deposits consist of coarse-grained gravel, cobble, and boulder-size sediment interbedded with gravelly sands. Between the arroyos are cobble and boulder covered surfaces darkly stained with desert varnish.

Mid Fan

The **mid fan** area is dominated by coarse-grained braided stream deposits. Broad, shallow channels are filled with low-angle trough cross-stratified cobbles, gravels, and sand. Extensive box-work calcrete is present in poorly-sorted cobble strata of the upper part of the mid fan. The calcrete zone (K-horizon) is best exposed on the north side of the basin in the Bristol Mountain fan arroyos (directly north of Amboy). However, the calcrete zone is widespread throughout the fans. The calcrete zone is approximately 1-3 metres below the surface of the fan and it is overlain by a dark red paleosol. The paleosol is disconformably overlain by a thin 1-3 m uncemented layer of low-angle trough cross-bedded braided

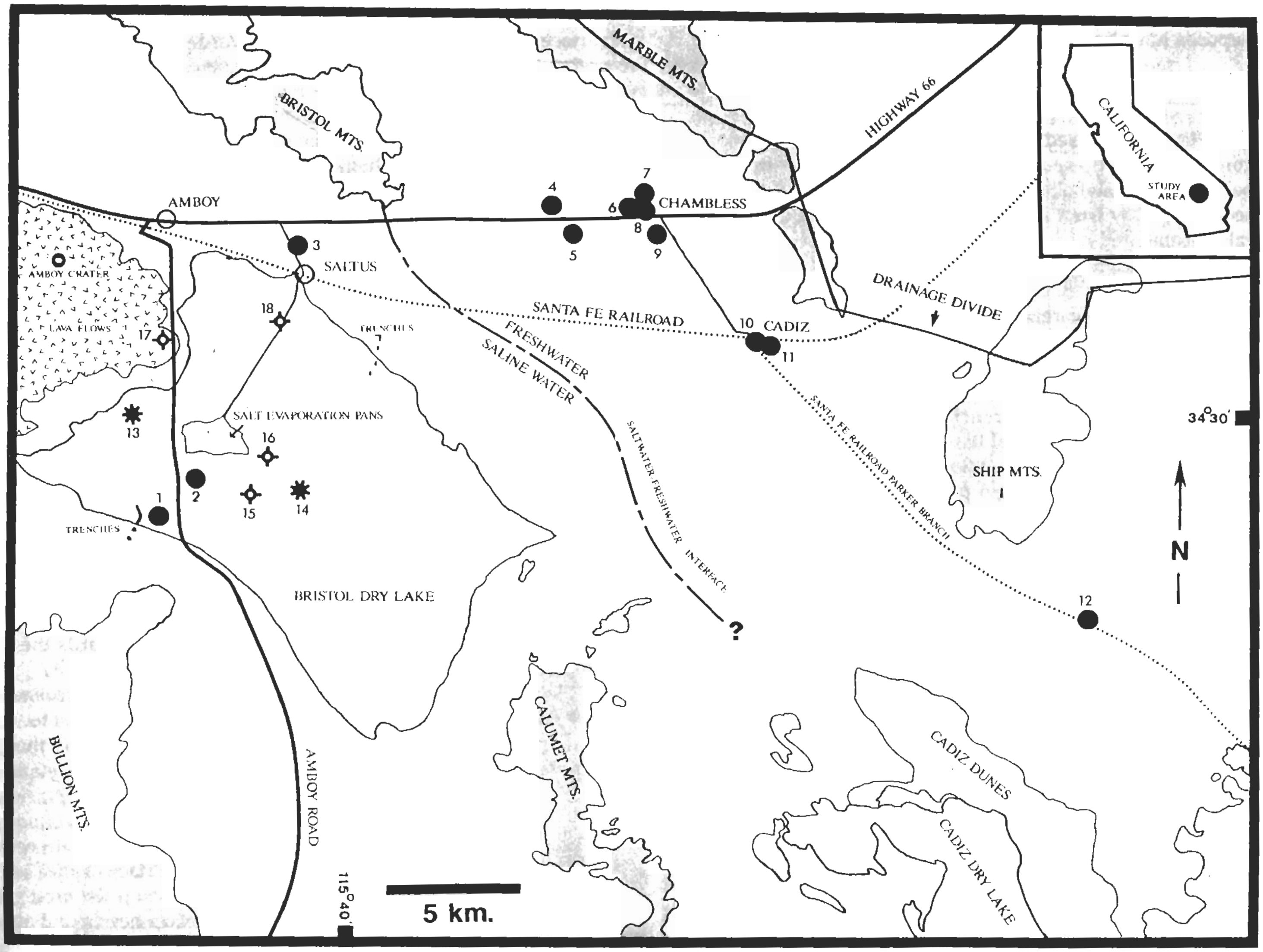

Figure 1. Bristol Dry Lake Basin modified from Rosen (1991) showing location of cores & surface sampling localities. Numbers refer to cores in Rosen (1989, 1991). CAES #1 is no. 17; CAES #2 is no. 16.

stream deposits.

Distal Fan

The **distal fan** is dominated by sheet flood deposits mostly composed of sand-size particles and finer. A thin veneer of the distal fan overlies, and grades into, the playa margin sediments. The well-developed paleosol and calcrete which extend up into the lower proximal fan are also exposed in the arroyo walls of the distal fan, approximately 1-3 m below the surface fan deposits.

In general, aeolian deposits are ephemeral in Bristol Dry Lake because the playa surface lacks sufficient moisture to trap sediment. Although the wind regime in the basin is seasonal, the dominant wind direction is to the southeast. Strong winds have plastered sand up onto the sides of the divide separating Bristol Dry Lake from Cadiz Dry Lake. In addition, a large dune field, with some dunes up to 15 m high, is migrating to the southeast directly down wind of the divide in the Cadiz Basin. This suggests that a great deal of the sand is transported out of the Bristol Dry Lake basin and over the divide into the Cadiz Basin. No unequivocally aeolian

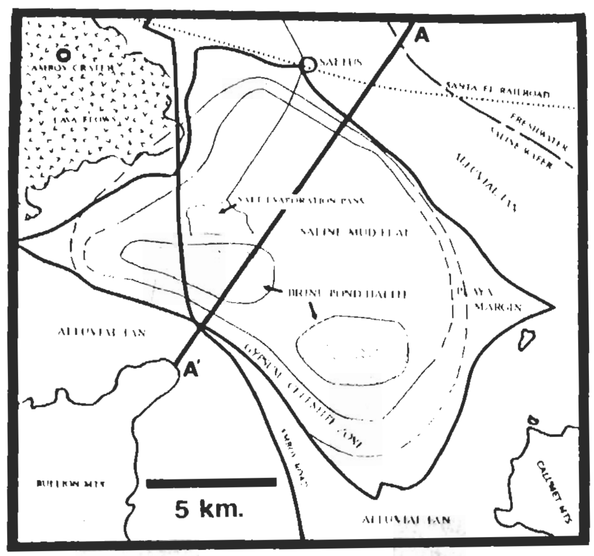

Figure 2. Surface facies distribution showing bulls-eye pattern of facies from alluvial fans to the gypsum-celestite zone of the play margin, to the saline mud flat and the halite salt pan in the basin center.

deposits have been seen in core.

Playa Margin

Playa margin sediments are deposited in a transitional zone vegetated by sparsely populated halophyte shrubs, between the distal alluvial fan and the saline mud flat. The sediments vary from silty sands to sandy muds that contain calcite-cemented nodules surrounding root holes of former halophyte shrubs.

Numerous wadi channels and distributary channels dissect the playa margin and bypass sediment and organic matter from the alluvial fans directly across the playa margin to the playa center. Some of the distributary channels are up to 1 m deep and 100 m across, and extend well out onto the saline mud flat. Shallow trenches across these wadis and distributary channels reveal flaser bedded sands and muds, as well as planar bedded, sheet flow sands.

Where the playa margin passes into the saline mud flat, a 0.15-0.5 km wide zone of gypsum and celestite is present in a lens-shaped body in a ring around the entire basin. The details of this zone are presented in Rosen and Warren (1989). Field and chemical data indicate that discharging groundwater in this zone is saturated with respect to gypsum and celestite. Vertically aligned lens-shaped gypsum crystals grow displacively in the sediment near the top of the groundwater table where the sediment is still water-saturated and easy to move. Where the groundwater table has been constant for some time, the precipitation of gypsum has been greatest. In this area, gypsum may account for up to 90% of the volume of sediment in trenches that are 2 m deep. The matrix surrounding the gypsum consists of a mixture of detrital siliciclastics (quartz, feldspars, clays, and a large suite of heavy minerals), ranging from sand to clay-size particles, and authigenic rhombic calcite less than 1 μm in size. Trenches that were excavated in 1988 are still present on the south side of the playa west of the road to Twenty-nine Palms (Fig. 1). Although they have collapsed significantly, they still show many of the playa margin features and are the only exposures of this facies in the basin.

Saline Mud Flat

The saline mud flat is dominated by mostly homogeneous detrital mud. The dominantly gray brown mud is oxidized to a reddish brown color at the surface. Individual mud units are not thick, usually less than 0.5 m, and some are capped by mudcracked surfaces. The modern saline mud flat is up to 5 km across, and at times in the past extended across the entire basin. In addition to the muds, some of the larger wadis and distributary channels extend into the mud flat, creating flaser bedded sand and mud out into the basin center. Aeolian processes also bring some detrital sand and mud out to the basin center as well as some detrital gypsum and calcite. In addition, millimetre-laminated authigenic calcite is precipitated along with the detrital muds when there is a sufficient water body in the basin.

The surface of the saline mud flat is hummocky and in many places water saturated. It has been variously described in the literature as "self-rising ground" and "puffy" (Thompson, 1929; Bassett and others, 1959; Gale, 1951). The "puffiness" is due to displacive halite growing just below the surface and pushing the sediment upward. Desiccation cracks and millemetre-think efflorescent halite and calcium chloride crusts are also common on the mud flat surface.

Where the groundwater table is below the surface of the playa, the water rises by capillary action to the surface and evaporates. Displacive halite crystals form in the sediment creating the "self-rising ground" mentioned above. The water that rises to the surface keeps the surface cohesive enough so that the sediment is not deflated. Cementation by halite in the "self rising ground" also helps retain sediment. However, where the surface lacks moisture because the groundwater table is too deep for water to rise to the surface by capillary action, the sediment dries out and is deflated. Deflation of the modern surface is particularly noticeable in the playa margin facies. Here, once buried displacive celestite and gypsum crystals are exposed on the surface of the playa.

Near the playa margin facies, the muds contain abundant molds of what were 5-10 mm displacive halite cubes. These molds are now empty but are rimmed with a manganese oxide stain. The molds form a network of cubic isolated pores in the mud which may be up to 35% of a given volume of mud. Towards the center of the playa, the molds are larger and are filled with halite. In the most basinward area of the saline mud flat, giant hopper crystals of halite can be found, some up to 0.5 m across. Although the crystals are larger towards the basin center, they are also less abundant.

Small (5-30 cm) calcite concretions also occur in the saline mud flat. Some of these concretions contain molds of halite crystals. These concretions are abundant at the surface in the saline mud flat, in the basin center, and in the cores.

Salt Pan

In 1988, there were two salt pans at the surface of the basin, one on the east side of the basin and one on the west (Fig. 2). According to local residents, water has been ponding in the east salt pan only since 1982. Both salt pans are about 0.4 m thick, composed of almost pure layers of 10-20 mm-size, vertically elongated chevron halite, separated by millimetre bands of detrital silt or mud. The halite is 99% pure, and both pans consist of approximately 30% chevrons, 40% clear diagenetic halite, and 30% porosity. Porosity decreases from 30% in the modern salt pan to 6% 3 meters below the surface.

The most striking feature of the salt pan is the large halite tepee structures, caused by the force of crystallization of the halite. They are up to 0.6 m tall and form in a regular polygon pattern reminiscent of mud crack patterns. However, after rainfall, sufficient water undersaturated with respect to halite, ponds on the salt pan and dissolves the tepees flattening the salt pan to an almost billiard table smooth surface. If there is enough rain to pond water that is undersaturated with respect to halite, 0.3 to 0.4 m diameter dissolution circular pits will form on the surface of the salt pan. Subsequent evaporation of the water and precipitation of new halite fills the pits with clear diagenetic halite and creates new tepee structures. Below the tepee crust of the halite is a water saturated mud which is just below the point of halite precipitation. The mud is thixotropic and structureless and there are no displacive halite cubes in the mud.

Although the salt pan halite is only tens of millimetres thick at the surface, 3 meters below the surface there is a 1 m thick halite bed. In some cores, relatively pure (80-90%) halite may be tens of metres thick. The halite beds in core have retained relatively little primary fabric. Although chevrons have been observed in the basin-center cores, they generally make up less than 2% of the fabric. Most of the subsurface halite is composed of interlocking centimetre-size crystals of clear equant halite. The amount of siliciclastic matrix mixed in with the halite varies from 3 to 50%.

Original data from CAES #1 and CAES #2 (Rosen, 1989); original data from USGS Bristol Core 2 (Bassett et al. 1959). (Figure from Brown and Rosen, submitted).

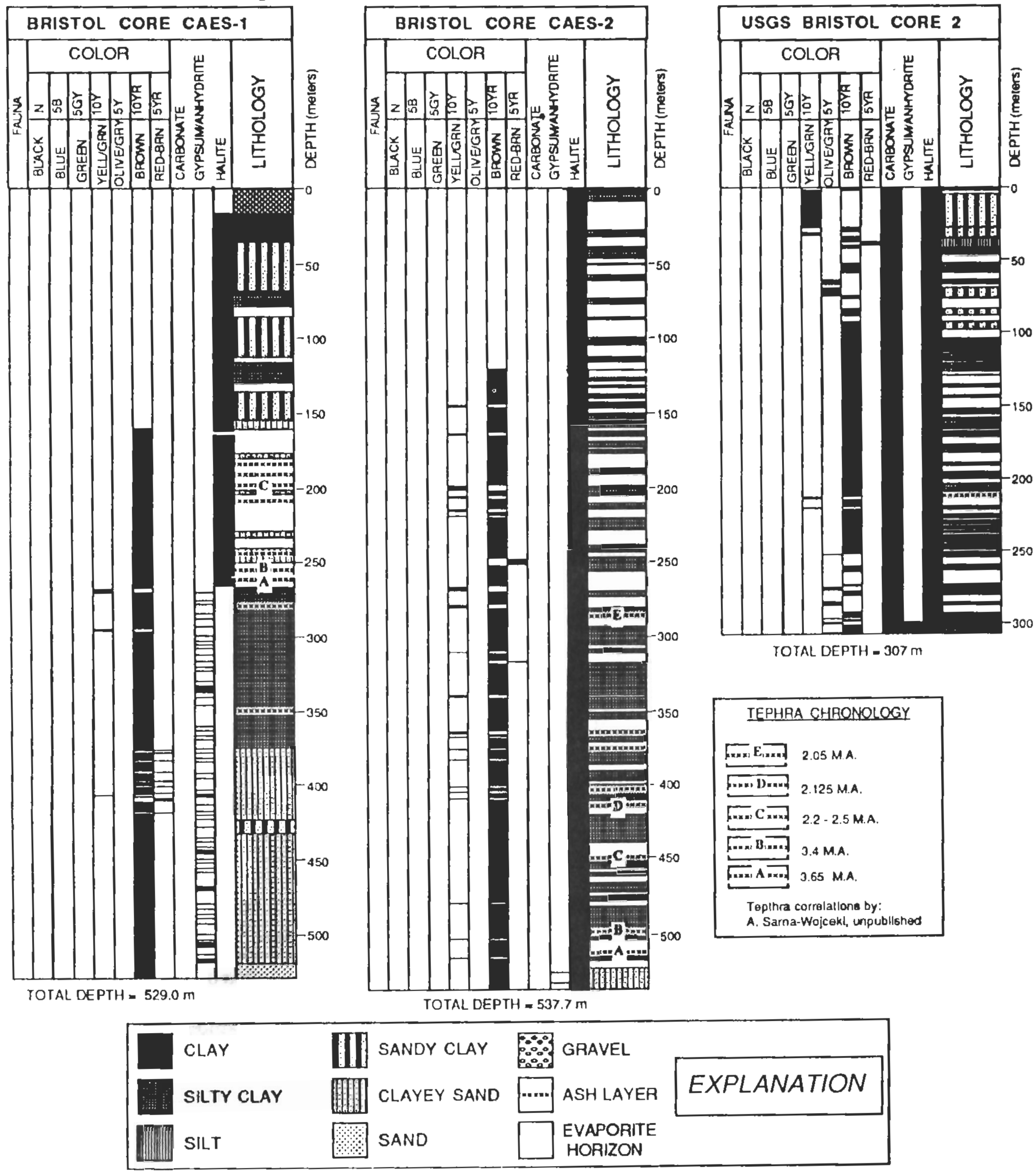

Figure 3. Simplified core logs, Bristol Dry Lake, California. Sediments must contain >50& evaporite minerals before they are considered an "evaporite horizon."

FACIES DISTRIBUTION AND STRATIGRAPHIC FRAMEWORK

Facies Distribution

The distribution of evaporite minerals into concentric rings around the basin (Fig. 2) implies a simple evaporation path of a relatively homogeneous groundwater as it moves towards the basin center. The saline mud flat, between the gypsum-celestite zone and the brine pan, is dominated by siliciclastic muds with only ephemeral displacive halite cubes and hopper crystals precipitating in the facies. The separation of the gypsum zone from the subaqueously deposited halite is a lateral characteristic of Bristol Dry Lake. The cause of this separation is most likely due to the limited availability of sulfate and the fact that the groundwater salinity must increase approximately three-fold before it reaches halite saturation. By the time this has occurred, the water has been ponded at the basin center. Only small amounts of sulfate minerals (gypsum or anhydrite) are found in the saline mud flat and basin center, and much of this is wind transported from the deflating playa margin sediments.

In core, this spatial separation of the sulfate and halite zones is also evident in vertical succession. In the cores taken in the basin center (Fig. 3), brine pan and displacive halite alternates with muds from the saline mud flat deposits for over 500 m and there is no appreciable accumulation of sulfate minerals. Therefore, it appears that in any given vertical sequence it is unlikely that gypsum will be overlain by halite as one might expect in a normal prograding type of marine sequence. Figure 4 shows an idealized cross-section through the middle of the playa to illustrate the facies and mineral distributions. Although Handford's (1982a) subenvironment terminology is retained, the vertical and lateral facies relationships shown here differ significantly from his reconstruction in that the bulk of the gypsum is precipitated in the playa margin sediments and not in the saline mud flat.

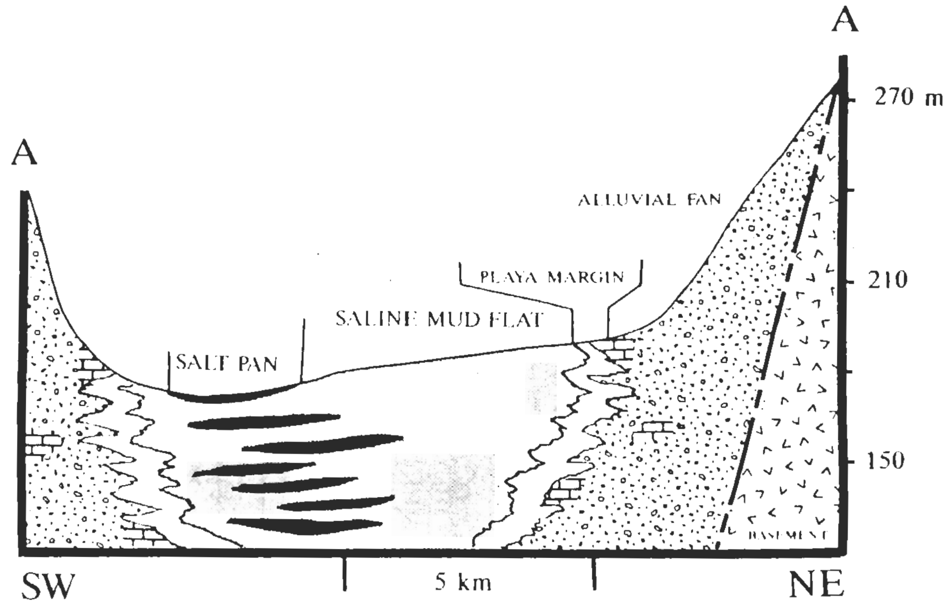

Figure 4. Idealized cross-section, Bristol Lake Playa, emphasizing vertical and horizontal facies distribution. Location of cross-section A—A' shown on Fig. 2.

AGE OF THE BASIN

The correlation of the tephra layers in Bristol Dry Lake with the tephra layers of known ages from other basins provides the basis for the chronology of the basin. In all, nine tephra layers from two cores have been correlated. Three of the tephra layers can be correlated between the two cores (Fig. 3).

There are no previous estimates or measurements of the age of the Bristol Dry Lake basin. Although data from ^{36}Cl measurements in halite indicate that by 150 m depth, the chloride is approximately 2 million years old (Jannik, pers. comm., 1988), this dating technique is still experimental and may have large errors associated with it (Phillips and others, 1983). Nevertheless, there is an indication that the basin is relatively old. Correlations of the deepest tephra layers indicate that the basin is at least 3.7 ± 0.2 million years old at those levels in the cores. In CAES #2 the lowest tephra is almost at the base of the core (500 m), but in CAES #1, there is still almost 270 m of sediment below the deepest ash. Using the lowest and highest calculated sedimentation rates for the core (see Rosen, 1989), the bottom of CAES #1 would be somewhere between 6 and 10 million years old. Although no structural complications are apparent in CAES #1, faulting, such as seen in CAES #2, could greatly reduce the estimates.

Given that the bottom of CAES #1 is between 6 to 10 million years old, this is still not the maximum age limit for the basin. No cores taken in the center of Bristol Dry Lake have reached basement rocks. Therefore, there is still an unknown depth of sediment below the cored interval that is older than the estimate. A crude estimate of the depth of the basin can be attempted by interpreting the Bouguer gravity map for the Needles quadrangle. Using an equation from Telford and others (1976), the total thickness of the basin center is estimated to be approximately 680 meters. If this calculation is correct, bedrock is approximately 100 m below the bottom of CAES #2.

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Saline Minerals Extraction from Southern Mojave Desert Playas of California

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INTRODUCTION

Past and present commercial sources of several mineral salts from four major playas of the southern Mojave Desert reside in southeastern San Bernardino County, California. The playas are Dale Lake, Bristol Lake, Cadiz Lake and Danby Lake. This paper presents an overview of the playa types in relation to the saline minerals produced, reviews some mineral production history, and provides selected references about these saline playas, mineral resources, mineral potential and commodity reports.

Previous Work

Previous work on southern Mojave Desert playas is abundant and many of the published references are listed here. "The Saline Deposits of California" (Bailey, 1902) may have been the first report with a map (back cover, this volume) portraying "Burts Lake" (now Dale Lake) and Bristol, Cadiz and Danby Lakes. The reader is referred to one review of the original literature by Papke (1976) for greater elaboration of general playa characteristics, classifications, or definitions. Droste (1961) describes the variety of playa surfaces of the Mojave Desert, and reports on several groups of clay minerals (traceable to source areas) present for Dale, Bristol, Cadiz, Danby Lake, and other playas.

Saline minerals and mineral resources are described in many sources of information. The Bibliography includes mineral commodity reports compiled over the years on salines in general and on specific mineral resources pertaining to southern Mojave Desert playas. By 1992, there had been no comprehensive coverage since the 25 year period 1942 through 1967, such as the works by Tucker and Sampson (1943), Wright and others (1953), Mumford (1954), Smith (1966a-c), and Moyle (1961 and 1967). Substantial work to estimate mineral resources and reserves by the U. S. Geological Survey and U. S. Bureau of Mines has been performed to assess the availability of specific mineral commodities (Brobst and Pratt, 1973; Knoerr, 1980; and U. S. Bureau of Mines and U. S. Geological Survey, 1981). However, exact mineral reserves information is not known for minerals and brines of Dale, Bristol, Cadiz and Danby Lakes.

Work to delineate mineral lands for special consideration among potential users and land managers has been undertaken by several agencies. The U.S. Geological Survey, and later the Bureau of Land Management (1985), produced maps identifying lands valuable prospectively and defining lands known to be valuable for certain minerals (Known Leasing Areas). For identification, and for advice to land planners and managers, mineral resource occurence or development potential maps of overall areas have been and are currently being produced with coverage in the Dale, Bristol, Cadiz and Danby areas (Calzia and others, 1978, 1979; U.S. Bureau of Land Management, 1980; California Division of Mines and Geology, work in progress). Systematic classifications exist for mineral deposit types in certain tectonostratigraphic terranes (Albers, 1982; Albers and Fraticelli, 1984) and for mineral deposit models (Cox and Singer, 1986). For an in-depth overview from exploration, mining and mineral classification, to unfavorable circumstances regarding minerals usage and availability, see Cameron (1986).

MINERAL DEPOSIT CHARACTERISTICS

Dry Lake Classification

Bailey (1902) referred to four ancient lake systems, two of which had previously taken on the U.S. Geological Survey names Lake Bonneville and Lake Lahontan for the respective northeast and northwest portions of the Great Basin. Bailey proposed names for the two ancient inland lake systems to the south. In Bailey's accompanying "Maps of Lake Le Conte and Aubury" (Fig. 1), Lake Le Conte was used for the area now superposed by the term Lake Cahuilla in the Salton Trough of California, and Lake Aubrey for the areas presently considered the southwestern Basin and Range and the Mojave Desert portions of southeast California. (Bailey's two fold-out map plates depicting dry lakes omits Bristol Lake, and a third map plate of the entire state omits Bristol and Cadiz Lakes.)

In subsequent literature, many ideas have been proposed explaining how various drainage basins in the southwest Basin and Range and the Mojave Desert have been interconnected in past times (eg., Blanc and Cleveland, 1961a, 1961b; Rosen, 1991; Brown and Rosen, this volume). These include paleo-surface flow-paths from the central Mojave Desert through the southern Mojave Desert "Bristol-Cadiz-Danby Lake chain" to the Colorado River. Models to assist in characterizing the change in water chemistry from one basin to another, and the resultant evaporite types deposited in each basin over time in a chain of interconnected basins, are based on study of the paleo-Owens River drainage (Hardie and Eugster, 1970; Drever, 1988). Based on the Hardie-Eugster model of chemical divides, without elaborating in significant detail here, playas such as Bristol, Cadiz, and Danby (and respective brines) can be categorized chemically (see below).

A number of terms and categories have been used to describe or characterize playas. Simple names for them are numerous: dry lake, dried lake, dry lagoon, desiccated lake-bed, mud flat, clay pan, salt pan, alkali flat, soda lake, salt flat, river sink, sink, salina, saline placer, saline lake, alkali marsh, salt marsh, or simply, marsh. Reporting on the investigations for locating sources of potash and other salts, Young (1915) differentiates between mud playas (dry and likely not suitable for finding economic saline minerals) and marshes. Young (p. 61) depicts Bristol and Danby Lakes being doubtful concerning existence of saline beds and value. Continuing the search for potash and other salts, Foshag (1926) presents considerable descriptive details on saline lakes of the Mojave Desert and presents two classes of playas, wet and dry, based on an earlier 1920 suggestion from David Thompson (Foshag, p. 57). The two-fold classification was further developed in a comprehensive U. S. Geological Survey report on Mojave Desert water resources (Thompson, 1929). The breakdown was then extended by Stone (1952) based on thesis work, and described in a detailed report on California salt by W. E. Ver Planck (1958). This classification consists of five types: dry type, clay-encrusted/salt-encrusted wet type, crystal body type, compound type, and artificial. According to Ver Plank, Dale, Bristol, Cadiz, and Danby Lakes are 4 of the only 5 crystal body type playas in California (Searles Lake, San Bernardino County being the fifth).

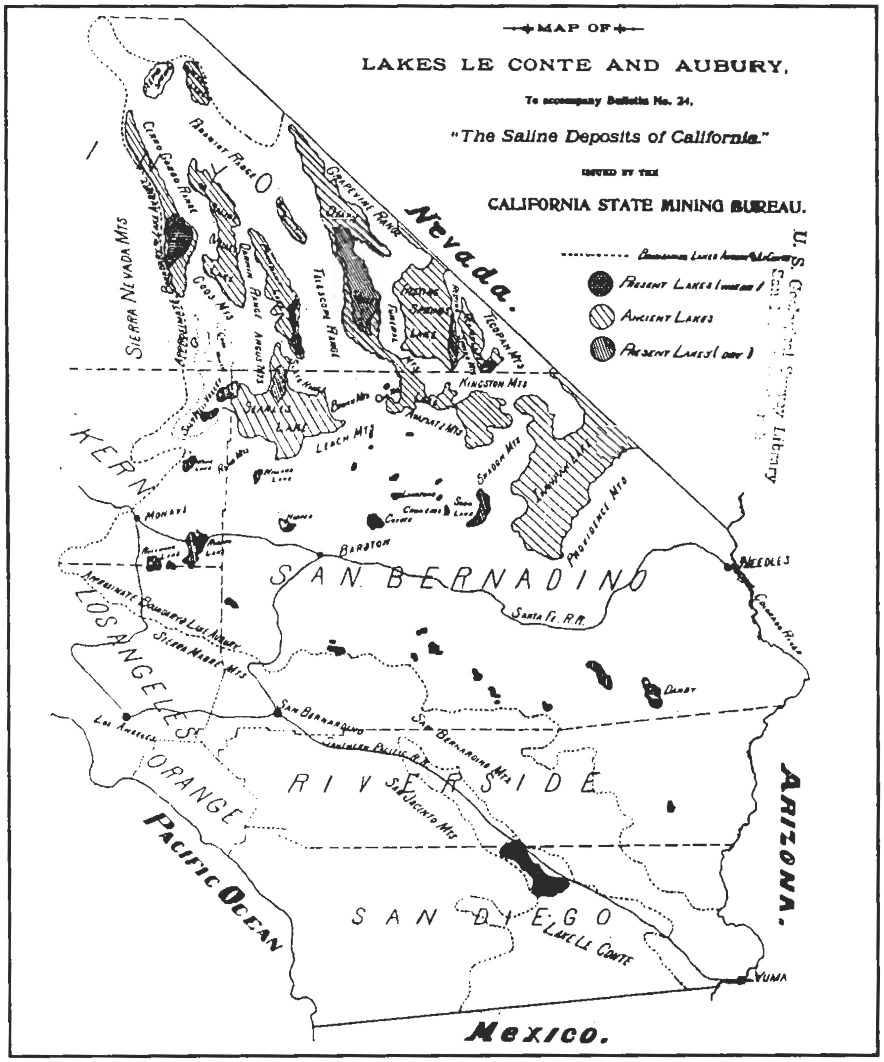

Figure 1. Map of Lake Le Conte and Aubury, showing playas along Bristol-Danby trough. From Bailey (1902).

Mineral Classification

The southern Mojave Desert lies within platform facies cratonal terrane delineated by Albers (1981). As indicated on a mineral resources assesment map (Albers and Fraticelli, 1984), Dale, Bristol, Cadiz, and Danby Lakes all are classified within geologically favorable terrane that contain known saline deposits of potential commercial value as indicated by drill-hole information. There has been other classification of non-marine evaporite deposits and there is a need to perform modeling studies (Sheldon and Raup, 1981). Mineral deposit models for the salines are not recognized in many works (Ridge, 1968; Skinner, 1981) including a more recent, major work and follow-up Open-File Reports on mineral deposits (Cox and Singer, 1986). Although short of a mineral deposit definition or standard, the Dale, Bristol, Cadiz, and Danby Lake mineral deposits can be characterized by their contained minerals and fluid.

The fluids in playas can be described in increasing saline concentration as fresh, brackish, hypersaline, or evaporitic, using a convention based on the chemical activity of water in contrast to indicated salinity (Drever, 1988). Due to a deficit of carbonates (which indicate fresh, brackish or hypersaline conditions), the subject playa brines are evaporitic. In another convention (Lloyd and Heathcote, 1985) the terms in increasing salinity are fresh, brackish, saline, and brine, brine having at least 100,000 total dissolved solids (TDS) or miligrams per liter (mg/l). Due to significant mineral saturation, all subject playa fluids are classed as brine using the latter scheme. Respective concentrations of total dissolved solids expressed as parts per million (ppm) for Dale, Bristol, Cadiz, and Danby reported by Ver Planck (p. 123, 1958) are 298,000, 279,149, 73,600 and 271,200 ppm. Respective specific gravity measurements reported by Calzia (1979a-d) for Dale, Bristol, Cadiz, and Danby brines are about 1.025, 1.2, 1.1, and 1.14 grams/cm^3. According to Ver Plank (1958), Dale brines were reported with a specific gravity of 1.21 to 1.25 gm/cc. Using a brine classification system by Drever (1988), calcium build-up in the brine remains while alkaline carbonate constituents are absent. Under Drever's scheme Bristol and Cadiz Lake brines are chloride brines, and Dale and Danby Lake brines are carbonate-free chloride-sulphate brines.

Dale, Bristol, Cadiz, and Danby Lakes are all classified valuable prospectively for sodium leasable minerals (U. S. Bureau of Land Management, 1985). Dale Lake contains a further classification as Known Leasing Area for sodium (a portion near producible wells). Bristol and Danby Lakes are both classified as Known Leasing Areas for sodium. Based on the California Desert Plan Geology-Energy-Mineral Resource Area Assessment, all four lakes are classed as having high potential for sodium mineral resources (Bureau of Land Management, 1980). For other minerals not leasable, a mineral lands classification map depicts the producing deposit areas of Bristol and Danby Lakes, known or possible deposit limits of Cadiz Lake, and covered (unknown) identity of Dale Lake (Calzia and Smith, 1978). Mineral resource estimates of crystalline sodium compounds for the four playa deposits are reported in Calzia and others (1979).

SALINE MINERAL PRODUCTION HISTORY

Playa evaporite sediments have been worked for sodium chloride (rock salt, NaCl), sodium sulphate (salt cake, Na_2SO_4), and calcium sulphate (selenite-gypsum and gypsite, $CaSO_4.2H_2O$). Associated brines have been the source of common salt (NaCl); salt cake ($NaSO_4$); and liquid and anhydrous-flake calcium-chloride ($CaCl_2$) with potassium chloride (KCl).

There is past production of sodium sulphate from Dale and Danby Lakes, and calcium sulphate from Bristol and Danby Lakes. There is past and continued production of calcium chloride from Bristol and Cadiz Lakes, and sodium chloride from Dale, Bristol, Cadiz, and Danby Lakes. During settlement and continual developmental expansion of the southern California region, the exploration, study and production of these economic saline minerals accompanied the establishment and growth of transportation routes and other mineral industries in the southern Mojave Desert.

Two drill holes to depths of 504 feet and 815 feet bottomed in salt in Bristol Lake (Bassett and others, 1959). Two additional holes, 1006 and 1007 feet deep, bottomed in salt at Bristol Lake (Calzia, 1979a). A 500-foot drill hole revealed salt only to a depth of 9 feet in Cadiz Lake (Bassett and others, 1959). One additional hole to 415 feet in depth encountered salt only near the surface (Calzia, 1979b). Two drill holes to depths of 880 feet and 460 feet did not encounter salt in Danby Lake (Bassett and others, 1959). In two additional holes to 504 and 503 feet in depth at Danby Lake, salt was found at a depth of 440 feet in one and no salt was encountered in the other (Calzia 1979c). A 440-foot deep drill hole revealed salt only to a depth of 2.5 feet in Dale Lake (Calzia, 1979d).

Dale Lake

Initial exploration drilling took place between 1920 and 1924 by Irving Bush (Wright and others, 1953). Dale Chemical Company produced salt and sodium sulphate from well brine from 1939 to 1946, according to Ver Planck (1958). Dale Chemical Industries Incorporated continued production in 1947 and 1948. Several brine production wells, over a dozen solar evaporation ponds, and a processing plant were established. A unique spray method produced brines using gravity separation to drop out sodium sulphate (Glauber's salt) from the remaining salt brine (Moyle, 1961). The crude sodium sulphate was made into salt cake in the plant (Wright and others, 1953). The property was leased by Don's Salt Service to glean unharvested salt remaining from Dale Chemical Industries (Ver Plank, 1958). Between 1949 and 1986, Southwest Salt Company acquired ownership of the operation and mineral rights. Southwest Salt turned over the property to Gerry Grott and family in the mid-1980s (as a result of Southwest Salt property acquisition by Morton-Thiokol, ie., Morton Salt, in several Mojave Desert dry lakes). An application for a sodium prospecting permit was made by Superior Salt Company (relation to several Grotts) and was granted in 1989 for leasing peripheral federal lands adjacent to the private holdings dominating the deposit. No exploration occurred by the November 31, 1991, permit expiration.

Bristol Lake

In 1908, Crystal Salt Company made placer claim locations for calcium chloride (Gale, 1951). Crystal Salt Company produced salt by stripping and blasting from 1909 to 1913, as Consumers Salt Company in 1916 and 1917, as Pacific Rock Salt Company in 1918 to 1920, and as California Rock Salt Company from 1921 to 1950. In *U.S. vs. California Rock Salt Company* (1950, 60 L.D. 420, 422), all but two of 38 contiguous placer claims of California Rock Salt Company were declared invalid based on the one claim per locator provision of the Act of January 31, 1901 (Saline Placer Act, 31 Stat. 745). California Salt Company operated here from 1942-1951 (Gale, 1951), and became California Salt Company in 1950 (Ver Plank (1958). Leslie-California Salt Company organized in 1924 and became Leslie Salt Company in 1936, currently owned by Cargill Corp. Original placer mining claims were located in 1908, with unreported production of calcium chloride until 1921. Operations by Leslie Salt Company continue to present with production of sodium chloride and calcium chloride. Brines are produced through a combination of brine seepage ditches, pits, and wells, with solar evaporation taking place along approximately 27-miles of canal-ways and solar evaporation ponds.

Aal Salt and Chemical Company produced salt in 1921, followed by Saline Products Company from 1924 to 1936. Ownership or operator names changed to Hollar Chemical, and then to Desert Properties Company in 1939, to eventual ownership by National Chloride Company in 1950. Up to the present time, all efforts were principally making salt as a by-product of calcium chloride production.

Near the southwest margin of the lake, and on U.S. Gypsum holdings northwest of Leslie Salt claims, gypsum was mined by Consolidated Pacific Cement Plaster Company from 1907 until idled in 1924 (Wright and others, 1953). Since 1940, the Hill Brothers Chemical Company has produced flake calcium chloride in dehydrating plants from brines. Two cars of celestite nodules produced by U. S. Geological Survey workers were recovered from near-surface accumulations in 1942 (Durrell, 1953).

Cadiz Lake

Brines were being tested for possible recovery in the 1950s (Wright and others, 1953). The Hills Brothers Chemical Company constructed drill holes up to 30 feet deep (leased from Lee Bardsley of Amboy). There was no production until after the 1960s. There are two main operating areas on the Lake, Delta Chemical and Lee Chemicals. Operations have been developed on placer mining claims (several were patented in two episodes, during the 1970s and 1980s). Road and solar evaporation ponds, berms, and dikes were constructed, and many brine wells were drilled and constructed to pump lake brine for calcium chloride. Salt is produced and used as construction material or stockpiled. There are no sodium leases issued from the U. S. Government.

Danby Lake

The Crystal Salt Company was working the Surprise salt mine at the turn of the century. A 40 acre area was developed into bedded rock salt, according to Bailey (1902). A shaft 35 feet into rock salt existed as early as 1882; later, a second to a depth of 65 feet encountered 22 feet of rock salt. The enterprise comprised 800 acres of claims by the locators. Bailey's account includes a photograph of desert wagons drawn by "locomobile trains" (boiler-powered steam tractor). Salt product was hauled in traction wagons to the Santa Fe Railroad Danby station (now A.T. & S.F.). The A.T. & S.F. rail line route from Cadiz station to Parker and Phoenix alongside Danby Lake (now leased by the Arizona and California Railway Company) was not yet in existence. Interestingly, Bailey's report mentions the unconfirmed existence of niter, or "cubic" or "chili saltpeter" discovered and located in 1901 in the playa deposits near the salt beds at Danby Lake (this "nitrate of soda" is otherwise known as soda niter, sodium nitrate, or $NaNO_3$). Both deposits are shown at the north end of Danby Lake. Much of the produced salt was sold to silver mills for use in chloridizing.

According to Ver Planck (1958), the name Crystal Salt Company changed to Crystal Rock Salt Mining Company and operated by stripping and blasting rock salt between 1890 and 1895. Milligan Salt Company operated solar evaporation ponds with a plant near the Milligan siding. Towards the south end of the playa near Ward Siding (now Saltmarsh), production by blasting and stripping occurred from 1914 to 1916, and in 1920 by R. B. Evans on Avery-Evans holdings. Rock Salt Products Company (J. W. Reeder) produced salt by stripping and blasting at the southern part of the playa near Saltmarsh from 1934 to 1942. Arthur Doran and Tom Schofield produced surface mound deposits of mirabilite (sodium sulphate) west of Saltmarsh, shipping 90 to 100 car loads in the 1920s (Wright, and others, 1953).

Metropolitan Water District of Southern California (MWD) acquired sodium preference right leases from prospecting permit discoveries and developed the leases from 1940 to 1950. Salt was to be used to treat water at the La Verne Plant. Danby Salt Corp. and MWD continued operations into the 1960s (Moyle, 1967). Operations by MWD ceased in the 1980s. National Chloride Company of America obtained sodium leases from prospecting permit discovery. Brines were produced from well-pumping and from a brine seepage ditch nearly 3000 feet long. In 1987, National Chloride assigned its federal lease and use permit to Salt Products Company, Inc., currently producing specialty salts. Three current, inactive sodium leases are held by W. C. Reeder, F. Rielly and Morton-Thiokol (Morton Salt Company).

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Some Thoughts on the Development of Amboy Crater

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LOCATION

Participants in the 1992 Mojave Desert Quaternary Research Symposium field trip will pass the eastern margin of the Amboy lava field along the edge of Bristol dry lake. This field of young flows sprawls over 62 square kilometers of nearly flat, alluviated desert floor. The field is roughly circular in shape, at its widest stretching some 12 kilometers across. Near its northeastern corner, easily visible from the highway, is the primary source vent—Amboy Crater—a cinder cone 80 meters high and nearly 500 meters in diameter.

Amboy Crater is associated with a group of about a dozen youthful cinder cones scattered across an area stretching approximately 80 kilometers WNW-ESE. Among the best known vents are Dish Hill, Siberia Crater, Sunshine Cone, and Mt. Pisgah. Amboy is the easternmost of the cones, and one of the youngest.

These small volcanoes are roughly aligned along the northwestward continuation of the Bristol-Danby trough, a string of basins containing the saline playas of the same names. The basins may have resulted from transtensional tectonism between northwest-southeast striking strike-slip faults, or perhaps from detachment faulting, or a combination of both processes.

LITHOLOGY

The volcanic rocks of Amboy Crater and related vents are alkali basalts. Relative to other basalts, alkali basalts are distinguished by high alkali and low silica content. The principal alkali ion in continental alkali basalts is potassium. Combined with low silica, alkali undersaturation may occur, leading to the crystallization of leucite in place of potassium feldspar. Given the typically high eruption temperature of alkali basalt lavas, and relatively low crystallization temperature of leucite, this phenocryst is not often found. Some alkali basalt lavas, however, contain abundant leucite phenocrysts. The flows at Amboy contains sparse, small forsteritic olivine and labradoritic plagioclase phenocrysts. The silica content of Amboy lava is about 47 weight percent (Parker, 1979, 1963; Janet Hirsch, written communication, 1985). This compares with silica values of 50-52 weight percent for other types of basalt.

Laboratory study of phase equilibria in the basalt system indicate that alkali basalts must form from partial melting at greater depths than other basalts. Because of the great thickness of ordinary continental crust, basalts erupted in interior continental settings tend to be alkalic in composition. Many alkalic basalt flows include mantle-derived peridotite xenoliths. No such xenoliths have been found in the Amboy lava; however, they are abundant at nearby Dish Hill (Wilshire and others, 1985; Wilshire and Nielson-Pike, 1986).

REGIONAL VOLCANISM

Alkalic eruptions mark the earliest stage of continental rifting, accompanying thermal arching, and the incipient foundering of fault blocks along the axis of the arch. This foundering marks the site of a future rift valley. In the immature southern East African Rift system, alkalic volcanism is common. With progressive extension of the crust and foundering of rift valley floors, the locus of partial melting rises to shallower depths, and ultimately tholeitic basalts characteristic of typical seafloor are erupted.

Apart from geographical coincidence, it is not clear how the volcanism is related to the Bristol-Danby trough. One might suppose that the trough and its extension toward Barstow act as an intracontinental rift system. But vents locally line up in north-south, or NNE-SSW-trending groups, cutting at high angles across trough-bounding faults and other regional tectonic fabric. Possibly this reflects the effect of some deep-crustal structure on the ascent of the magmas. Extensional alignments and dikes with this orientation are present in basement crust of mid-Proterozoic and greater age in this region. Also, some vents lie atop thick mountain blocks rather than in basinal lowlands. In short, the magmatotectonic significance of recent volcanism in the central Mojave remains enigmatic.

A Quaternary alkalic volcanic province very similar to the Amboy-Dish Hill-Pisgah group exists in China, where extensive strike slip faulting has also contributed to the development of pull-apart basins (Liu Jiaqui, written communication, 1989). Both in central east Asia and in the Mojave, rifting may be regarded as weak and widely dispersed, rather than concentrated, as in the African or Rio Grande rift valleys.

AGE

Most young-looking cinder cones in the central Mojave are Pleistocene in age, with fission-track and K-Ar ages falling around two million years (Wilshire and others, 1985). Basalt flows interbedded with lacustrine deposits deep within the Bristol lakebed are probably associated with even older volcanic events in the Bristol-Danby trend. Amboy is exceptionally young. While some fluviolacustrine sediment associated with Bristol Lake laps up against the lava field, the lava for the most part overlies the youngest sedimentary layers. If the last significant deposition of sediment occurred in Bristol Lake shortly after the end of the last Ice Age, Amboy Crater would be less than 10,000 years old. If the lake level was high during Tioga glaciation, the volcanic field could be even younger—less than 6,000 years (Parker, 1963). Secular paleomagnetic field study has not yet provided a more definite age (Ray Wells, 1991, personal communication). In any event, the excellent state of preservation of these volcanics certifies they are quite young. That future eruptions may take place in the region is probably, though Amboy itself is very likely extinct.

AMBOY FLOW FIELD

Eruptive History

The fact that the Amboy flow field is in few places thicker than a few meters, and that the flow surface is pahoehoe throughout, suggest that the erupted lava was very fluid, and spread quickly—much like a small scale outpouring of flood basalt. High temperature, high gas content, and low silica abundance could account for this. The existence of original high temperatures is supported by the sparsity and smallness of phenocrysts found in hand samples. A high gas content is suggested both by the high vesicularity of some samples, and the abundance of tumuli (mound-like "gas blisters"). The very low silica value clearly favors low viscosity effusion, if combined with the other two factors above.

Several other factors probably also played a role in developing the wide lava field at Amboy: discharge rate must have been very rapid, and the early phases of the eruption may have occurred along a fissure, or from several aligned vents.

High discharge is suggested by the fact that the flow surface is generally level throughout the lava field. If discharge had been slow, cooling of small flow lobes would have caused a build-up of lava around the main vent, creating a broad lava shield or lava cone. It is possible that the entire flow field could have extruded within a period of several weeks, given historical rates of eruptions observed at high-discharge basaltic volcanoes elsewhere.

In addition to Amboy Crater itself, lava may have extruded from another vent simultaneously, about three kilometers to the southwest. This area, termed The Plateau by Parker (1963), is an elevated flatland broken by piles of blocky fragments, and by shallow pits enclosed in low rims of ejecta. These features could mark the sites of eruptive blasts, though Parker (1963) suggests these features could also have formed from weak phreatomagmatic explosions where lava overrode wet sediment. The tableland itself may have grown as a perched lava pond, with or without the presence of an underlying vent.

If indeed The Plateau represents a second vent structure, then it seems probable the Amboy eruption began with opening of a fissure along which the outpouring of lava became concentrated at only a few localities. This eruptive style is typical of basaltic volcanoes. In the final phase of activity, eruption is constrained to a single vent along the fissure, around which a cone is built. Amboy Crater could represent such a late stage cone.

However, eruptive fissures are generally preserved as open oxidized vents or spatter ramparts. These features are not present in the Amboy lava field. If an eruptive fissure initiated the Amboy eruption, it cannot have remained active for long, and was subsequently buried by lava flowing from the site of the cone.

Many basaltic volcanic fields contain lava channel and lava tube systems. These are conspicuously lacking in the Amboy lava field. In part, this may be due to a short duration eruption; well-established channel-tube networks require many days of time and steady-state conditions to form. More importantly, perhaps, the lava field spread out over essentially flat terrain. Lacking a slope to guide flow movement, well-directed channel-tube forming currents could not develop. Parker's (1963) mapping shows a very turbulent pattern of flow in the Amboy lava field.

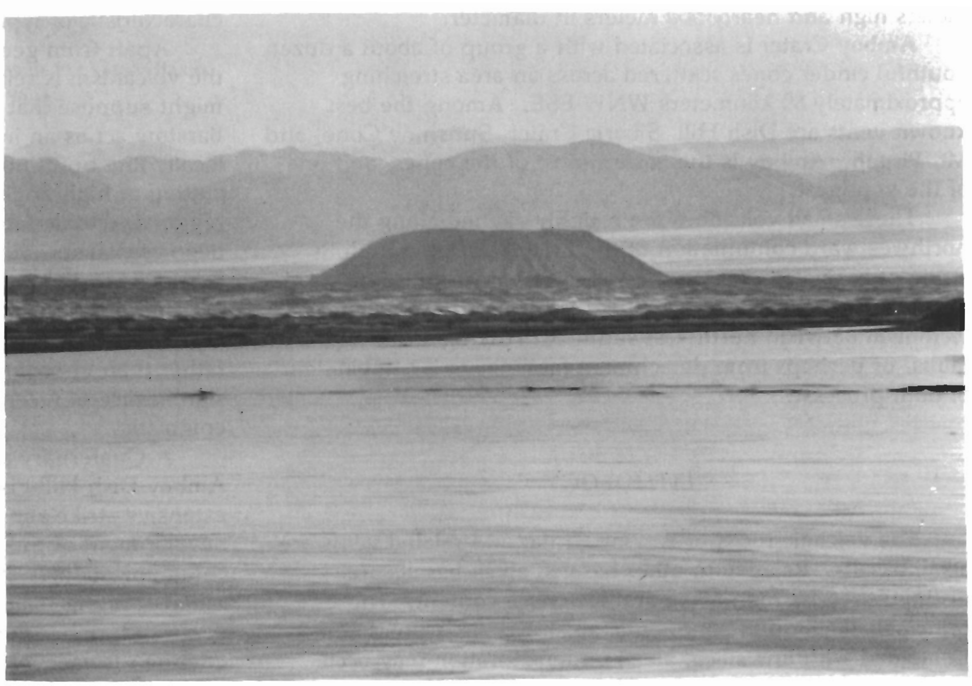

Figure 1. The low, broad cinder cone and lava field at Amboy, viewed across the Bristol Lake playa. ***R.E. Reynolds photo.***

Amboy Cinder Cone

As in historically observed eruptions which have formed cinder cones, cone growth at Amboy probably occurred throughout the period of lava extrusions. But in several respects, the Amboy cone is unusual: given the size of the lava field, Amboy cone is small. It is also unusually low and broad (Fig. 1); most cinder cones have smaller craters in proportion to their heights.

The unusual size and shape of the Amboy cone can be attributed to several factors. Perhaps explosive discharge of gas was less vigorous in this eruption than that of many cinder cones. This would create less ejecta to build up a cone. Also, the focal point of explosions shifted during the eruption, so that a broad crater containing at least three nested conelets resulted. The rims of these intra-crater cones may still be seen inside Amboy Crater. They formed during the waning stages of activity. Wind blown sand and clays trapped by the rims of the nested cones have formed a hardpan soil on the floor of the crater in several places, allowing for development of miniature playas.

A wide breach in the southwest wall of the main crater was probably opened as lava extruding from the flank of the main cone undermined the crater wall. To most easily enter the crater, one ascends this breaching flow. Climbing to the crater rim, one may find a profusion of lava bombs—most spindle, ribbon, and fusiform in shape. These bombs are also scattered across the flow surface to the northeast of the cone as far as a kilometer away (Parker, 1963). One may ponder this in considering the power of the volcanic blasts which once occurred here.

A short walk around the rim will reveal that the eastern flank of the cone is dissected by an intricate gully system. This stands in marked contrast to the little-eroded western flank. The origin of this gully system poses an interesting geological question. Almost all cinder cones are the products of single eruptions, though two cones may grow at different times in close proximity to one another. This is because the magma feeding cinder cones is produced in small, discrete batches. If indeed Amboy cone grew during a single brief eruption, why do its flanks show such striking erosional contrast? Parker (1963) suggested a prolonged erosional interval may have occurred between eruptive phases, with the western portion of the cone forming at a later time. Alternatively, he proposed the channels resulted from rill-like collapses in the eastern flank which have subsequently been modified by erosion.

If a prolonged erosional interval occurred, there is certainly no trace of it in the appearance of the surrounding flows. The lavas all appear to be of the same general age. The possibility of older flows having been completely buried by younger ones seems unlikely. The older lavas would more probably stand out in kipuka, or as raised areas which guided the flow of younger lava.

Though mass-wasting is an important process on volcanoes, even during periods of eruption, no known process produces a pattern of rilled channels in active cinder cones. Given this, and the above reservation about an erosional interval, a more likely explanation for the contrast seems to be differential erosion of the cone surface due to differences in material composition. The western flank of the cone is reinforced by spatter agglutinate deposited during an early stage of the eruption. This material is relatively resistant to erosion, and does not dislodge easily during periods of high runoff. The eastern flank is much less consolidated, however. It is also higher, favoring a greater supply of runoff. Channel formation proceeded in a headward manner—which is typical of stream erosion on volcanoes—with channel heads now lying only a short distance beneath the crater rim. Slope wash deposits of loose cinder and bomb fragments have entered most channel heads.

Amboy cinder cone has acted as a giant barrier for the prevailing movement of windblown sand across the lava field. The light-colored, sandier flow surfaces stand out relative to the darker, sand-starved areas. Using this albedo difference, a wind-shadow zone may be discerned extending from the crater rim eastward toward Bristol dry lake.

The Amboy volcanic field is an excellent area for studying volcanology. It is the site of one of the youngest eruptions in southern California. Basic questions concerning the age, duration, and pattern of eruption remain. But important clues to answer these questions may be seen in the present landscape.

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Bolo Hill Rendezvous: History, Ethnography, and Prehistory of the Bristol Lake Area

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INTRODUCTION

Studies conducted over the past two years in connection with the environmental review of a proposed landfill project known as the RAIL•CYCLE Bolo Station Facility have resulted in the documentation of historic and prehistoric cultural resources in the project area (Lerch 1992). That project site stretches from near Bolo Hill for a distance of four miles to the shoreline of Bristol Lake, with a width that varies from one to three miles. The proposed landfill will be located on the southern half of the project site, on the opposite side of the railroad. The area between Bolo Hill and the railroad is to be maintained as a desert preserve. Based on the cultural resources studies conducted for the RAIL•CYCLE project, this paper presents a brief review of the historic, ethnographic, and prehistoric background of the area.

HISTORY

Historic use of this area centered around transportation, both by rail and by road. To the southwest, mining for salt and gypsum in Bristol Lake also has been an important activity since the turn-of-the-century.

The earliest known historic activity in the project region occurred in January and February of 1868, when General William J. Palmer and his party conducted a survey for the railroad route (Palmer 1869). The railroad ultimately was built by the Southern Pacific Railroad and service to Amboy began on March 12, 1883. The section from Amboy to Needles was completed on April 19, 1883. In October, 1884, the line was purchased by the Atlantic and Pacific Railroad, and subsequently acquired by the Atchison, Topeka, and Santa Fe Railroad in 1890. The Santa Fe Railroad continues to operate the line up to the present (Myrick 1963:766,788).

Along this stretch of the railroad, section camps were named in alphabetical order, starting with Amboy and continuing eastward to Bristol, Cadiz, Danby, Essex, Fenner, Goffs, Homer, Ibis, and Java. The section camp and siding in this area, located two miles south of here along the railroad, was known successively as Bristol, Bombay, Bengal, and finally, Bolo. The original name of Bristol is shown on a train schedule for Dec. 1, 1883 (Myrick 1963:770). The name was changed by Santa Fe in 1898 to Bombay and then to Bengal, and it later was changed again in 1915 to Bolo (Gudde 1960:33).

The section camp is depicted on a number of historic maps as both Bengal and as Bolo. The site is shown as Bengal on Rueger's 1903 *Automobile and Miners' Road Map of Southern California*, on Mendenhall's 1909 map of *Desert Watering Places in Southeastern California and Southwestern Nevada*, and on the Automobile Club of Southern California's 1914 *Map of a Portion of Southern California and Southwestern Nevada Embracing the Arid Region of Mohave Desert, Colorado Basin, and Death Valley.* It appears as Bolo on Thompson's 1921 *Relief Map of Part of Mohave Desert Region, California, Showing Desert Watering Places*, and on the State Railroad Commission's 1926 *Official Railroad Map of California*.

Section camps housed maintenance crews and served as water stops for the steam locomotives. Water was hauled to the Bolo section camp in tank cars from Newberry Springs. Beginning in 1941, the railroad introduced diesel-powered locomotives which did not require frequent water stops as did the steam locomotives. The conversion to diesel power was completed by 1952, after which time there was no further need for section camps such as Bolo (Duke and Kistler 1963:75; Bryant 1974:312-316). The Bolo section camp apparently was abandoned by that time, and useable materials most likely were salvaged for use elsewhere. Nothing is shown at the site on the 1956 USGS 15' Cadiz topographic quadrangle, surveyed in 1954. The placename "Bolo" had by then been shifted to designate the small hill adjacent to this stop.

Another aspect of transportation history which is relevant to this stop is that of automobile travel on the National Old Trails Road. The route is first shown on an Auto Club map of 1914. It was described in 1918 by Thompson (1921:123) as an oiled road which ran approximately one-fourth mile south of the current paved road. Thompson also described a junction with a branch road to Cadiz which was located one-half mile west of Bolo Hill. On his map of the area, the branch road to Cadiz is labeled as the "old main road abandoned" east of Cadiz (Thompson 1921:Plate XII).

In 1926, the oiled road described by Thompson was designated part of U.S. Route 66, which connected Chicago to Los Angeles. It was paved through the California desert by 1934 (Scott and Kelly 1988:31). The current paved road runs approximately one-quarter mile north of the old road through the area. Route 66 was eventually bypassed as a major automobile route when Interstate 40 was completed in 1972 (Scott and Kelly 1988:183).

Thus, in this region there is evidence of the historic change of travel routes through the desert. The earliest route was the railroad itself, beginning in 1883. The first automobile route ran from Amboy to this vicinity, and then ran southeasterly to Cadiz on the railroad and around the south end of the Marble Mountains. That road had been abandoned for travel purposes (although it continued in use until recently as a pole-line access road) by 1914, when it was supplanted by an oiled road originally known as National Old Trails Road and later as U.S. Route 66. The oiled road in turn was abandoned when the route was moved a short distance north when it was paved in 1932. And finally, Route 66 became a "back road" in 1972, when most of the traffic across the desert began traveling on the newly completed Interstate 40, 12 miles to the north.

As technology improved and the historic travel routes through the area changed, the historic settlement pattern also changed. Section camps such as Bristol/Bombay/Bengal/ Bolo no longer were needed as water stops on the railroad and were abandoned. Roads that once were the major routes through the region became little more than dirt tracks across the desert. And once-thriving stops on Route 66 such as Amboy were reduced to vestiges of their former selves when they were bypassed by the freeway. Some, like Bagdad, have disappeared altogether.

ETHNOGRAPHY

Ethnohistoric data suggest that the populations that occupied this area of the desert during much of the Protohistoric period likely were the Desert Mojave, speakers of a Yuman language who were related to the historic Mojave Indians of the Colorado River (Lerch 1990). The Desert Mojave were succeeded late in the Protohistoric period by the ancestors of the ethnographic inhabitants of the region, the Chemehuevi Indians. Other groups that may have utilized the project area were the Serrano and the Mojave Indians.

The project site is shown by Knack (1980:82) as an area of intermittent use bordering the territories of the Chemehuevi to the northeast and the Serrano to the southwest. Both groups were hunters and gatherers whose territory included several different biota, ranging from the margins of ephemeral lakes on the desert floor to higher mountain elevations. Hunting was primarily a male activity, while gathering was largely the province of the women. Animals commonly taken as game included deer, mountain sheep, pronghorn, hares, rabbits, rodents, and birds. Hunters made use of bows and arrows, throwing sticks, dead falls, or snares. Floral resources used included such items as mesquite and screw beans, agave, piñon nuts, cactus fruits, seeds such as chia, and a variety of roots and tubers.

Social groups in the project region consisted of small bands which stayed in temporary camps while they exploited the resources of the surrounding area. It is probable that one or more of the various ethnographic groups that used the region collected salt from Bristol Lake. Both the Chemehuevi and the Mojave Indians sang the Salt Song, which told of an itinerary that took the singer to various places in the Mojave Desert, including this area (Kroeber 1972; Laird 1976). Chemehuevi from the Parker area are reported to have traditionally gathered salt at Bristol Lake (Cultural Systems Research, Inc. 1979:7-26). Kroeber (1972:38) recorded one stop in the Mojave Salt Song as *Selye'aya-kuvataye,* which his informant identified as "sandhills south of Amboy, two deserts [i.e., valley systems] away to the west from the Colorado River at Parker."

PREHISTORY

The archaeological research for the RAIL•CYCLE project was guided by a number of research questions designed to address paleoclimatic changes and their effect on the use of the study area by prehistoric human populations. Among the questions posed were: (1) Was Bristol Lake utilized during the late Pleistocene/early Holocene epochs? (2) How have prehistoric hunter-gatherers used ephemeral lakes in the Mojave Desert since the desiccation of Pleistocene lakes? (3) What systems of exchange with distant cultural groups were operating in the project region? (4) What effect did the eruption of the volcano now known as Amboy Crater have on aboriginal land use in the project region? and (5) To what extent were sparsely occurring lithic materials found on the surfaces of alluvial fans exploited?

One of the assumptions made at the start of the research was that Bristol Lake was once the northern lobe of a perennial Pleistocene lake. In one of the earliest comprehensive topographic and geologic studies of the Mojave Desert, Thompson (1929:696) concluded that Bristol Lake and Cadiz Lake to the southeast were remnants of a former, larger Pleistocene lake, which he named Amboy Lake to avoid confusion with the modern playas. Bristol Lake is one of numerous playas in the Mojave Desert that have generally been considered to have contained standing water during the Pleistocene until their desiccation some 10,000 years ago (Blackwelder 1954; Blanc and Cleveland 1961). However, based on analysis of cores taken from Bristol, Cadiz, and Danby playas, Bassett *et al.* (1959:100-101) concluded that the two playas probably were never united into a major lake, even during pluvial periods, and that at most "these basins contained only very shallow—perhaps ephemeral—lakes that did not overflow during the late Pleistocene epoch." Because sediments in cores from the three playas correlated so poorly, they concluded (1959:111) that "the deposits were irregularly laid down in ephemeral lakes as a result of intermittent floodings." Recent work by Handford (1982), Rosen and Warren (1990), and Rosen (1989, 1991) appears to confirm the findings of Bassett *et al.* (1959) that Bristol Lake has been a playa rather than a lake throughout the Pleistocene epoch.

Given the discrepancies in the literature regarding whether a perennial lake did exist here during the Pleistocene, an abrupt topographic change at the 640-foot contour (which corresponds to the overflow elevation from Bristol Lake into Cadiz Lake) was thought to be a relict beach feature when it was discovered during the initial archaeological survey of the project site. Such a feature could indicate that a Pleistocene lake was present and that archaeological remains on the feature might be associated with ancient cultures, similar to the situation at Pleistocene Lake Mohave (Campbell *et al.* 1937). Further work designed test this possibility resulted in the finding that the "shoreline" feature actually was the result of playa deflation (cf. Blackwelder 1931), thus confirming the recent work by Rosen and others.

Eight prehistoric archaeological sites and eleven isolated artifacts were located and recorded during the archaeological studies on the RAIL•CYCLE project site. All but one of these are related to the procurement of lithic resources from the alluvial fan portions of the project site. Nodules of jasper, rhyolite, and basalt occurring as float in the alluvial fans shed from the Bristol Mountains to the north were "assayed" by prehistoric travelers through the area to determine whether they contained material suitable for the manufacture of stone tools. Many of these nodules were simply tested and abandoned. Others were found to contain useable material and were prepared by trimming away the cortical surfaces and removing the resulting core for further reduction into stone tools elsewhere. In those cases, all that remains are the discarded cortical flakes and an occasional hammerstone.

A prehistoric campsite dated on typological grounds to

8,000-10,000 years ago was discovered at the margin of Bristol Lake near the mouth of a drainage system that enters the lake from the northeast. This site contained evidence that even during the earliest well-accepted periods of Mojave Desert prehistory, people were attracted to ephemeral lake stands in the desert, perhaps by the bloom of brine shrimp and the waterfowl they attracted. While there, broken thrusting-spear points were removed and discarded, and new ones were fashioned. Seeds collected from the surrounding area were ground on milling stones at the site, in an apparently earlier use of milling equipment that has been known from studies in other parts of the desert. Obsidian nodules procured from Tertiary volcanic strata in the Bristol Mountains (Casey 1981) were prepared using bipolar reduction into small cutting tools. Other obsidian artifacts recovered from the site originated in the Coso area in Inyo County and in the Spring Mountains in southern Nevada.

With respect to the questions posed at the outset of the research, it was determined that the Bristol Lake area was indeed used during the late Pleistocene/early Holocene epochs by people attracted to ephemeral stands of the lake. Obsidian artifacts from distant locales suggest that regional trade for obsidian began very early in Mojave Desert prehistory. Due to the deflated nature of the campsite on the margins of Bristol Lake, no evidence regarding the effect of the last eruption of Amboy Crater (Parker 1963) on local human populations was obtained. Finally, the pattern of lithic resource procurement on the alluvial fans of the area suggests that prehistoric prospecting for useable lithic materials was an ongoing search embedded in the daily activities of hunting and gathering.

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The Depositional History of Several Desert Basins in the Mojave Desert: Implications Regarding a Death Valley—Colorado River Hydrologic Connection

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ABSTRACT

Since the turn of the century, many authors have postulated a Plio-Pleistocene connection between the Death Valley-Owens River pluvial system and the Colorado River drainage basin based on the regional distribution of fish species. The most commonly proposed routes involve: 1) an overflowing Death Valley Lake system; or 2) migration of the Mojave River between its present course through Soda Dry Lake and a more southerly route through Bristol Dry Lake. Under the present topographic regime, a Death Valley lake capable of overflowing the bedrock saddle at Ludlow, California (594 masl) and discharging southward into the Bristol, Cadiz and Danby Lake basins (and the Colorado River) would be over 12,000 km^2 in size. However, there is a distinct lack of surface and subsurface indicators to support the existence of either a fluvial or lacustrine connection. Evidence from cores and boreholes drilled up to 536 m deep in Soda, Bristol, Cadiz and Danby Dry Lake basins indicate that neither of these connections has occurred during the last 3.7 ma. B.P. No distinct paleoshorelines, wave-cut terraces, gravel bars, or tufa mounds have been located at elevations corresponding to the above hydrologic systems in Death Valley, Silver-Soda or Danby Lake basins. In the Bristol, Cadiz, Silurian and Broadwell basins, these features have not been found at all. We, therefore, conclude that a hydrologic connection between the Death Valley-Owens River system and the Colorado River has not occurred along either of these routes since the mid-Pliocene.

INTRODUCTION

For almost a century, regional topography and the distribution of vertebrate fish faunas in widely separated, now isolated basins in the Mojave Desert, have fueled a debate as to whether the Colorado River and the Death Valley pluvial system were connected at one time (Blackwelder, 1933 and 1954; Hubbs and Miller, 1948; Miller, 1981; and Hale, 1985). Although this theory has gained wide acceptance and is featured in the Visitor's Center of Death Valley National Monument, the evidence for this connection is minimal. This paper represents part of a larger study concerning this topic (Brown and Rosen; submitted).

During the Pleistocene, three major river systems emptied at one time or another into the Death Valley basin (Fig. 1) (Hubbs and Miller, 1948): 1) the Owens River system draining the eastern Sierra Nevada Mountains; 2) the Amargosa River system draining the Spring Mountains in southwestern Nevada; and (3) the Mojave Rivers system draining the San Bernardino Mountains of southern California. Under the present arid regime, only the Amargosa River discharges into Death Valley and only then during spring runoff or major flooding events. However, there is ample evidence from lacustrine deposits and paleoshoreline features (eg. tufa mounds, gravel bars, wavecut benches) that each of these river systems sustained one or more deep, fresh water to saline lakes for selected intervals during the Pleistocene (Fig. 1). For the purposes of this paper, only the Mojave River system will be discussed in detail.

During the late Pliocene, the Mojave River may have drained westward (Weldon, 1982). By early- to mid-Pleistocene, the Mojave had shifted eastward and began filling, overflowing and breaching a series of downstream basins (Figure 1). Dissected lacustrine sediments near Victorville, California contain Irvingtonian Land Mammal Age (LMA) assemblages and were deposited sometime between ~700 ka. and ~450 ka. B.P.) (Reynolds, 1989). The Manix basin experienced several prominent lake stands between ~350 ka. and ~14 ka. B.P. when Afton Canyon was cut and the basin breached (Jefferson, 1985; Meek, 1989). The exact timing and nature of the first overflow of Lake Manix into the Soda-Silver basins remains somewhat unclear; however, detailed reconstructions of latest Pleistocene Lake Mojave by Wells and others (1989) place the beginning of Lake Mojave at ~22 ka. B.P. The first documented overflow of Lake Mojave and subsequent connection of the Mojave River with Death Valley occurred soon after this event (Wells and others, 1989; Brown and others, submitted). Therefore, prior to the latest

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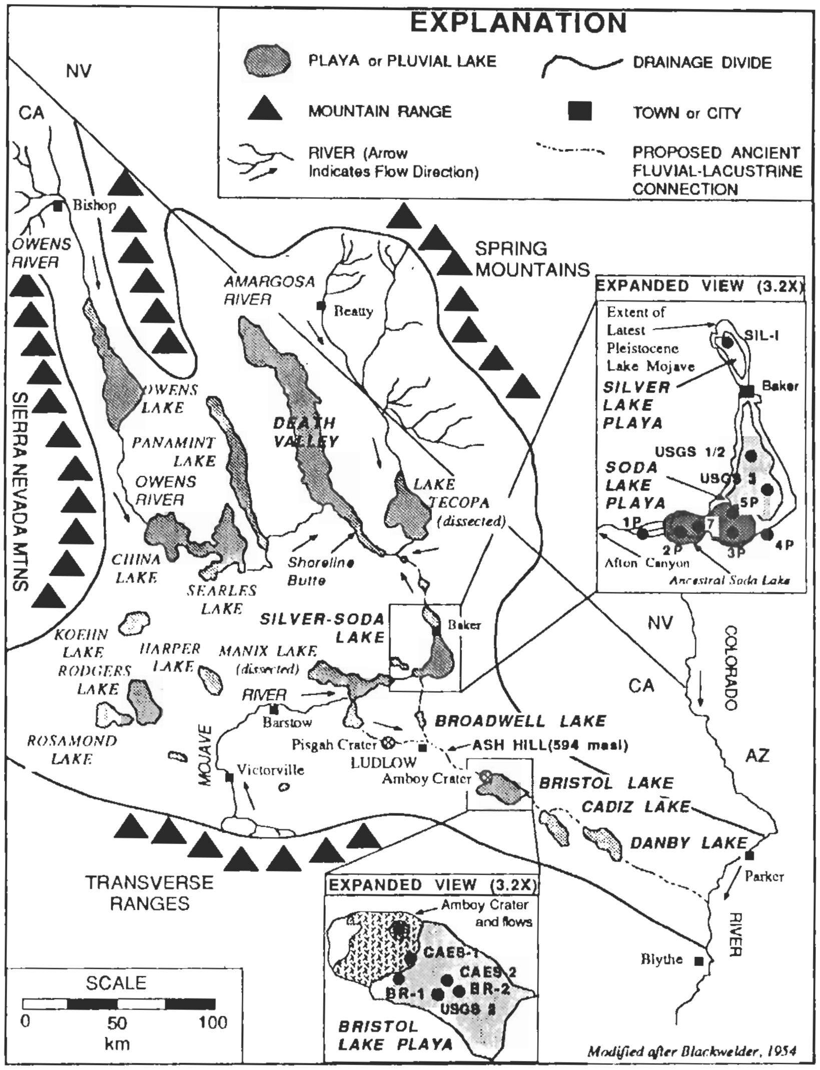


Figure 1. Pluvial lakes and playas of the Death Valley-Colorado River region. Insets of Silver-Soda & Bristol dry lakes show location of boreholes and drillcores discussed in text (after Blackwelder, 1954).

Pleistocene, the Mojave River was not part of an integrated Death Valley drainage network over which fish species could migrate.

PHYSIOGRAPHY OF THE BASINS

Physiographic Setting

The study area lies predominantly within the Basin and Range geographic province of North America which is characterized by northwest-southeast trending, elongate fault-block mountains separated by desert basins (Hunt, 1975). Elevations range from a maximum of over 4400 masl at Mount Whitney, CA (Owens River Drainage Basin) to -87 masl at Badwater in Death Valley. This prominent relief produces significant variations in temperature, precipitation and evapotranspiration within the study area. Under the current arid regime, Furnace Creek in Death Valley receives an average annual precipitation of only 42 mm (Hunt, 1975). During the summer months, temperatures frequently exceed 40°C, resulting in evaporation rates of up to 3800 mm/year (Hunt, 1975). In contrast, the mountainous headwater regions of the Amargosa, Mojave, and Owens river systems characteristically receive over 1000 mm of precipitation annually (Wells and others, 1989).

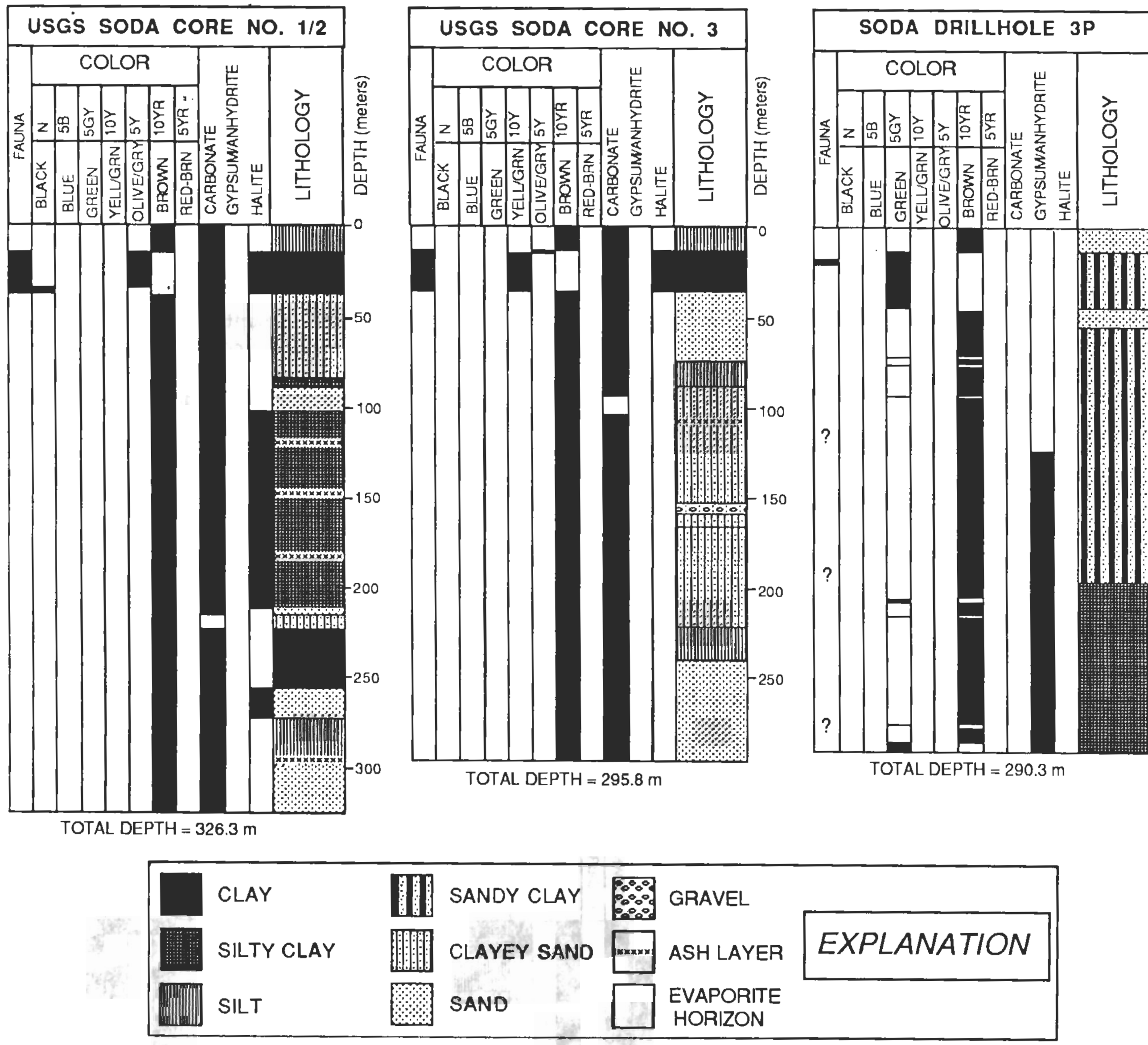
Soda Basin

The Soda Lake basin is a structurally formed basin approximately 30 km long and up to 25 km wide (Fig. 1, inset) and is flanked by the bedrock outcrops of the Soda Mountains to the west, the Cowhole Mountains to the east, and the Mesquite Hills to the south. Geophysical studies and core and drillhole data indicate the basin is asymmetric, and depth to bedrock increases to the south and east where the basin contains over 670 m of fill (Dickey and others, 1979; Negrini and others, unpublished data). Nineteen cores and thirteen boreholes have been drilled in the Silver and Soda Lake basins to depths of over 326 m (Fig. 1, inset) (Meussig and others, 1957; Dickey and others, 1979; Wells and others, 1989; summarized in Brown, 1989). These cores show evidence of a *single* sustained basin-wide lacustrine interval that deposited sediments between depths of 3 to 36 m below the present playa surfaces. These sediments correspond to latest Quaternary Lake Mojave which existed between ~22 ka. B.P. and ~9 ka. B.P. and formed shoreline features at or below 287 masl, over 300 meters below the current overflow height at Ludlow (Fig. 1) (Wells and others, 1989). Lithologic logs for the three deepest drillholes in the Soda Basin are presented in Figure 2. USGS Core 1/2 (326 m deep) and USGS Core 3 (290.6 m deep) were drilled in northern and central Soda Lake (Fig. 1, inset) and encountered oxidized sediments below the Lake Mojave clays characteristic of playa, alluvial fan, and eolian depositional facies (Meussig and others, 1957). Analyses of these deposits indicate that they are lacking in lacustrine faunal assemblages except for the Lake Mojave deposits (Meussig and others, 1957; Wells and others, 1989; Brown and Rosen, submitted).

In southern Soda Lake and along the Mojave River wash (Fig. 1, inset), drill holes 2P, 3P, 5P, and 7 (drilled to depths of 247, 290.3, 151, and 87 meters, respectively) encounter fine-grained, brown-to-green clastic sediments below the latest Wisconsinan Lake Mojave deposits that contain minor to moderate amounts of evaporate minerals (Fig. 2). The nature and distribution of these sediments in drill holes, in conjunction with geophysical studies (Dickey and others, 1979), suggest that a small wet playa-to-shallow lake existed in this location during the early-mid(?) Pleistocene. This "Ancestral Soda Lake" was temporally more persistent, although areally more restrictive, than latest Quaternary Lake Mojave (Fig. 1).

Although no age correlations are available in the Soda Lake basin below the Lake Mojave deposits, average sedimentation rates from nearby basins can provide estimates on the antiquity of sediments penetrated in the above drill holes. Dating of older, compacted sediments in Searles Lake

Figure 2. Simplified core and borehole logs, Soda Dry Lake, California. Explanation of lithologic symbols given in lower left of figure. Sediments must contain >50% evaporite minerals before they are considered an "evaporate horizon." Fauna column indicates that lacustrine fauna are present in core. Original data from USGS Soda Cores 1/2 and 3 (Meussig et al., 1957); original data from Soda Drillhole 3P (Dickey et al., 1979).



(Fig. 1) indicates average sedimentation rates on the order of 22 cm/1000 years during the last 3.18 ma. B.P. (Smith and others, 1983). Tephra correlations and ^{36}Cl dating of sediments in the Bristol basin (Rosen, 1989) indicate sedimentation rates averaged 10 cm/1000 years during the last 3.7 ma B.P. (discussed below). Using these rates as bracketing ranges for deposition in the Soda Basin, the deepest sediments in core 1 are between 1.5 and 3.3 ma. old. The above information suggests that a large, basin-wide lake with overflow to the south has not existed in the Soda Basin since the early Pleistocene or before.

Bristol Lake Basin

A detailed history of the Bristol Lake basin in presented in this guidebook (Rosen, 1992). Six deep cores have been drilled in Bristol Dry Lake (Fig. 1, inset): USGS Bristol Cores 1 and 2, BR-1, BR-2, CAES#1, and CAES#2 to depths of 306.5, 307, 153, 247, 529, and 537 m, respectively (Bassett and others, 1959; Rosen, 1989, 1991) and have been used to reconstruct the depositional history of the basin during the last 3.7+ ma. B.P. (Rosen, 1989). Throughout this period, the Bristol Lake basin has been the site of brackish, ephemeral, shallow water bodies that alternated with halite-precipitating, brine-pond playa conditions (Rosen, 1989, 1991). No evidence exists to support a large basin-wide lacustrine event or a through-flowing fluvial system in the Bristol basin.

Danby Lake Basin and the Bouse Embayment

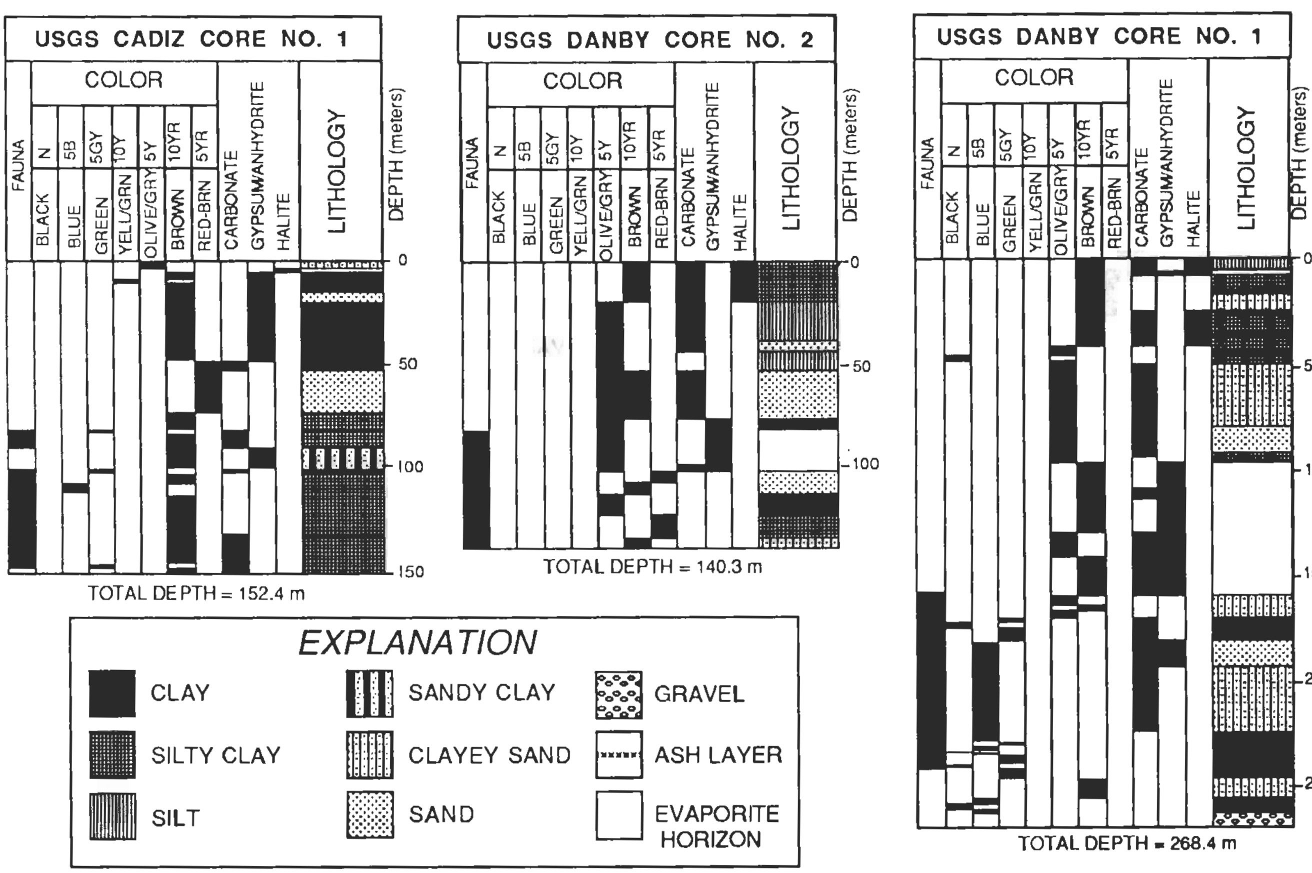
Unlike the Bristol basin, cores drilled in the Danby Basin contain lacustrine and/or probable estuarine deposits at depth. Two deep cores have been drilled in Danby Lake to depths of 268.2 m (USGS Danby Core 1) and 140.2 m (USGS Danby Core 2) (Figs 1 and 3). Interbedded blue, green, and brown, clay- to sand-sized sediments predominate in the lower portions of each of these holes and contain fossil assemblages characteristic of estuarine or brackish water environments (Fig. 3) (P.B. Smith, 1960, 1970). These sediments have been correlated with the Bouse Formation (P.B. Smith, 1960, 1970) which was deposited during the proto-gulf of California marine embayment (Luchitta, 1972) sometime during the mid-late Pliocene. Bouse sediments are found in isolated outcrops in the Blythe-Parker area (Fig. 1). These deposits thicken to the south where they are found in the subsurface. Depositional characteristics and faunal assemblages of the Bouse Formation suggest that Pliocene topography of the region was similar to that of today and that embayment water depth decreased and salinity increased to the north (Lucchitta, 1972; A. Buising, pers. comm., 1990). It is probable that the lower sediments in the Danby cores represent the extreme northwesterly extension of the Bouse embayment.

Overlying the estuarine section in the Danby cores are thick sections of gypsiferous silt which may represent retreat of the Bouse embayment (Fig. 3) (A. Buising, pers. comm., 1990). These sediments are in turn overlain by alternating green, brown, black, and yellow fine-grained clastic sediments with minor amounts of evaporite minerals (Bassett and others, 1959) which are interpreted to have been deposited during closed basin lacustrine-playa phases in the mid to late Pleistocene.

Cadiz Lake Basin

The Cadiz Basin, located between the Bristol and Danby basins, is the smallest of the three southern basins (Fig. 1). A single deep core, USGS Cadiz Core 1, 152.4 m deep, has been drilled in the Cadiz Basin (Bassett and others, 1959) and is shown in Figure 3. This core consists of detrital clays, silts, and sands exhibiting predominantly 10YR (oxidized) colors until the uppermost 3 meters of core where a 2 m thick halite unit is present (Fig. 3). Gypsum is present in relatively minor quantities between 90 and 105 m depth and in the upper 50 m of sediment. Rosen (1989) has suggested that the relative lack of evaporites in Cadiz Basin and the present regional groundwater flow in the area may indicate that during the

Figure 3. Simplified core logs, Cadiz and Danby Dry Lakes, California. Explanation of lithologic symbols given in lower left of figure. Sediments must contain >50% evaporite minerals before they are considered an "evaporate horizon." Fauna column indicates that lacustrine fauna are present in core. Original data from USGS Cadiz Core 1 and USGS Danby Cores 1 and 2 (Bassett et al., 1959).



later part of its existence groundwater flow did not terminate in Cadiz Basin until the prograding fans cut off flow to Bristol Dry Lake. Water was, therefore, unable to reach halite saturation in Cadiz Lake until it became a closed basin.

Shallow, brackish water fossil assemblages have been found in the Cadiz core at depths below 81 m (Bassett and others, 1959; P.B. Smith, 1960, 1970). In addition, sediments analyzed from between 81.4 and 83 m below the playa surface contain rare representatives of a single foraminifera species, *Ammonia beccarii* (P.B. Smith, 1970), which have also been found in the Bouse Formation. It is possible that Cadiz Basin was connected with the Bouse embayment for a short period, however, it is also possible that these foraminifera were accidentally transported via migratory birds from the Danby basin. Evidence for this type of migration is suggested by the presence of the foraminifera *Elphidium* in a 2 m thick package of sediments (43 m depth) from Panamint Core 1 drilled in Panament Valley (Fig. 1) (Smith and Pratt, 1957). Extensive studies of the Owens River chain of lakes (Smith and Pratt, 1957; Smith and others, 1984) support the conclusion that these Owens River basins were not part of a marine embayment during this period of deposition.

DISCUSSION

Our examination of subsurface and surface stratigraphy, sedimentology, and geomorphology in the many basins along these proposed routes does not support either: 1) migration of the Mojave River between its present course and a more southerly route; or 2) an overflowing Death Valley lake system.

The Mojave River was not part of an integrated Death Valley drainage network until after ~22 ka. B.P. when overflow from Manix Lake first reached the Soda Basin in significant quantities. The reconstructed history of the Bristol Basin clearly indicates that during the last 3.7 ma. B.P. neither the Mojave or any other large river system discharged water into the basin. In fact, the sedimentological evidence from Bristol Lake indicates that it has been a closed basin with brackish, ephemeral, shallow water bodies alternating with halite precipitating brine pond playa conditions during this interval (Rosen, this guidebook). Therefore, a shifting in the course of the Mojave River between the Soda-Death Valley and Bristol-Cadiz-Danby basins has not occurred during the last 3.7 ma. B.P.

Evidence for an extensive "Death Valley Lake system" (eg. Hale, 1985) is also lacking. Under the current geographic setting, a lake capable of producing overflow at Ludlow would be over 12,000 km^2 in size and up to 650 m deep. Prominent shoreline features characteristic of such a large, multi-basin pluvial lake system (e.g., Lake Lahontan and Lake Bonneville) have not been found anywhere near the 594 masl elevation currently needed for overflow, nor have they been found at elevations below this level. Hubbs and Miller (1948), Blackwelder (1954), Miller (1981), and Hale (1985) have argued that the absence of surficial fluvial-lacustrine features is due to their antiquity and subsequent erosion. If this is indeed the case, then the lacustrine sediments deposited during this event should still be preserved in the subsurface. Our examination of many of the basins along the proposed route (Fig. 1) indicates that this is not so.

Discounting latest Quaternary Lake Mojave (~22 ka. B.P. to 9 ka. B.P.), the Soda Basin has not experienced a basin-wide lacustrine episode since before the early Pleistocene. Most importantly, extensive coring of the Bristol Basin indicates that it has been a closed basin experiencing saline brind pond conditions since before 3.7 ma. B.P. and although the Pliocene Gulf of California Bouse embayment incorporated the Danby and possibly the Cadiz basins, it was not connected with the Bristol Basin.

Several possible explanations would account for the similarity of fish fauna in the Colorado River and the Death Valley-Owens River system other than a connection through the Bristol and Soda Basins: 1) there may have been a paleo-connection between the two drainage basins, however, it was along a different route; 2) a connection between the two areas did occur, but prior to 3.7 ma. B.P.; 3) eggs of fish living in the Colorado River were transported coincidentally by migratory birds flying between isolated lake basins and river systems. Although the latter seems implausible upon first inspection, accidental transportation of fauna and flora into barren or underpopulated basins is a common process for exploiting new ecological niches and has occurred with foraminifera species in this region (Smith and Pratt, 1956; P.B. Smith, 1960, 1970).

It is apparent that during the last 3.7 ma. B.P., the amount of precipitation entering the study area (Fig. 1) has been insufficient to overcome the physiographic constraints of the region and form a hydrologic connection with the Colorado River drainage basin.

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Pleistocene Faunas in the Bristol-Danby Trough

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ABSTRACT

More than 250 paleontologic localities have been found in the Bristol-Danby trough during the past seven years. The assemblages, at Cadiz, Archer, and Saltmarsh, represent a period of time from the middle? to late Pleistocene. The elevations of the fossiliferous sediments relative to current playa surfaces, combined with observed surface features and satellite imagery, suggest that a southeastward extension of the Bristol Mountain Fault was active during late Pleistocene times.

BACKGROUND

The trough that connects Bristol Lake, Cadiz Lake, and Danby Lake has yielded fossil assemblages of Pleistocene age. Since 1985, the San Bernardino County Museum (SBCM) has monitored excavations and inspected surface exposures in the Bristol-Danby trough and has recorded more than 250 localities that are considered to be late or possibly middle Pleistocene in age.

CADIZ ASSEMBLAGE
SBCM 1.52.1—1.52.195, 1.46.1—1.46.46

Location and Description

The Cadiz localities, at the southwest end of the Fenner Valley (Reynolds, 1991) are found in sediments east of Bristol Lake and north of Cadiz playa, south of Chambless, and north and south of the railroad siding of Cadiz. These distal fluvial deposits were derived primarily from the Fenner drainage and secondarily from the Clipper and Orange Blossom washes. Fenner Valley drains Lanfair Valley (elevation 4000′) and the New York Mountains (elevation 7500′), approximately 45 miles to the northeast. The sediments at Cadiz are pink to gray silts and silty sands interspersed with carbonate kernals and layers of pedogenic carbonate. The section is poorly exposed, but the elevational distribution of paleontologic localities suggests that it may be 80 feet thick. The fauna recovered from the Cadiz localities is given in Table I.

Faunal Assemblage

The faunal assemblage at Cadiz differs from other Irvingtonian and Rancholabrean assemblages in southern California. No single taxon stands out as distinctly representing the Irvingonian Land Mammal Age (LMA) or Rancholabrean LMA. *Bison*, an indicator Rancholabrean taxon, and *Mammuthus* and *Mammut*, common in most Mojave Desert Pleistocene localities, are absent (see Jefferson, 1989 and 1991). *Geochelone* is relatively common in Irvingtonian LMA sediments in southern California (Reynolds and Reynolds, 1990; Reynolds and others, 1991) but is rare or absent in Rancholabrean deposits in California and Arizona (G.T. Jefferson, Page Museum, pers. comm. to RER, 1991; E.H.

Table I. Cadiz Assemblage

Pisidium casertanum	clam
Physa sp.	snail
?*Succinea* sp.	snail
?*Planorbula* sp.	snail
Bufo sp.	toad
Testudinidae	tortoise
Geochelone sp.	giant tortoise
Xantusia vigilis	desert night lizard
Cnemidophorus tigris	western whiptail
Sceloporus sp. (lg)	spiny lizard
cf. *Uta stansburiana*	side-blotched lizard
cf. *Crotaphytus* sp.	collared lizard
Phrynosoma sp.	horned lizard
Pituophis melanoleucus	gopher snake
Coluber/Masticophis	racer/whipsnake
Colubridae (sm)	small colubrid snake
Colubridae (med)	medium-size colubrid snake
Crotalus sp.	rattlesnake
Aves (sm)	small birds
Aves (med)	medium-size bird
cf. *Lepus* sp.	jack rabbit
cf. *Sylvilagus* sp.	cottontail
Leporidae	medium-size rabbit
Sciuridae (sm)	small squirrel
cf. *Eutamias* sp.	chipmunk
Spermophilus (Xerospermophilus) sp. or *S. (Ictidomys)* sp.	ground squirrel
Thomomys bottae	Botta's pocket gopher
Thomomys bottae?	gracile gopher
Perognathus sp. (sm)	small pocket mouse
Perognathus sp. (lg)	large pocket mouse
Dipodomys sp. (sm)	small kangaroo rat
Dipodomys sp. (?med)	medium-size kangaroo rat
Dipodomys sp. (lg)	large kangaroo rat
Peromyscus sp. (lg)	large deer mouse
Peromyscus sp. (sm)	small deer mouse
Neotoma sp. (sm)	small wood rat
Neotoma sp. (lg)	large wood rat
cf. *Canis* sp. (lg)	large dog
Felidae (med)	medium-size cat
?*Tetrameryx* sp.	medium-size pronghorn
?*Capromeryx* sp.	small pronghorn
cf. *Camelops* sp.	large camel

Lindsay, Univ. of Arizona, pers. comm. to RER, 1992), although Rancholabrean records are common in Florida and Texas (Parmley, 1986; Holman, 1969, 1978; Milstead, 1986; Moodie and VanDevender, 1979).

The gracile *Thomomys bottae?* resembles a form found in the late Irvingtonian deposits of Murrieta (Reynolds and Reynolds, 1990; Reynolds and others, 1991). The medium-sized leporid is similar morphometrically to specimens from deposits considered to be Irvingtonian LMA in the west central Mojave Desert (Reynolds, 1989).

The small pronghorns *Capromeryx* and *Tetrameryx* are both known from the Irvingtonian LMA in southern California, but *Tetrameryx* is rare in latest Rancholabrean deposits (Kurten and Anderson, 1980; Savage and Russell, 1983). The medium-sized felid is an unidentified cat not known from the Rancho La Brea deposits (Jefferson, pers. comm. 1992).

In general, the fauna from Cadiz suggests an early Rancholabrean or late Irvingtonian LMA. The presence of *Pisidium casertanum* suggests the presence of standing water; the cricetines *Neotoma* and *Peromyscus* (small and large), and the *Sylvilagus* sp. suggest brushy and grassy habitat. Deposition may have been in a marshy discharge area at the distal end of the Fenner Valley drainage.

ARCHER ASSEMBLAGE
SBCM 1.42.2 — 1.42.5

Location and Description

The Archer localities are approximately 8 miles southeast of Cadiz on the southern flanks of the Ship Mountains, at an elevation similar to that of sediments at Cadiz. The Archer sediments are lithologically distinct, however, perhaps due to their proximity to the Ship Mountains. The section consists of approximately 4 feet of sands and angular gravel clasts which fine upward into an iron oxide-stained horizon with root casts and carbonate kernals that may have been a stable surface with a poorly developed paleosol (Fig. 1). This surface is capped by a 2 feet thick layer of calcium carbonate. The abruptness of the contact does not suggest a pedogenic carbonate, but rather may indicate carbonate deposition by ground water discharge. The fauna (Table II) was recovered below the carbonate layer in the silty sands that contain the root casts.

Faunal Assemblage

The fauna contains no extinct elements, and all taxa with the exception of the *Peromyscus* sp. are probably present in the near vicinity today. This suggests that the sediments represent latest Pleistocene deposition, and the overlying carbonate layer indicates hydrologic regimes that were not active in recent times.

Table II. Archer Faunal Assemblage

Plantae	root casts
Succinea sp.	land snail
Phrynosoma sp.	horned lizard
P. platyrhinos	desert horned lizard
Boidae	boa
Colubridae	colubrid snakes
Aves sp. a	bird
Aves sp. b	bird
Aves sp. c	bird
Passerina cyanea amoena	common bunting
Passerculus sandwichensis	Savannah sparrow
Carduelinae	finch
Vespertilionidae	small bat
Lepus sp.	jack rabbit
L. californicus	black-tailed jack rabbit
Sylvilagus sp.	cottontail
Thomomys bottae	Botta's pocket gopher
Perognathus sp. (lg)	large pocket mouse
Perognathus sp.	pocket mouse
Dipodomys sp.	kangaroo rat
D. ordii	Ord's kangaroo rat
Peromyscus sp.	deer mouse
Neotoma sp.	wood rat
N. cf. *N. lepida*	desert wood rat
Reithrodontomys megalotis	harvest mouse

SALTMARSH ASSEMBLAGE
SBCM 1.41.1 — 1.41.8

Location and Description

The Salt Marsh locality is in Ward Valley, on the northeast margin of Danby Lake, approximately 8 miles northwest of Highway 62 and 9 miles southeast of the site of Milligan. Sediments consist of tan playa silts overlain by a soil containing carbonate kernals and root casts (Fig. 2). Hematite-stained, stabilized, cross-bedded dunes are inset within these silts. Vertebrate fossils (Table III) occur in deflation areas below the dunes.

Faunal Assemblage

The *Camelops* sp. and small horse are extinct taxa typical of Rancholabrean LMA faunas described from southern California (Jefferson, 1989, 1991; Kurten and Anderson, 1980). The remainder of the taxa are known from the late Pleistocene and are also found in the area today.

Table III. Saltmarsh Assemblage

Plantae	root casts
Lepus sp.	jack rabbit
L. californicus	black-tailed jack rabbit
Dipodomys sp.	kangaroo rat
D. sp. cf. *D. deserti*	desert kangaroo rat
D. cf. *D. merriami*	Merriam's kangaroo rat
Vulpes sp.	fox
Taxidea taxus	badger
Equus sp.	horse
Equus sp. (sm)	small horse
Camelops sp.	large camel
Artiodactyla (sm)	deer-size artiodactyl

TOPOGRAPHY

The relative elevations of fossil localities in the Bristol-Danby trough compared to present-day playa surfaces are shown in Figure 3.

The Saltmarsh locality is at elevation 640′ in sediments that are no more than 20 feet above the surface of Danby playa. Danby Lake is separated from Cadiz and Bristol playas by the metamorphic rocks of the Kilbeck Hills.

Sediments at the Archer locality, at elevation 750′, are approximately 200′ above the surface of Cadiz Lake. The Archer sediments are within the elevation range of vertebrate fossil localities at Cadiz. The Cadiz localities span a 100′ range in elevation, from 720′ to 820′. The elevation of the lowest locality is approximately 100 feet above the surface of Bristol playa (610′), which lies to the west.

STRUCTURE

It is interesting to note that Bristol and Danby playas are at approximately the same elevation (610′ and 620′, respectively), and that Cadiz playa is approximately 70 feet lower in elevation than either. Had there been a through-going drainage in the Bristol-Danby trough (Gardner, 1980; Thompson, 1929), today's elevations indicate that Cadiz Lake, not Danby, would have received the cumulative drainage.

The difference in elevation may be due to structures beneath the playas. Bishop (1963) maps a northwest-trending fault on the north side of the Iron Mountains that strikes towards the Kilbeck Hills and roughly toward the Bristol Mountains Fault, part of the Eastern California Shear Zone (Howard and Miller, 1992). Satellite imagery provided to the SBCM by Dr. John Ford (Jet Propulsion Laboratory, California Institute of Technology) shows a lineation which may indicate the continuation of this structure beneath the Kilbeck Hills to the Bristol Mountains. Surficial expressions of a hypothetical fault along strike of this lineament include the outcrop of Bolo Hill; "ruins" which include wells for fresh water; locally flourishing, large creosote bushes; and a series of dissected bluffs with relief of more than 10 feet that contain fossiliferous silt and carbonates. The bluffs may be eroding on the "up" side of a lateral fault. Agricultural projects draw fresh water from Fenner Valley on the east side of this lineation, and Rosen (this volume, Fig. 1) indicates a boundary between the fresh water at Cadiz and the saline water in Bristol basin.

Taken together, these imagery, surface, and hydrologic features suggest that the Bristol Mountains Fault might continue southeastward and, since it cuts fossiliferous sediments of middle? to late Pleistocene age, that the fault was active at least into late Pleistocene times. Such a structure may also, in part, be responsible for the difference in elevations between Cadiz and Danby lakes.

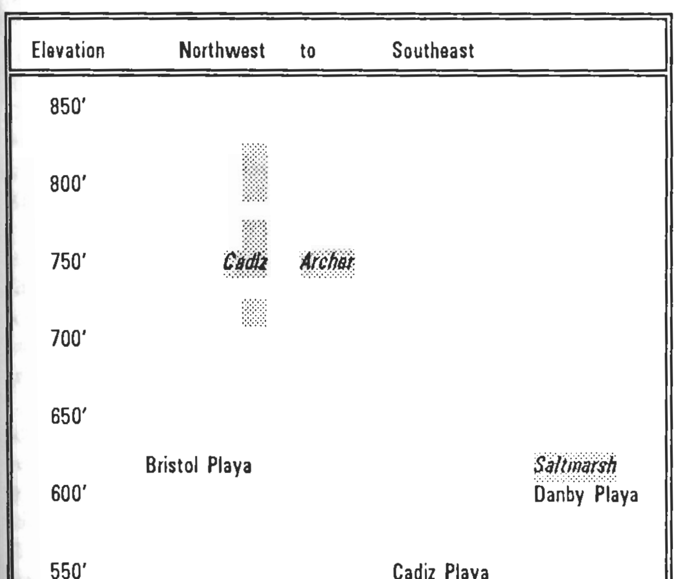

Figure 3. Distribution of faunal assemblages. Fossil localities are indicated by shading.

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Geology and Saline Resources of Danby Playa, Southeastern California

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INTRODUCTION

Danby Playa (or dry lake) is 38 mi southeast of Amboy, California in a large, northwest-trending structural trough that also includes Bristol and Cadiz Dry Lakes (Map 1). Danby Playa is approximately 2 to 3 mi wide by 14 mi long and is bounded by the Old Woman Mountains piedmont to the north, by the Iron Mountains piedmont to the south and west, and by the Turtle Mountains piedmont to the east. The playa surface is about 620 ft above sea level and is separated from Cadiz Dry Lake by an alluvial divide approximately 500 ft higher than Danby Playa. Thompson (1929) concluded that Danby Playa is the sump of a large drainage basin that includes Ward Valley and adjacent upland areas.

Saline resources are present in the northwest, central, and southeast areas of Danby Playa (Map 1) and have been explored by numerous shallow (generally less than 40 ft deep) test wells. Two deeper wells, DAN-1 and DAN-2, were drilled in the southeast and northwest areas, respectively, to determine the chemical characteristics of the saline resources at depths greater than 40 ft. This report describes the geology and the saline resources of Danby Playa.

GEOLOGY

Ver Planck (1957) reported that the surface of Danby Playa is a mud flat with minor relief that includes an elongated low area underlain by a mixture of mud, salt, and brine (Qhb on Map 1); the low area is often covered with a white saline efflorescence and(or) a salt crust approximately 1 in. thick. Thompson (1929) reported that the surface of the playa is cut by numerous small drainage channels 4 to 5 ft deep and up to 100 ft long that slope to the south; small, gypsum-capped tabular hills up to 12 ft high near the southern end of the playa are erosional remnants of a former lake surface (Blackwelder, 1931). Combined, these features suggest that the surface of Danby Playa has been lowered by regional structural tilting (Thompson, 1929), by the interaction of wind deflation and water erosion (Blackwelder, 1931), or by a combination of these processes as well as differential compaction, solution, and redeposition of salt deposits (Bassett and Kupfer, 1964).

The late Pleistocene (Bassett and others, 1959) lacustrine deposits that underlie Danby Playa consist of lenticular beds of very fine- to medium-grained sand, clayey silt, silty and sandy clay, and massive clay; gypsum and halite are present throughout this sequence (Bassett and others, 1959; Calzia, 1991). The lacustrine deposits are 380 ft thick in the southeast area, more than 500 ft thick in the northwest area, and overlie fine- to coarse-grained sand, gravel, silt, and clay (Calzia, 1991). The homogeneous and monotonous nature of the thick lacustrine deposits, the abundance of discontinuous beds that cannot be correlated between test wells, and (save for one locality east of Saltmarsh) the absence of shore line features, wave-cut terraces, and(or) gravel bars, suggest that these sediments were deposited in a series of shallow ephemeral lakes over a long period of time (Bassett and others, 1949; Bassett and Kupfer, 1964).

Smith (1960, 1970) described foraminifers associated with marine and brackish water mollusks, barnacles, and ostracods in a thick (300 ft) interval of sand, calcareous silt, and laminated blue and green clay 400 to 500 ft below the surface of Danby Playa. She concluded that this fossil assemblage represents marine conditions and correlated the host sediments with the late Miocene and Pliocene (Lucchitta, 1979) Bouse Formation. Northwest-trending gravity profiles across the playa suggest that the relatively low density lacustrine deposits and the underlying Bouse Formation are approximately 2800 ft thick within the northwest-trending structural trough (Calzia and others, 1979).

SALINE RESOURCES

The saline resources of Danby Playa consist of gypsum ($CaSO_4.2H_2O$), halite (NaCl), and chloride-dominated brines. Gypsum is present as disseminated crystals and as several thick (one more than 200 ft thick) beds in the lacustrine deposits; halite is found in clay-rich beds in the lacustrine deposits and as coarse-grained crystals in large, discontinuous crystalline salt bodies (Bassett and others, 1959; Ver Planck, 1957, 1958). Relatively rare bassanite ($CaSO_4.5H_2O$) and mirabilite ($Na_2SO_4.10H_2O$) crystals were found in clay beds near the northern and central area of the playa, respectively (Allen and Kramer, 1953; Ver Planck, 1957).

The largest crystalline salt body is located in the southeast area of Danby Playa and consists of numerous tabular halite beds 0.5-3 ft thick. This salt body underlies a 2 to 3 mi^2 area and is 1 to 10 ft thick (Fig. 1). Assuming a density of 2.16 gm/cc, this salt body contains 13.75x10^6 tons of salt at an average grade of 92.2-04.9 percent NaCl (Ver Planck, 1958). A resource estimate for the salt bodies in the central and northwest areas is difficult to calculate because the salt deposits are discontinuous, porous, and contain clay interbeds. Nevertheless, Ver Planck reported that the central salt body contains 0.8 to 0.9x10^6 tons of salt at an average grade of 73-84 percent NaCl; the northwest salt body contains 15-20x10^6 tons of salt.

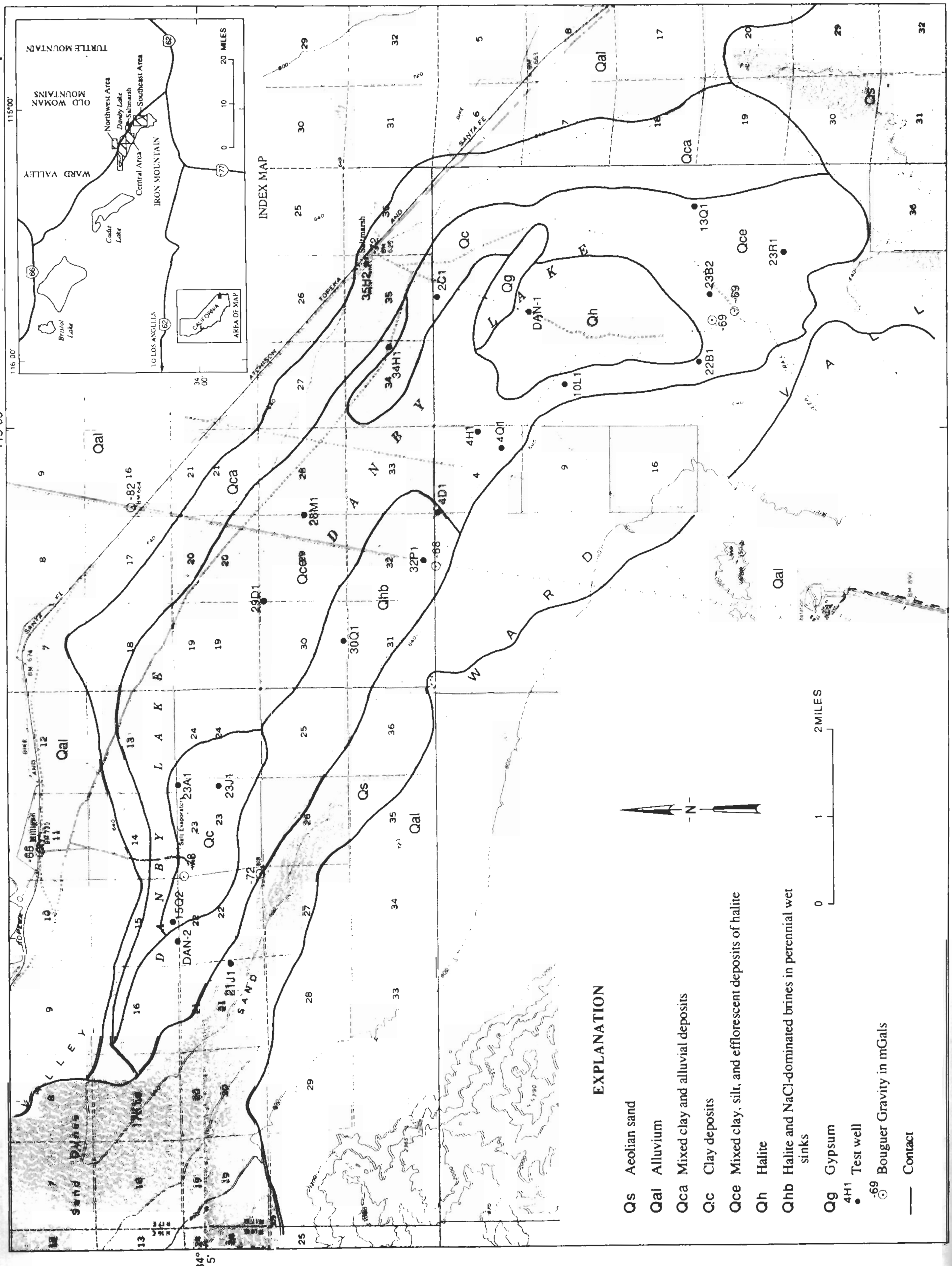
Brines from the crystalline salt bodies are dominated by sodium chloride with lesser amounts of sodium sulfate; the concentration of sodium sulfate increases as the brines precipitate halite during evaporation. Bitterns from Danby Playa contain much less calcium chloride than evaporated brines from Bristol and Cadiz Dry Lakes; in general, the calcium chloride concentration increases to the northwest and

Figure 1. Cross sections and isopach map of crystalline salt deposits on Danby Playa. Cross sections from Verplanck (1958);m contour interval is 2 ft.

Chemical analyses of brines from Danby Playa. All concentrations in ppm.

Well no. (fig. 1)	Total depth (ft)	pH	Temp °C	Hardness (a)	Ca	Mg	Na	HCO3	SO4	Cl	TDS (b)	Reference
							Northwest Area					
DAN-2	503	7.6	23.8	500	82	72	98,000	171	31,000	140,000	187,000	Calzia and others (1979)
15Q2	21	8.1		2,660	638	259	95,200	265	20,700	33,000	251,000	Moyle (1967)
17K1	34	7.9		2,070	500	200	111,000	200	29,700	150,000	292,000	Moyle (1967)
21J1	21	8.1		2,210	483	245	105,000	233	23,400	46,000	275,000	Moyle (1967)
23A1	41	7.5		1,550	575	27	59,400	78	4,070	89,700	154,000	Moyle (1967)
23J1	32	7.9		2,200	729	92	127,000	90	10,600	189,000	328,000	Moyle (1967)
							Central Area					
28M1	41	8.0		1,170	375	204	105,000	162	21,500	148,000	276,000	Moyle (1967)
29D1	41	8.1		2,570	498	324	109,000	232	32,200	146,000	288,000	Moyle (1967)
30Q1	41	7.9		4,810	1,600	195	119,000	65	6,030	182,000	309,000	Moyle (1967)
32P1	31	7.7	26.7	1,190	662	81	10,500	60	1,140	16,500		Moyle (1967)
34H1	41	8.3		375	109	25	14,300	227	3,740	19,400	38,000	Moyle (1967)
35H2	594			27	7.1	2.2	150	110	114	93	482	Moyle (1967)
							Southeast Area					
2C1	51	8.0		1,920	465	185	89,000	93	17,000	126,000	233,000	Moyle (1967)
4D1	41	8.0		5,030	1,560	277	51,600	143	7,470	77,600	139,000	Moyle (1967)
4H1	41	8.0		4,900	1,620	208	56,000	129	8,230	84,900	152,000	Ooyle (1967)
4Q1	21	8.0		1,680	540	81	40,200	104	3,420	60,600	105,000	Moyle (1967)
10L1	37	7.7		2,640	979	47	123,000	51	3,400	189,000	317,000	Moyle (1967)
13Q1	40	8.1		2,540	775	148	15,900	283	3,950	23,300	44,500	Moyle (1967)
22B1	21	7.9		3,400	1,010	212	46,800	110	5,620	70,300	134,000	Moyle (1967)
23B2	56	7.5		3,200	1,060	136	124,000	45	3,400	192,000	321,000	Moyle (1967)
23R1	32	8.3		3,200	146	22	10,700	181	2,960	14,600	28,800	Moyle (1967)
DAN-1	504	5.8	24.3	1,400	370	110	81,000	50	3,800	130,000	172,000	Calzia and others (1979)

(a) As CaCO3.
(b) Total dissolved solids at 180°C



Map 1. Geologic map of Danby Playa (from Bassett and Kupfer, 1964).

the sulfate content decreases to the southeast across Danby Playa. Although brines in the southeast and the central areas are as saline as brines in the northwest area, salinity decreases with distance and depth from the crystalline salt bodies (Calzia, 1991). For example, brines in the northwest area contain approximately 25 percent NaCl and generally less than 2 percent Na_2SO_4 (Table 1); brine from a test well along the northeastern side of the playa contains less than 5 percent total dissolved solids (Ver Planck, 1958). This spatial relation suggests that the brines are produced by the interaction of crystalline salt deposits with groundwater imported from the Ward Valley drainage basin. Assuming the crystalline salt deposits are not removed by mining, the brines are a renewable resource within the limits of the regional hydrologic cycle.

Ver Planck (1957, 1958) reported that production of the saline resources from Danby Playa has been sporadic since the late 19th century. The crystalline salt deposits were mined by the Metropolitan Water District of Southern California and by National Chloride Company from the late 1940s to the mid-1980s for water softeners in the Los Angeles and southern California areas. Currently, salt is produced from brines pumped into 5-acre solar evaporation ponds by Salt Products Company for water softeners in Arizona and for livestock in southern California (Richard R. Gundry, geologist, BLM California Desert District, personal communication, 1992).

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Miocene Vertebrates in the Little Piute Mountains, Southeastern Mojave Desert

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INTRODUCTION

Although mountain ranges in the southeastern Mojave Desert contain considerable thicknesses of Miocene sedimentary rocks, vertebrate fossils of the same age are rare. Two localities within the Little Piute Mountains contain vertebrate fossils and ichnofossils of Miocene age. These fossils offer a sense of Miocene vertebrate communities in the region and aid in the reconstruction of past climatic conditions and depositional environments.

GEOLOGIC SETTING

The Little Piute Mountains are located at the northeast end of the Old Woman Mountains in eastern San Bernardino County, southeastern California (Fig. 1). The range contains Miocene sedimentary and volcanic strata and structures which record extensional tectonism that affected much of the region during the mid-Tertiary (Howard and John, 1986; Knoll, 1986; Nielson, 1986; Hileman and others, 1990). Rock fragments in the Tertiary and Quaternary sedimentary units of the Little Piute Mountains are of local provenance and were derived primarily from metamorphic and igneous rocks of the adjacent Old Woman Mountains (Fig. 2) as they were uplifted and denuded. Provenance and paleocurrent directions indicate that sediment was shed east from the uplifted source into basins of the Little Piute Mountains, where it was deposited mainly as coalescing alluvial sheets and braid plains emanating from the Old Woman Mountains highland (Knoll, 1988).

STRATIGRAPHY

An extensive record of Miocene basin filling is found within the southern Little Piute Mountains, where over 1100 m of Tertiary strata are exposed (Fig. 3). Basin fill can be divided into two sections based upon the presence of the Peach Springs Tuff, a felsic ignimbrite emplaced about 18.5 ± 0.2 Ma (Nielson and others, 1990). The basal unconformity is defined by coarse sandstones and conglomerates resting on an erosional surface developed on Cretaceous granodiorite, Paleozoic metasedimentary rocks, or Precambrian gneiss (Fig. 3). A variation in thickness of the pre-Peach Springs Tuff section of 0-800 m attests to considerable paleotopography on the original depositional surface. An approximately 200 m thick volcanic section composed mainly of alkali, olivine-bearing basalt flows, subordinate interbedded volcaniclastic sandstones and lahars, and rare andesite flows underlies the Peach Springs Tuff (Fig. 3) (Knoll, 1986; Miller, 1986).

The upper, post-Peach Springs Tuff section, 390 m thick, is dominated by sandstones and conglomerates with minor siltstone and a single air fall tuff (Fig. 3). The petrology and stratigraphic position of this tuff suggest that it may be the distal air fall deposit of the 17.8 ± 0.15 Ma Wild Horse Mesa Tuff of the Providence Mountains to the north (McCurry and others, 1989).

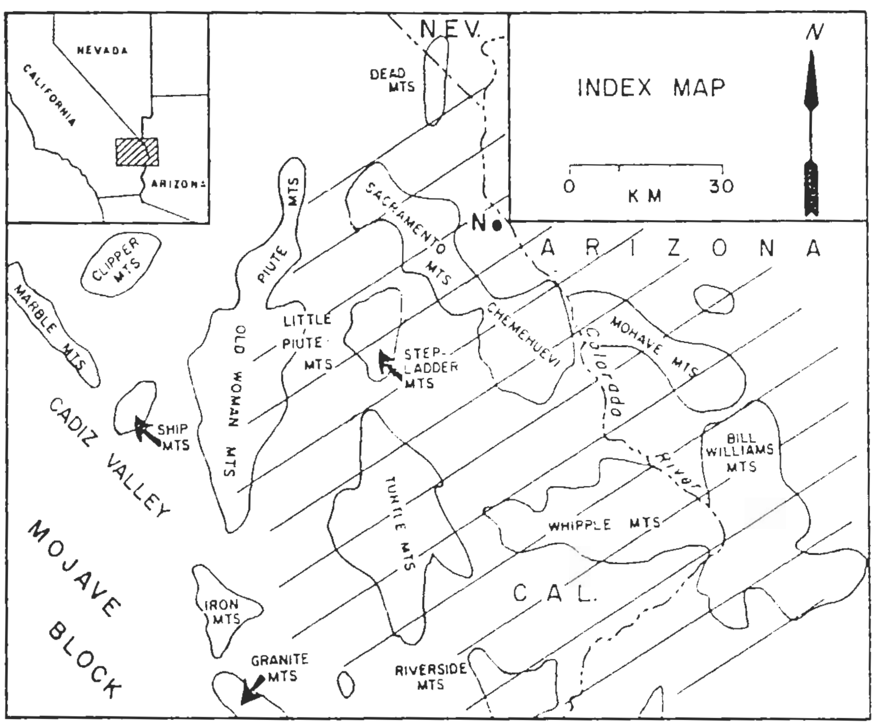

Figure 1. General location map. Ruled pattern is the highly extended Whipple Detachment Terrane. N = Needles.

FOSSIL-BEARING STRATA

Vertebrate fossils are found in strata above the Peach Springs Tuff; fossil tracks are found in strata beneath this tuff.

Ichnofossils

Vertebrate trackways are found in a single locality known as the "footprint gully" (Fig. 2), 115 m beneath the Peach Springs Tuff. Here, a 3 m thick, red-colored claystone exposes a single bedding plane containing the trackways of lower

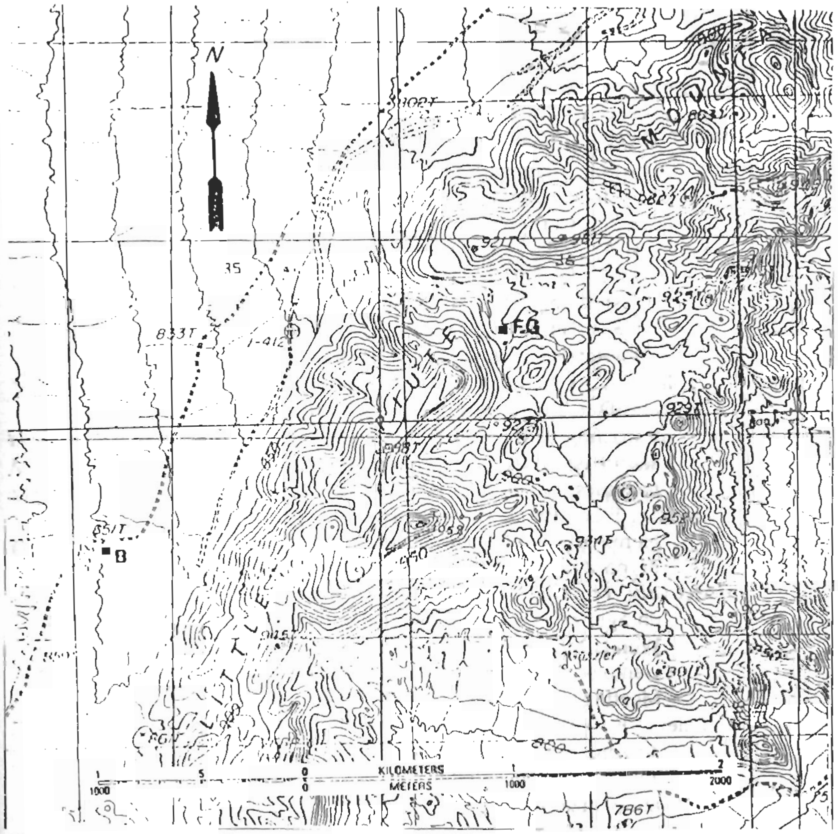

Figure 2. Topographic map of a portion of the southern Little Piute Mountains. FG = "footprint gully", B = vertebrate fossil locality. (from USGS 7.5' Little Piute Mts, prov. 1985).

The only reliable age constraint for this unit is provided by the 18.5±0.2 Ma Peach Springs Tuff, which lies 115 m upsection. The unit lies 10.5 m above Paleozoic meta-sedimentary basement rocks. Knoll (1988) estimated that Tertiary sedimentation began in the region approximately 19 to 23 Ma. Thus, the age of the trackways should fall between 18.5 Ma and 23 Ma.

Vertebrate Fossils

The vertebrate fossil locality in the Little Piute Mountains is approximately 2 km southwest of the "footprint gully" (Fig. 2), in a 12 m thick, fine- to medium-grained, massive sandstone 140 m stratigraphically above the Peach Springs Tuff (Fig. 3). The fauna is listed in Table I.

Table I. Little Piute Mountains Assemblage

Proheteromys sulculus	pocket mouse
Perognathus sp. cf. *P. furlongi*	pocket mouse
Hesperocamelus sp.	large camel
Michenia sp. cf. *M. agatensis*	small camel
Merychippus sp. (sm)	horse

Merychippus sp. (sm) was reported by Whistler and Tedford (pers. comm. to Reynolds, 1985); Dr. James Quinn identified the camels.

Miocene artiodactyls, shore birds, and a possible reptile. The most numerous tracks are of an artiodactyl, possibly a large cervoid (Fig. 4a). The shore bird trackways may represent the families Scolopacidae or Charadriidae. One unusual trackway is from an animal with four toes. The individual digits diverge broadly, each making a distinct depression of approximately equal size. The length and divergence of the digits eliminates Carnivora as an ichnofamily, and because perissodactyls have an odd number of digits, the tracks were most likely left by artiodactyls (Fig. 4b). Four-digit artiodactyls include pig-like forms and oreodonts. The symmetry of the digits from the Little Piute Mountains trackways suggest that the tracks were made by an oreodont (Romer, 1955, fig. 342), possible a mid-size genus such as *Merychyus* sp.

The fine-grained nature of the sediments, the presence of mud cracks and ripple laminations, and shore bird trackways indicate deposition in an ephemeral water body in the distal portions of a fan complex (Knoll and others, 1985).

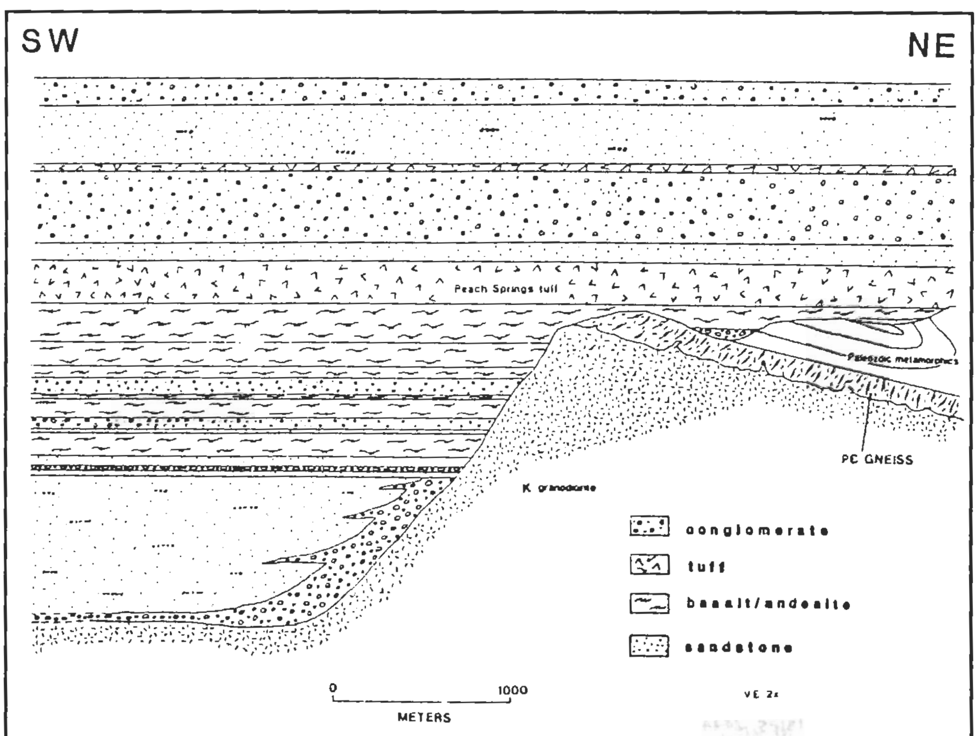

Figure 3. NE - SW cross section through the southern Little Piute Mountains.

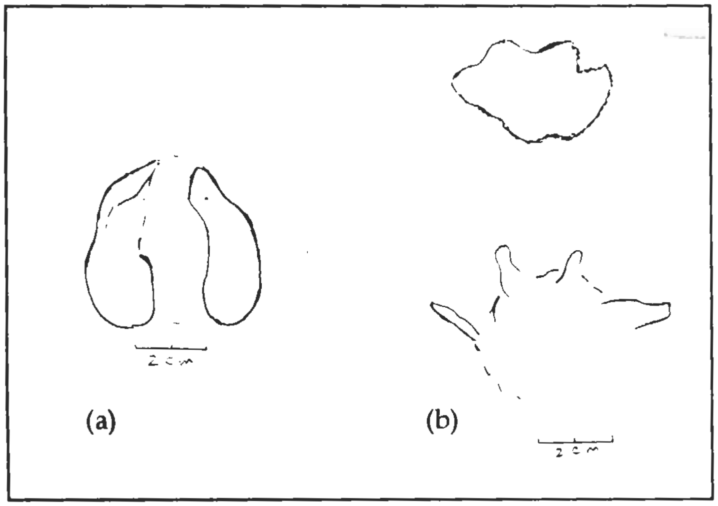

Figure 4. Sketches of tracks of "footprint gully". (a) artiodactyl (cervid?); (b) oreodont?

Michenia agatensis is known to occur in Hemingfordian LMA deposits in the Cady Mountains (Miller, 1980) and in the Crowder Formation (Reynolds, 1991a). The genus was described by Frick and Taylor (1971) who indicate it is known from early Hemingfordian to late Clarendonian times. *Hesperocamelus (Aepycamelus)* is known from the Hemingfordian and Barstovian deposits in Cajon Pass and at Barstow (Davidson, 1923; Miller, 1980; Reynolds, 1991b).

Perognathus furlongi is known from Hemingfordian and stovian deposits in and around the Mojave Desert (Reynolds, 1991b; Lindsay, 1972; Whistler, 1991a, 1991b). *Proheteromys sulculus* is known from the Cajon and Crowder formations (Reynolds, 1991a) and from Barstow (Lindsay, 1972), and its usefullness as a Hemingfordian LMA indicator is discussed by Reynolds (1991b).

The age of this material is constrained by the underlying 18.5 Ma Peach Springs Tuff and possibly by an air fall tuff which lies 36 m upsection. This tuff is probably the distal equivalent of the 17.8 ± 0.15 Ma Wild Horse Mesa Tuff (McCurry and others, 1989) of the Providence Mountains.

CONCLUSIONS

The vertebrate fossils and trackways exposed in sedimentary rocks of the Little Piute Mountains provide a rare glimpse of Miocene, Hemingfordian LMA faunal assemblages in the eastern Mojave Desert. The presence of grazing animals suggests a paleoenvironment with grasslands; shore bird tracks indicate that water was present locally.

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Geological Overview of the Turtle Mountains

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HISTORY

The course of the 1992 Mojave Desert Quaternary Research Symposium field trip leads east, skirting the southern tip of the Turtle Mountains along Highway 62, near the abandoned Second World War military airfield and railroad siding of Rice (Fig. 1).

Lying in one of the most remote, hottest, and driest regions of the Mojave accessible to civilian travel, the Turtle Mountains remained a geological mystery until recent times. Yet the history of human interaction with this region is long.

The range was frequented by Halchidoma, Chemehuevi, Mojave and other native peoples into the present historical era. These tribal groups hunted bighorn sheep and other game, establishing camps at Mopah Spring and similar water holes throughout the range. Arrowheads, fragments of pottery and other artifacts may still be found near these sites–now all protected and still culturally important to local natives. Petroglyphs carved in desert varnish adorn boulders lying on the slopes of Vidal Valley, a reentrant in the southeastern part of the Turtle Mountains. To these were added carvings by cattlemen who first visited this country in the late 19th century.

PREVIOUS STUDIES

Surveying in this region may actually have begun with Spanish exploration, though this is uncertain. While tales persist of buried Spanish treasure in the northern part of the Turtle Mountains, the first authenticated surveys in this region began with the Whipple Expedition in the 1860s. Mapping continued as the U.S. Army established a set of fortified posts across the desert, and as rail lines were graded. In 1954 the U.S. Geological Survey developed a topographic map of the Turtle Mountains from plane tabling and multiplex air photogrammetry (Turtle Mountains Quadrangle, 1:62,500). By this time the area lay under the administration of the U.S. Bureau of Land Management.

While prospectors had long been active throughout the Turtle Mountains, especially between the world wars, the first geological study was not published until 1949, when Chesterman described a spectacular swarm of Tertiary dikes and plugs marking the site of a once vigorously active volcanic field at the northern extremity of the range. A few years later, Wright and others (1953) cataloged large test pits and active mines for the California Division of Mines and Geology. They noted mining of perlite glass from vitric lava flows, and of lead, zinc, copper, and silver from crystalline rocks. Their study came near the end of the heyday of mining operations in the Turtle Mountains, though interest in marginal deposits of placer gold have continued up to the present (Howard and others, 1988).

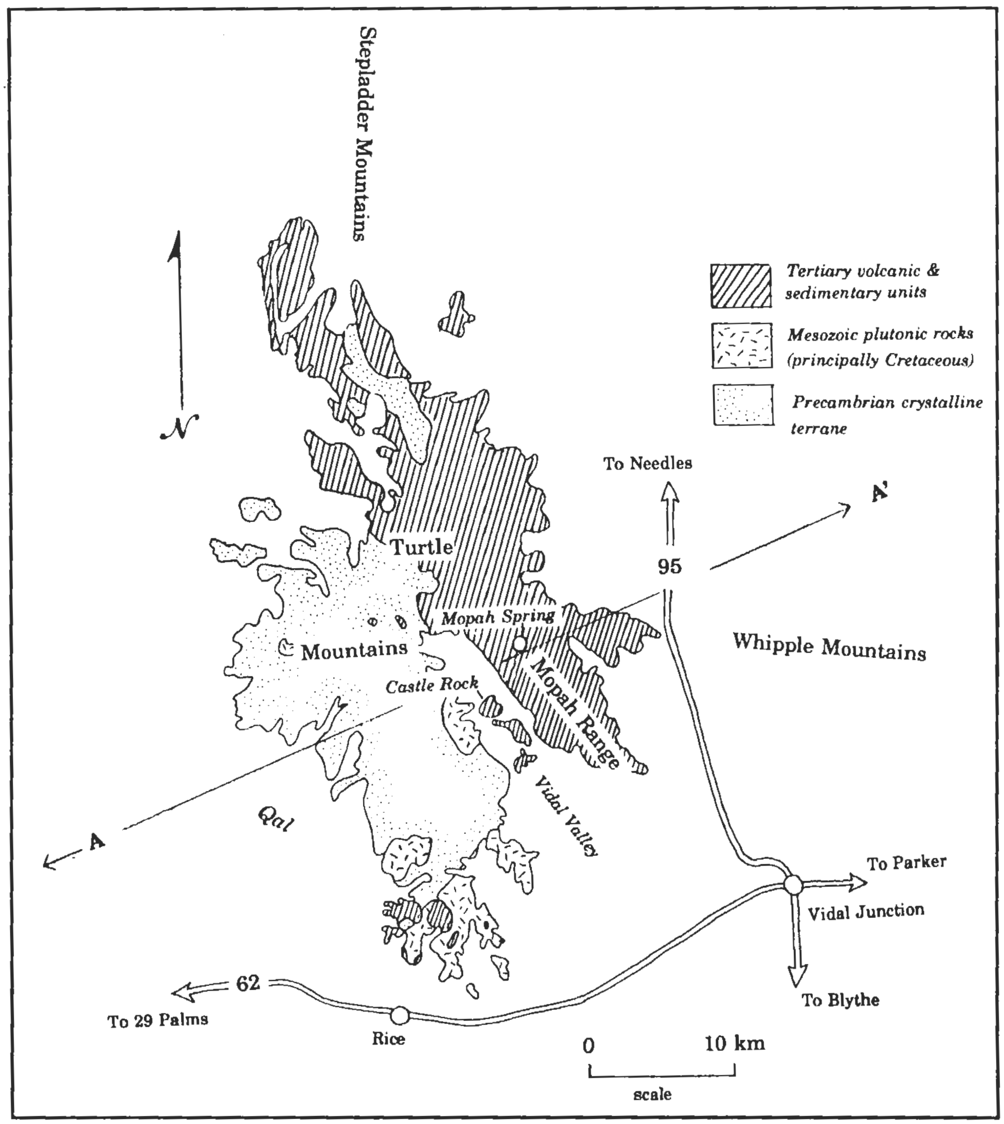

Figure 1. Generalized geological map of the Turtle Mountains. (Line of cross section shown in Figure 2.)

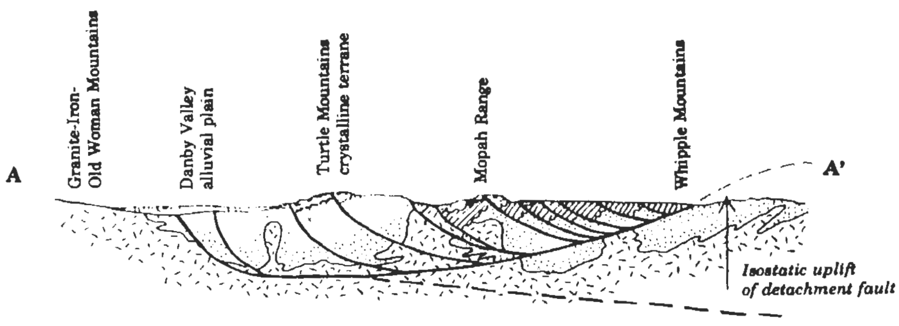

Figure 2. Generalized cross-section of Turtle Mountains (A-A' in Fig. 1). Adapted from Howard et al, 1982. Pattern of rock units same as in Fig. 1.

The earliest large-scale geological reconnaissance was done in the late 1950s by railroad company geologists (Cooksley, 1960a, b; Coonrad, 1960) and by R.B. Saul (1963), gathering information for the California State Geologic Map. Embree (1967) undertook the first truly detailed geologic mapping, returning to Chesterman's field area at the northern end of the range to describe the volcanic suite there.

In the early to mid-1970s, the southern extension of this volcanic terrane, a spur of the Turtle Mountains called the Mopah Range, was roughly mapped by geologists working for the Southern California Edison Company (Woodward McNiell and Associates, 1974). A nuclear power plant had been proposed for development in Vidal Valley, between the Mopah Range and the main bedrock massif of the Turtle Mountains. To assess the potential for future volcanic activity, the first radiometric dates for the region were collected from young plugs near Mopah Spring. A late Oligocene-early Miocene age was established for volcansim--too old to be of concern to the power company. However, lack of cheap water supply together with economic and political questions regarding nuclear power forced a decision against development. The wilderness remained wilderness.

In the late 1970s, Carr and others (1980) mapped in detail the southern tip of the Mopah Range. Over the next five years, most of the rest of the Turtle Mountains was intensively mapped and geologically explored as well (eg. Howard and others, 1982; Nielson and Turner, 1986; Allen, 1986; Hazlett, 1990). This work, funded by the U.S. Geological Survey, was done to evaluate mineral resources for the Bureau of Land Management, which under the Wilderness Act, was obliged to report on the economic potential of the area before its proposed Turtle Mountains Wilderness Area could be authorized.

As this work was being done in the Turtle Mountains, detailed studies were also continuing in the nearby Whipple and Chemehuevi mountains (John, 1982; Davis and others, 1980). A large-scale regional detachment structure, the Whipple Fault, had been identified. The lower plate of the fault is exposed by isostatic uplift in the central area of the Whipple Mountains. The upper plate consists mostly of Tertiary volcanic rocks noncomformably overlying detached basement, all extended by rotation along northeastward-dipping listric normal faults. The work in the Turtle Mountains made it clear the Whipple Fault continued westward beneath this range, which, in fact, could be construed as lying in the headwall terrane of the fault. Industry seismic profiling with CALCRUST data, and other geologic studies indicate that the Whipple Fault "daylights" beneath alluvium in Danby Valley, between the Turtle Mountains and the Granite-Iron-Old Woman mountains to the west (Okaya and Frost, 1989). The entire Turtle Mountains therefore is part of a detached allochthon of great regional extent (Fig. 2).

GEOLOGY

Turtle Mountains

From Highway 62, heading east toward Rice, the Turtle Mountains appear as a mass of high, very rugged hills, having a light-greyish coloration. A few hilltops are capped nonconformably by dark mafic basaltic and andesitic lava flows of Miocene age. This is the main body of the range—a Precambrian terrane made up of high-grade gneisses and other metamorphic rocks intruded by subophitic dikes of probable mid-Proterozoic age, middle to late-Proterozic granitic apophyses and veins, and Cretaceous granitic plutons. Resetting of fission track and K-Ar ages by multiple metamorphic and other heating events make it difficult to establish dates here. The oldest reliable apparent age, reported by Howard and others (1982), is 1.35 Ga. Foliations in gneisses strike generally east-west, matching foliations in similar rocks of mid-Proterozoic vintage throughout the region.

Preservation of quartzites, metaconglomerates, and amphibolites in some outcrops indicate the basement terrane in large part originated as siliciclastic sediments and basaltic rock. Ensuing multiple episodes of metamorphism and partial melting radically changed this crust over a period spanning at least a billion years. The abundance of migmatitic fabric in the terrane is an indication of the extreme level reached by metamorphism, and suggests that many intrusive bodies are locally derived. Paleodrainages preserved in the Miocene volcanic terrane show that the resulting crystalline rock had been uplifted and was shedding material similar to that making up the present terrane by early Miocene time. In general, the basement highland of the Turtle Mountains represents erosion to a mid-crustal level, much of this accomplished since the end of the Cretaceous.

Vidal Valley and the Mopah Range

Continuing beyond Rice, the wide prospect of Vidal Valley with the dark, rugged Mopah Range beyond comes into view. The fortress-like pinnacle rising from the center of Vidal Valley is Castle Rock. This is a mass of resistant ash-flow tuff overlying low angle slope-forming andesite flows. The darker hills in Vidal Valley lying closer to the road are also andesitic lavas interbedded with tuffs and tuffaceous epiclastic deposits. Towering plugs, some rising as much as 350 meters above the surrounding valley floors, give spectacular relief to the nearby Mopah Range. The two most prominent plugs, North and South Mopah Peaks, lie near Mopah Spring. These are hypabyssal dacite intrusions, which may have fed some of the thick flows making up the surrounding flat-topped ridges and mesas. Further east, on the approach to Vidal Junction, dark cuestas of subalkalic to alkalic basalt flows may be seen marking the eastern edge of the Mopah Range. These cuestas are fault blocks rotated gently westward due to transport along the underlying detachment fault. Their lavas are among the youngest erupted in the volcanic field.

Volcanic Activity

Volcanic activity in the Mopah Range-northern Turtle Mountains took place between 22.5 and 14.5 Ma (Hazlett, 1990). Detachment faulting occurred during much, if not all of this period of time, resulting in development of extensive angular unconformities within the volcanic section. Plugs and dikes are aligned parallel to and concentrated within the the most intensely faulted areas, showing an intimate connection between extensional faulting and magmatism. Composition of volcanics ranges unimodally from mantle-derived alkali-olivine basalts to rhyolites derived from mid-crustal fusion events. Andesites and dacites are the most abundant rock types. Two general pulses of activity occurred, as in volcanic areas throughout the southern Basin and Range: an early mafic-to-silicic period of activity, often explosive, preceeded a late-stage burst of effusive mafic activity. An interval of uncertain duration separated the two pulses of activity, marked structurally by a prominent unconformity. Preservation of vent structures indicates this volcanism took the form of eruptions from clustered cinder cones and domes, not unlike what is presently seen in the much younger Coso Hills volcanic field near China Lake.

The faulted volcanic terrane making up the Mopahs continues across Highway 95, north of Vidal Junction, into the western Whipple Mountains. Here volcanic strata dip more steeply westward, showing ever greater amounts of displacement and rotation as one moves eastward across the detachment terrane. Steepening of strata is in part an artifact of the domal uplift in the Whipple Mountains core. While the rocks making up the Turtle and Whipple mountains are quite different, both mountain systems are structurally linked in a manner important to understanding the regional geology.

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Mid-Miocene Sedimentation Patterns and Landform Development in the Whipple Mountains, Southeastern California

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INTRODUCTION

The area along both sides of the Colorado River between Lake Havasu City and Parker, Arizona, known as the Colorado River Extensional Corridor (CREC) (Fig. 1), is beautiful, displaying dramatic landforms and strong color contrasts between different rock types. Studies of sedimentary strata in this region indicate that the present-day topography developed as a result of early to middle Miocene detachment faulting, and has experienced little change since its formation. The landform history is particularly well exposed in the Whipple Mountains on the California side of the river.

The Whipple Mountains are a classic example of Cordilleran metamorphic core complexes. These features, the result of extreme Tertiary extension, are irregularly exposed in a belt from southern British Columbia to northern Mexico. More than 20 geographically separate "core complexes" have been recognized in this belt (Coney, 1980) in areas that experienced large amounts of Cenozoic crustal extension. Core complexes are characterized by: (1) domal or antiformal mountain ranges; (2) flanking low-angle normal faults (detachment faults of sub-regional to regional extent); (3) lower plate ("core") assemblages of crystalline rocks, commonly including mylonitic gneisses that formed at a depth of approximately 12-15 km; and (4) upper plate rocks that are highly distended by closely-spaced normal faults (Davis and Lister, 1988). The Whipple detachment fault has an overall northeastward dip, and extension occurred as the lower plate moved up and out to the southwest, away from the Colorado Plateau.

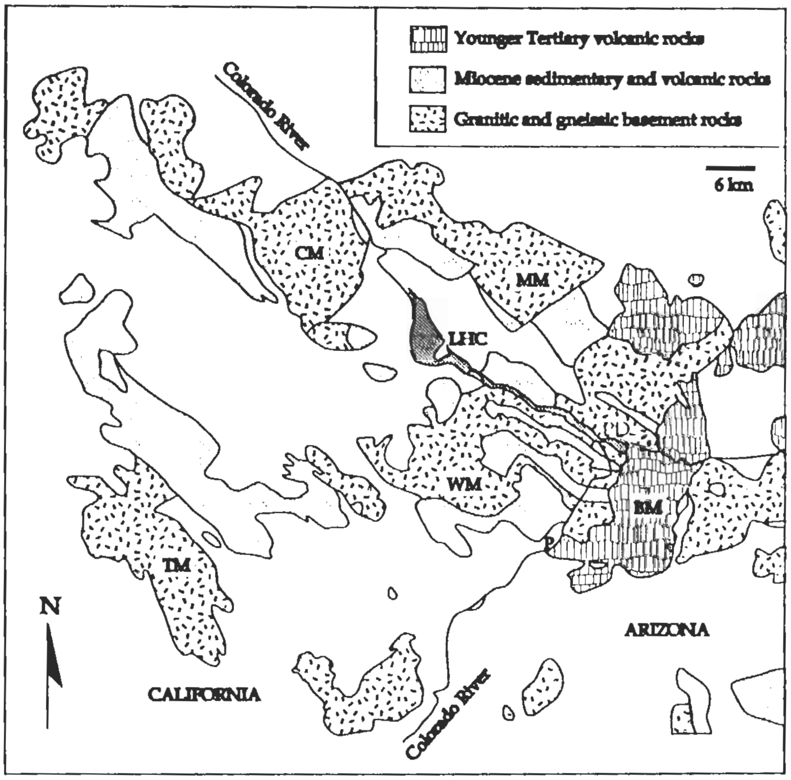

Figure 1. Schematic geologic map of the Colorado River Extensional Corridor, showing the distribution of Tertiary sedimentary and volcanic rocks. BM = Buckskin Mountains; CM = Chemehuevi Mountains; LHC = Lake Havasu City; MM = Mohave Mountains; P = Parker; PD = Parker Dam; TM = Turtle Mountains; WM = Whipple Mountains.

Nielson and Beratan (1990) and Beratan (1991) suggested that sedimentation patterns during extension in the CREC were controlled by two orthogonal fault sets. Small (less than about 15 km long) sedimentary basins formed as a result of tilting on high-angle normal faults oriented perpendicular to the NE-SW extension direction. Differential motion on adjacent tilt blocks was accommodated along transfer faults, high-angle fault zones oriented parallel to the extension direction. These two sets of faults controlled the location of basin boundaries, and thus the size and shape of the basins. Fault activity was episódic. The period of most active faulting probably occurred early in the history of the basin, followed by a time of relative quiescence and a final episode of block tilting. Uplift of the core of the Whipple Mountains, probably as a result of isostatic readjustment in response to tectonic denudation, affected strata deposited during the quiescent period.

Strata deposited within one of these syn-extension basins are now exposed in the southern and eastern Whipple Mountains (Fig. 2). These rocks are dominated by braided stream and sheetflood deposits. In the eastern part of the area, these rocks are associated with monolithologic breccia beds deposited by rock avalanches and mass flows. Lacustrine limestone forms an extensive unit in the western part of the basin. Volcanic flows and pyroclastic deposits are scant. These exposures represent the best and most complete sedimentary sections in the entire CREC; strata in other areas have been more dismembered by faulting and disrupted by coeval volcanic intrusions.

A more complete description of the stratigraphic units and the basin development history is given in Beratan (1990).

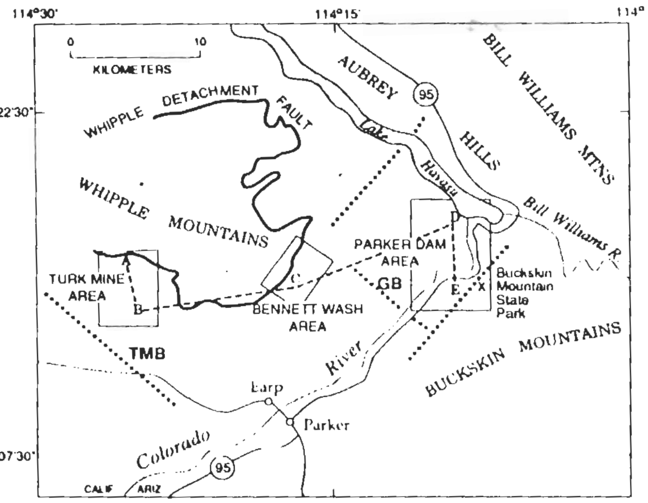

Figure 2. Map showing the location of some geologic and geographic features named in the text. **Location of Whipple detachment fault from Davis and others (1980). Heavy dotted lines indicate the approximate position of the basin boundaries (FB = inferred position of the high-angle normal fault bounding the basin during Twin Lode Mine—Copper Basin time). Stratigraphic sections**

STRATIGRAPHIC NOMENCLATURE

Syndetachment strata within the study area can be divided into two distinct depositional sequences, separated in the eastern part of the area by an angular unconformity. The two sequences are distinguishable by differences in outcrop extent and location, facies distribution, and clast composition.

Ransome (1931, 1933) described and informally named two Tertiary units—the older Gene Canyon Formation and the younger Copper Basin Formation—on the basis on a prominent angular unconformity within exposures near Parker Dam. Beratan (1990, 1991) redefined these formations on the basis of (1) stratigraphic position relative to the Peach Springs Tuff of Young and Brennan (1974), (2) presence or absence of Peach Springs Tuff clasts in conglomerate beds where outcrops of the tuff are lacking, (3) presence or absence of lava flows and intrusive rocks, (4) stratigraphic position relative to angular unconformities, (5) dominance of limestone compared to siliciclastic rocks, and (6) paleocurrent directions relative to basin depocenters.

Using the above criteria, Beratan (1990, 1991) also introduced two new units, the Turk Mine and Twin Lode Mine formations, which are equivalent to parts of the Gene Canyon Formation. An informally named "basal conglomerate" also was described. The defining characteristics of these stratigraphic units are summarized in Table I.

STRATIGRAPHY

Gene Canyon Formation

The Gene Canyon Formation is found only in the eastern part of the study area (Fig. 3). This unit is subdivided into three members, which from oldest to youngest are the Giers Wash Member, the Desilt Wash Member, and the Gene Wash Member.

The Giers Wash Member is characterized by dramatic lateral and vertical facies changes, lack of volcanic rocks either as primary deposits or as clasts, and by the presence of monolithologic breccia beds that have been interpreted as rock avalanche deposits. In the Parker Dam section (Fig. 3), the unit is dominated by coarse-grained sandstone and pebbly sandstone interpreted as streamflood deposits with subordinate matrix-supported mass flow deposits and small lenses of limestone. Cobble- to small boulder-bearing, sand matrix-supported strata, interpreted as mass flow breccias, dominate the Buckskin Mountain State Park section (Fig. 3), with subordinate stratified sandstones and conglomerates interpreted as streamflood deposits. Clast types include Proterozoic(?) and Cretaceous (?) granitoids and gneisses, and metasedimentary rocks including quartzite, phyllite, and white marble.

Table I. Summary of the defining characteristics of the stratigraphic units used in this study.

UNIT NAME	REFERENCE SECTIONS	DISTINGUISHING CHARACTERISTICS	BOUNDARY DEFINITION
Copper Basin Formation	Parker Dam, Turk Mine, Bennett Wash	Clastic sediment deposited after emplacement of the Peach Springs Tuff; heterogeneous clast assemblage, including Peach Springs Tuff; lacks interbedded volcanic flows	Angular unconformity Angular unconformity or first clasts of Peach Springs Tuff
Gene Canyon Formation	Parker Dam, Buckskin Mtn. State Park	Deposited before and during emplacement of the Peach Springs Tuff; deposits include a well-defined basin center and basin margin	First volcanic flow
Gene Wash Member		Volcanic flows and tuff with interbedded sandstone and conglomerate	Color change, facies change
Desilt Wash Member		Lack of volcanic flows; presence of minor quantities of Tertiary volcanic rock clasts; lack of monolithologic breccia	
Giers Wash Member		Lack of volcanic rocks either as flows or clasts; presence of monolithologic breccia; lateral and vertical variability in texture and grain size	Nonconformity on older granitic suite
Twin Lode Mine Formation	Turk Mine, North Turk Mine	Lacustrine limestone; lack of volcanic flows	Change from limestone to clastic deposition
Turk Mine Formation	Turk Mine, North Turk Mine	Dominated by volcanic flows of mafic to intermediate composition; sedimentary interbeds lack clasts of Peach Springs Tuff	Base of first limestone bed
basal conglomerate	Turk Mine	Coarse-grained clastic deposits derived from nearby sources; lack of Peach Springs Tuff or limestone clasts; deposits lack a recognizable basin margin or basin center	Nonconformity on older granitic suite

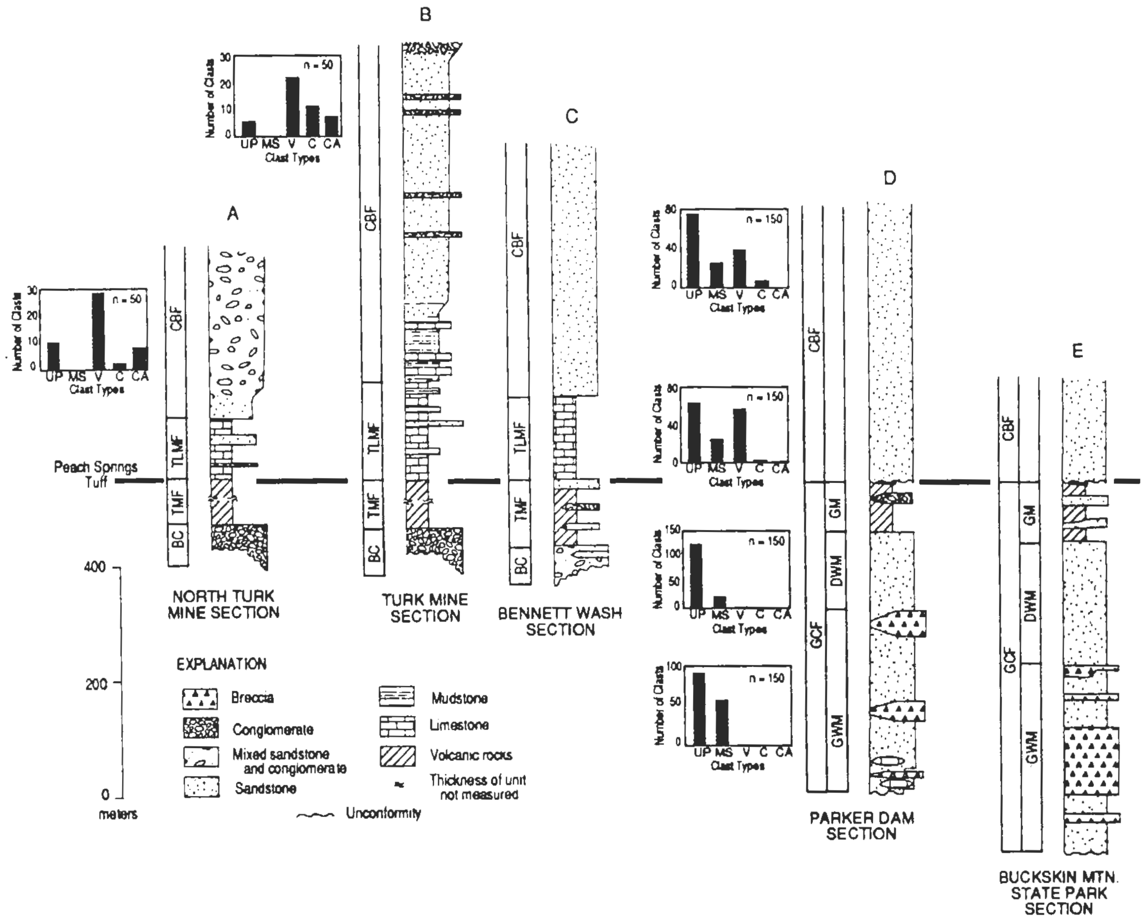

Figure 3. Simplified stratigraphic columns, showing the distribution of stratigraphic units in the southern and eastern Whipple Mountains. (GCF = Gene Canyon Formation; GWM = Giers Wash Member; DWM = Desilt Wash Member; GM = Gene Wash Member; BC = Basal Conglomerate; TMF = Turk Mine Formation; TLMF = Twin Lode Mine Formation; CBF = Copper Basin Formation). The histographs show typical clast compositions of the portions of the columns on which they are centered (UP = upper plate crystalline assemblage; MS = metasedimentary and metavolcanic rocks; V = Tertiary volcanic rocks; C = Tertiary clastic rocks; CA = Tertiary carbonate rocks and chert).

The Desilt Wash Member displays considerably less facies variation than does the Giers Wash Member. In the Parker Dam section, the unit comprises moderate reddish brown, moderately well-indurated, fine pebble-bearing coarse-grained sandstone, with subordinate interbedded medium and mixed medium-grained to very coarse-grained sandstone and pebble to cobble conglomerate. These strata are interpreted as streamflood deposits. In contrast, laterally persistent beds of thinly bedded fine-grained sandstone and siltstone, interpreted as playa deposits, and interbedded streamflood deposits dominate in the Buckskin Mountain State Park section (Fig. 3). Clasts within the Desilt Wash Member were derived primarily from Proterozoic(?) and Cretaceous(?) granitoid and gneiss and also include rare clasts are less common in this unit than in the Giers Wash Member.

The Gene Wash Member is distinguished from underlying units by the presence of thin, strongly altered, mafic to intermediate lava flows with interbedded coarse-grained sedimentary strata. A rhyolitic ash-flow tuff that tentatively has been identified as the Peach Springs Tuff caps the unit (Beratan, 1990). The lava flows are sparsely porphyritic, containing about three to five percent small oxidized pyroxene phenocrysts; plagioclase phenocrysts are rare to absent. Interbedded sedimentary strata consist of conglomerate and medium- and coarse-grained sandstone, similar to the underlying Desilt Wash Member with additional prominent boulder and cobble conglomerate beds. The sedimentary strata are generally coarser in the Parker Dam section than in the Buckskin Mountain State Park section.

Basal Conglomerate

Strata that form the base of the Turk Mine, North Turk Mine, and Bennett Wash sections in the western half of the study area (Fig. 3) are texturally variable and include poorly sorted, unorganized, clast- and matrix-supported breccia (rock avalanche deposits) and pebbly very coarse-grained sandstone, interpreted as streamflood deposits. The clasts in any given exposure commonly are limited in variety and are similar in composition to local basement rocks. The basal conglomerate varies in thickness, indicating topographic relief on the original depositional surface.

Turk Mine Formation

The Turk Mine Formation is found in the western half of the study area (Fig. 3). It conformably overlies the basal conglomerate and consists of medium grey lava flows of mafic to intermediate composition. The flows have been strongly altered by potassium metasomatism. The formation varies from about 100 to 500 m thick, and sedimentary interbeds are uncommon.

Twin Lode Mine Formation

The volcanic flows of the Turk Mine Formation are overlain conformably by as much as 200 m of lacustrine limestone (Fig. 3) predominantly composed of calcic micrite, which contains abundant algal material. The siliciclastic content of the limestone is variable but generally high. Rare conglomerate interbeds are dominated by pebble-sized mudstone and carbonate intraclasts. Secondary silica is a common and characteristic component of the Twin Lode Mine Formation, varying from incipient replacement of calcite by microcrystalline quartz and chalcedony to nodules and massive layers of chert.

Copper Basin Formation

The Copper Basin Formation is found throughout the study area. In the Parker Dam area, the Copper Basin Formation unconformably overlies the Gene Canyon Formation with 15-20 degrees of angular discordance. The unit is dominated by moderately sorted, medium-grained to very coarse-grained sandstone. The beds are laterally impersistent. Crude stratification and large-scale, low-angle cross-bedding are the most common sedimentary structures. These sandstone and conglomerate beds are interpreted as streamflood deposits. A large variety of clast types (Fig. 3) is present, including granitic gneiss and coarsely crystalline granite derived from the upper plate of the Whipple detachment fault, Tertiary volcanic rocks including the Peach Springs Tuff, Tertiary limestone, and quartzite. Tertiary volcanic rocks are the most abundant clast type near the base of the Copper Basin Formation, with clasts from the upper plate crystalline assemblage becoming more abundant higher in the unit.

In the western half of the basin, the Copper Basin Formation conformably overlies the Twin Lode Mine Formation. The unit is dominated by sandstone, ranging from well-bedded, fine-grained sandstone and siltstone interpreted as interbedded turbidite and resedimented deposits (Walker and Mutti, 1973) to well-sorted medium sandstone with rippled surfaces. In general, these deposits exhibit an overall coarsening-upward sequence. Locally, interbedded sandstones and conglomerates interpreted as streamflood deposits overlie the Twin Lode Mine Formation.

Osborne Wash Formation and Bouse Formation

The tilted, syn-extension strata in the eastern and southern Whipple Mountains are unconformably overlain by alkali-olivine basalts and interbedded fanglomerates and streamflood deposits belonging to the Osborne Wash Formation, and siltstones and marl of the Bouse Formation. These units are discussed in Buising (this volume).

TIME OF DEPOSITION

Ages of strata within the study area are poorly constrained due to the scarcity of datable material within the basin. A fission-track age of 23 ± 2.4 Ma (zircon) was obtained from a lacustrine tuff located near the base of the Tertiary section in the Aubrey Hills, just north of the study area (Beratan, 1990; Nielson and Beratan, 1990); this is the best estimate of the age of basin inception. The Peach Springs Tuff provides an upper age limit for the Gene Canyon Formation and a lower age limit for the Copper Basin Formation. The accepted age of this widespread ignimbrite is 18.5 ± 0.5 Ma ($^{40}Ar/^{39}Ar$, sanidine; Nielson and others, 1990). A sequence of basalt lava flows in the Aubrey Hills thought to be slightly younger than the Copper Basin Formation has yielded a K-Ar age of 14.1 ± 0.3 Ma (whole rock; Nielson and Beratan, 1990).

LANDFORM DEVELOPMENT

The landforms seen today in the Whipple Mountains resulted from the Miocene extensional event. The topography can be divided into five dominant landscape elements.

(1) The bulk of the range consists of a domiform topographic high, primarily composed of mylonitic gneisses in the lower plate. Synextension sedimentation patterns indicate that the topographic expression of the range core developed between approximately 18.5 and 13 Ma. This uplift is thought to have resulted from localized isostatic uplift due to non-uniform tectonic denudation (Spencer, 1982).

(2) Prominent northwest-trending ridges flank the range, held up by resistant Early to Middle Miocene andesite flows (Turk Mine Formation) and strongly cemented sandstones and conglomerates (Copper Basin Formation). The strata commonly are tilted to about 30 degrees to the southwest, and the southwest flanks of the ridges generally form dip slopes. Cliffs commonly occur on the northeast flanks of the ridges. Tilting along northwest-trending high-angle normal faults to form half-graben basins occurred three times during the detachment event; the ridges observed today resulted from the final tilting episode at about 14 Ma (Nielson and Beratan, 1990; Beratan, 1991).

(3) The area between the ridges consists of low, irregular hills separated by poorly interconnected washes. These rubbly hills formed on the highly fractured upper-plate granites and gneisses.

(4) A plateau capped by nearly horizontal late Tertiary basalt flows covers the tilted ridges along the east side of the Colorado River. This plateau forms the bulk of the Buckskin Mountains. The basaltic volcanism probably represents the gradual shutoff of detachment faulting (Buising and Beratan, in press).

(5) The Whipple Mountains are flanked by a broad alluvial apron deposited after the final tilting event, and low-lying regions between the other landscape elements contain nearly flat-lying fluvial deposits. These deposits range in age from late Miocene to present. Also included in this landscape element are the Bouse Formation (Buising, 1991) and younger Colorado River gravels.

Facies analysis of synextensional sedimentary strata and contact relations with younger units suggests that relatively little landscape modification has occurred since the end of detachment faulting. Although broken up and distended by the last phase of extensional faulting, most of the Middle Miocene depositional basin was preserved, including the margins and the depocenter. Erosion has reduced the size of most exposures; however, large blocks of Tertiary strata do not appear to be missing. Abundant silica cement in these rocks may account for this preservation. The only major landscape modification observed is some incisement of drainages in response to the introduction of the Colorado River to its present course (Buising, 1991).

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The Bouse Formation and Bracketing Units, Southeastern California and Western Arizona

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Mio-Pliocene sediments of the lower Colorado River area represent the northernmost well documented extent of the proto-Gulf of California, a tectonically enigmatic marine incursion that occupied much of what is now the Gulf of California region–including the lower Colorado River area and the Salton Trough–as much as 8 m.y. prior to the onset of spreading- and transform-related subsidence in that area. Fanglomerate and volcanic rocks informally referred to as the Osborne Wash strata interfinger with the conformably overlying Bouse Formation. The Bouse Formation includes carbonate and coarse terrigenous-clastic basin margin deposits that interfinger laterally with basin fill material; basin fill strata comprise a basal estuarine carbonate unit overlain by fine-grained terrigenous-clastic deltaic deposits. The upper portion of the Bouse interfingers with overlying cobble conglomerates referred to as the Colorado River gravels. Syndepositional folding of the Bouse Formation and bracketing units is believed to reflect slumping on oversteepened slopes, perhaps exacerbated by episodic tectonic activity. Syndepositional faults show both normal and reverse separation; outcrop relations allow but do not prove a strike-slip component of motion on some structures. Contemporaneous minor faults seem to reflect mutually incompatible stress orientations; this suggests that they record either localized stress fields or localized anomalous responses to regional stresses. Alternatively, they may reflect extension in two directions in the sediment pile overlying the subsiding basin floor. The Bouse Formation and bracketing units record four stages in the evolution of the northern proto-Gulf/lower Colorado River area: (1) dissection of preexisting, detachment fault-controlled topography and localized, interior-drainage alluvial deposition (about 14-9 Ma); (2) regional subsidence and proto-Gulf transgression (perhaps as early as about 8 Ma; not later than about 5.5 Ma); (3) progradation of ancestral Colorado River delta into the northern end of the proto-Gulf basin (prior to about 4.3 Ma); and (4) arrival of throughgoing Colorado fluvial channel (prior to about 3.5-4 Ma). Outcrop relations between the Osborne Wash strata and the Bouse Formation, and the relationship of these units to modern landforms, indicate that topography in the lower Colorado River area has not changed significantly since middle Miocene time. Subsidence of the Bouse basin is believed to have occurred via broad regional downwarping, which may have represented a sag formed as the locus of active proto-Gulf extension propagated northward into the lower Colorado River area from the block-faulted southern portion of the proto-Gulf. Timing suggests that incipient proto-Gulf extension in the lower Colorado River area was arrested by the approximately 5-Ma shift to the modern transtensional regime.

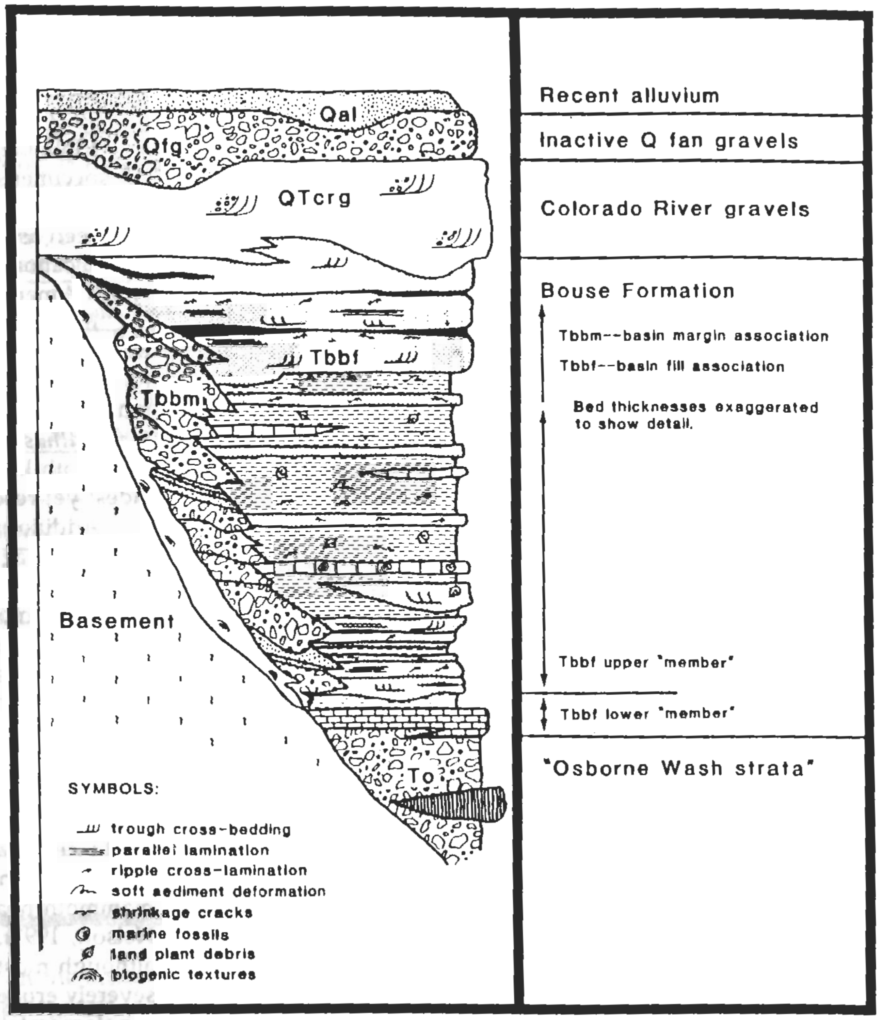

Figure 1. Generalized regional stratigraphy, post-mid-Miocene units, lower Colorado River area. Scale variable; exposed thicknesses differ.

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Mammoths in the Colorado River Corridor

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INTRODUCTION

In 1970, Saunders (1970) summarized the known distribution of mammoths in Arizona. His data provided for four mammoth localities between Yuma, Arizona and Lake Mead. Little additional information was obtained in the succeeding twenty year interval, and his summary did not address the neighboring states of Nevada, California, and Sonora, Mexico.

The Bureau of Land Management, Lake Havasu District, and Kingman District, contacted Northern Arizona University in 1990 with regard to a mammoth located near Golden Shores, Arizona. During excavation of that specimen, contact was made with persons in the region who were knowledgeable about other occurrences, and with the San Bernardino County Museum. The results of this interaction, plus published data, have more than tripled the known and reported occurrences of mammoths along the Colorado River Corridor (Fig. 1).

INVENTORY OF REPORTED SPECIMENS

Saunders' data (1970) located four mammoth sites in Yuma and Mohave counties. These sites were designated:

- WPB1—Yuma Mammoth (Blake, 1990; Hay, 1927)

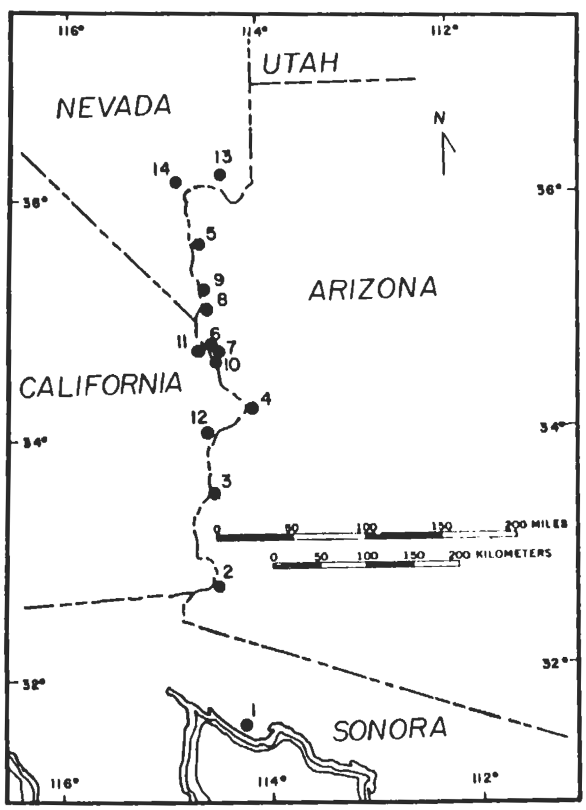

Figure 1. Recorded mammoth localities, Colorado River Trough. Numbers indicate descending sequence in Table I.

- USGS:LCRP 4-23-2—Ehrenberg Tusk Locality (Metzger, personal communication 1964; this locality was published at a later date in Metzger and others, 1973 and Bell and others, 1978)
- JSN1 Bill Williams Fork (Newberry, *in* Ives, 1861; Hay, 1927)
- JSN2 Elephant Hill (Newberry, *in* Ives, 1861; Hay, 1927).

All specimens were designated *Mammuthus columbi* by Saunders (1970); one specimen reported by Newberry (1961) had been assigned to *Mammuthus primigenius*, probably an identification error in light of our present knowledge of the North American distribution of this species.

In 1974, a mammoth represented by the right half of the mandible was discovered on the Arizona side of the Colorado River (McShan, 1974) and was stored at the Needles, California museum. Measurements in 1991 indicate that this specimen is *Mammuthus meridionalis*. This species is the oldest form of mammoth known in North America, and the specimen is the oldest yet recovered from Arizona.

Additional specimens housed at the Needles Museum include a fragmentary humerus from Needles, California (M. McShan, personal communication 1991) and a disintegrated mammoth molar from Topock, Arizona (M. McShan, personal communication 1991).

In 1971, a nearly complete cranium of a mammoth was discovered by two residents of the Colorado River Indian Tribes Reservation near Poston, Arizona, but on the California side of the Colorado River. Measurements in 1991 indicate that this specimen is also *Mammuthus meridionalis*.

During the winter of 1990-91, excavation teams from Northern Arizona University, in response to a request from the Bureau of Land Management, excavated the remains of a mammoth near Golden Shores, Arizona (Jackson, 1991; Nelson, 1991). A nearly complete skeleton was recovered, although most of the pelvic region and hind limbs had been severely eroded. Encased in fine clays overlying a conglomerate, the individual was preserved in a position indicative of settling on its back in shallow water. Abundant imprints of reeds and sedges suggest a locality similar to the marshes that currently exist north of Topock. A tooth had been removed from the specimen, so it was unavailable for exact species determination. An attempt to identify the species using a photograph of the tooth provides a questionable assignment as *Mammuthus meridionalis* or *M. imperator* (designation from measurements taken from a newspaper photograph).

During the excavations at Golden Shores, we were informed of mammoth remains curated at the Mohave County Historical Museum. Examination of those specimens indicated there were two individuals, from two localities "along the shore of the Colorado River," possibly in the Bullhead City, Arizona region. One specimen (MCM 468:67) also is *Mammuthus meridionalis*, based on dental measurements. (There is overlap in most dental criteria for some of the

Table I. Recorded Mammoth Localities for the Colorado River Trough

Location	Site Number	Reference	Description
El Golfo, Sonora	LACM 118451	Shaw, 1981	*Mammuthus imperator* M3 enamel plate fragments; L astragalus; R navicular; prox. L femur frag; frag. metatarsal; dist. metapodial
Yuma Co., AZ	WPB1	Blake, 1900 (Saunders #70, 1970)	Yuma, AZ. Maxilla w/ teeth, lower molar, tusk fragments. Valley fill along Colorado River, ±125' elev.
La Paz Co., AZ	USGS LCRP 4-34-3	(Saunders #47, 1970)	Ehrenberg Tusk Locality. 380' elev. Tusk in Chemehuevi Fm
Mohave Co., AZ	JSN-1	Newberry, *in* Ives, 1861; (Saunders #58, 1970)	Bill Williams Fork. 480' elev. Single tooth from valley fill along Colorado River in 1857
Mohave Co., AZ	JSN-2	Newberry, *in* Ives, 1861; (Saunders #59, 1970)	Elephant Hill. Single tooth at base of Elephant Hill, elev. 720'.
Mohave Co., AZ	SBCM 3.3.3	Needles Museum, 1974	*M. meridionalis* half mandible with tooth
Mohave Co., AZ	NAU-QSP 9155		Golden Shores Mammoth. *M. meridionalis*, nearly complete skeleton in lacustrine unit, evel. 645'
Mohave Co., AZ		Mohave County Museum	Femur fragment, Colorado River, Arizona
Mohave Co., AZ	MCM 468:67		Isolate tooth, Colorado River, Arizona
Mohave Co., AZ		Needles Museum	Topock, Arizona. Disintegrating molar.
San Bernardino Co, CA		Needles Museum	Needles, California. Partial humerus.
Riverside Co., CA	CRIT skull		Colorado River, California, west of Poston, AZ. Nearly complete skull, *M. meridionalis*
Clark Co., NV	NAU-QSP 9156		Lake Mead NRA, NV. Dentition, partial tusk, fragmentary ribs
Clark Co., NV	SBCM 2.6.21	Reynolds, 1991	North Las Vegas Mammoth. tusk, mandible with dentition, maxillary dentition.

species; the tooth was assigned to *M. meridionalis* on the basis of the length/laminar frequency). This specimen provides a second record of the species from Arizona, and three known examples from the Colorado River Corridor.

In the spring of 1991, Agenbroad and Mead were contacted by the National Park Service-Lake Mead National Recreation area with regard to a mammoth weathering out of sediments below the high water mark of Lake Mead. An earlier visit to this locality by representatives of the University of California, Berkeley (Mason, pers. comm. to Agenbroad, 1987) recovered maxillary teeth (V87021-131822). A cast of one tooth was analyzed during salvage excavation of the rest of the faunal remains indicate that the "Virgin" mammoth was *Mammuthus columbi*.

Table 1 summarizes the known and reported mammoth localities along the Colorado River, including specimens outside the river valley but relevant to the distribution of mammoth species in the region.

DEPOSITIONAL OCCURRENCE

The localities that provide data on the geologic depositional units containing the mammoth remains can be rather generally assigned to the Chemehuevi Formation (the Chemehuevi Gravels of Lee, 1908; Longwell, 1936). These deposits appear to be relatively continuous from Lake Mead to Imperial Dam, north of Yuma (Metzger and others, 1973). They have been attributed to two different depositional origins: lacustrine, and fluvial (Longwell, 1936). Longwell (1936, 1946, 1954, 1960, 1963, 1965) favors a ponding (lacustrine) environment of deposition; however, he can account for no suitable base level (natural dam). Metzger and others (1973) consider the deposits which are collectively called the Chemehuevi Formation (units D & E of Metzger and others, 1973) to be depositional units formed by the Colorado River, "during a time in which it was graded to the Gulf of California." Except for the minor base level changes produced by dam construction, it seems the Colorado River has *always* been graded to the Gulf of California!

Longwell (1936, 1965) gives one of the better descriptions of the Chemehuevi Formation, as a sedimentary unit consisting of clay, silt, and sand, "laid down in the old lake." He describes it as whitish clay and silt in sharp contact with the underlying Muddy Creek Formation in the Virgin and Muddy Valleys. Near Callville, Longwell (1936) describes the Chemehuevi Formation as consisting of an upper cream-colored member forming steep slopes, and a lower buff-yellow member standing in steep cliffs.

Bell and others (1978) describe the Chemehuevi Formation as "two dominant facies: 1) massive to thin-bedded pinkish silt and fine silty sand, with subordinate clay and sand, and 2) interbedded locally derived alluvial gravels. Basal and capping gravels described in previous work are not now recognized as part of the formation." They present a paleomagnetic analysis of 29 specimens as "strong monodirectional normal polarity." Their soil analysis data suggest that overlying caliches range from 30,000 to 90,000 BP. Johnson and Miller (1980) discuss the post-Chemehuevi depositional sequence.

Colorado River sediments near Parker, Arizona, which postdate the Bouse Formation, consist of tan, pink, and brown silts with interbedded well-rounded cobbles and gravel beds. The Colorado River Indian Tribes mammoth was found in these sediments, which may be equivalent in age to the Chemehuevi Formation at Needles and Blythe.

CHRONOLOGY

Few radiometric dates are available for the Chemehuevi Formation or its enclosed fossils. Bell and others (1978) cite a uranium-thorium date on the Ehrenberg tusk from the Chemehuevi Formation (unit D of Metzger and others, 1973) as $102{,}000 \pm 700$ BP. To our knowledge, no other proboscidean material from the Colorado River Corridor has been subjected to a radiometric age analysis.

Longwell (1936) cites a bison horn core in an upper sandy unit near Callville, indicative of a Rancholabrean Land Mammal Age for an upper unit of the Chemehuevi Formation, which is consistent with the presence of *Mammuthus columbi* at other locations. The presence of *M. meridionalis* in three localities, and possibly a fourth, suggests an Irvingtonian Land Mammal Age for the early part of the formation.

Based on the sparse information outlined above, the Chemehuevi Formation and its included fossils range from Irvingtonian through Rancholabrean LMA. The only absolute date is 102,000 B.P. for the proboscidean tusk at Ehrenberg, Arizona. Bell and others (1978) conclude the Chemehuevi Formation is pre-Wisconsin (100,000-200,000 BP). However, the presence of *M. meridionalis* suggests that deposition of the lower portion of the Chemehuevi Formation started in Irvingtonian LMA times, at least as early as 300,000 to 500,000 ybp (Repenning, 1987; Woodburne, 1987).

SUMMARY

Since Saunders' (1970) mammoth census for Arizona, the number of known localities along the Colorado River from Sonora to Lake Mead has tripled. All southern species for North America are represented: at least three *M. meridionalis* specimens are known from California and Arizona portions of the corridor; one *M. imperator* is known, from El Golfo, Sonora; several *M. columbi* specimens are known; and there are non-diagnostic specimens from several localities. It is anticipated that additional mammoth localities will be discovered (and, we hope, reported) so the faunal record and inferred paleoenvironment of this region can be reconstructed with increased accuracy.

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