OLD ROUTES TO THE COLORADO

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

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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 Peninsular 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 Yuccaipa and Palm Springs. Ahead you can see Yucca Ridge and San Bernardino Peak (elevation 10,525 ft). Yucca 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 Yucca Ridge south of the mountain front. The Mill Creek Formation is a series of nonmarine Tertiary sediments deposited in a pull-apart basin (Demiter, 1986; Sadler and Demiter, 1986). The formation contains fossils which suggest an early Pleistocene age (Asielrod, p.c. to Reynolds, 1985) but which may be as old as late Miocene or as young as middle Pliocene (Galston, 1971). The Wilson Creek Fault (Matt and others, 1983 and 1985) crosses Yucca 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, 1988). The maif is glaciated during the Pleistocene (Sharp and others, 1959; Debbal, 1964).

2.3 (1.9) Holocene alluvium is on both sides of the freeway; to the right at about 2.00, Sanley Heights is on a Pleistocene alluvial surface with a Pleistocene soil (Reynolds and Redder, 1986). This Pleistocene surface will be encountered repeatedly throughout the Yuccaipa/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 rancho, California State Historical Landmark 43. The first irrigation project in the county, the rancho was constructed in 1819 and 1820 by Serrano and Cahulla Indians under the guidance of Franciscan fathers from the Mission San Gabriel to develop agriculture at Guachuma, the Indian ranchería 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 (Qua 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 (4.8) Reserve Canyon. Cross the trace of the Craton Fault (Reservoir Canyon Fault) offsetting Quaternary alluvium and uplifting Precambrian metamorphic and igneous basement rocks (Rogers, 1967). The Craton Hills are a faulted complex of upper and lower plate rocks divided by the Vincent Thrust. Oceanus Decatur Gass held gold-bearing quartz veins on the Yuccaipa side of these hills in 1884. By 1889 the "Gold Bar Company" had developed a 60-foot tunnel and in 1890 the water-powered Yuccaipa Quartz Mill had been constructed to process gold ore. The mine property was located in the canyon north of Craton 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, those exposures of the upper plate gneisses include Permo-Triassic Lowe Granodiorite and cataclasites. Pelona Schist is exposed in the road cuts to the right near the top of Reserve Canyon.

6.8 (0.4) Marked by trees and bushes to the right, Crystal Springs come to the surface at the fault trace. These springs supported a small bottled water industry in the past. Reserve 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's pioneering flaring 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 presently-day Crafton (Beadle 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 granitic 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; Pickag Peak (elevation 5,487 feet) is at 10:30. Pickag Peak is south of the south branch of the San Andreas Fault and consists of upper plate granite and granitoid gneiss 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. 1996).**

11.2 (1.0) **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 Becter, 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 130 years.**

14.2 (1.7) **Terraces to the right are developed on Pleistocene and Holocene alluvium. The badland topography is developed in the Plio-Pleistocene San Timoteo Formation (Reynolds and Becter, 1986; Reynolds and Becter, 1986, 1991). At 10:00 the terrace 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 Raymond Flat area on the skyline to the left (Mato and others, 1983).**

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-10 West is on the right.**

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

19.7 (0.6) **San Gorgonio Pass is the lowest topographic break in southern California through the mountains to the inland desert, separating Mt. San Gorgonio (11,592 feet) and Mt. San Jacinto (10,786 feet), 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 migmatic gneiss and green schist (Pebina 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. Bree remnants have been recovered from the Heights Fanglomerate (Jefferson, 1986) indicating that it is less than 500,000 years (Savage and Russell, 1983). The Heights Fanglomerate unconformably overlies sediments similar in appearance to the San Timoteo Formation, which underlies 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 (Bortognino and Spittler, 1986; Dibblic, 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, Pleocene in age, are cut up between thrust faults along the base of the mountain front from this point to Stubbe Canyon (Allen, 1957; Dibblic, 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 larger gneiss derived from an area north of the Banning Fault between Cottonwood and San Gorgonio canyons. He also mentioned rare clasts of stilplicid 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, 1980).

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 (0.4) 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 Palisade metasediments (quartzofeldspathic gneiss and schist, phyllite, quartzite, and marble) intruded by quartz diorite of Mt. San Jacinto (Morton and others, 1980).

To the left is Millard Canyon; a fault scarp crosses the alluvial fan near the canyon mouth. The debris of the Millard Canyon fan overwheets 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 overlain 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 "bulldozer" 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 10:50 at Stubbe Canyon is seen where pink piemonte-bearing rocks are thrust over green epidote-bearing rocks. The distinctive piemonte-bearing gneiss is found as clasts in sediments on the south side 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 (shaded at 12:30 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 copped by a Pleistocene soil.

36.0 (0.8) Verbonia 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 9:30. 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, 1948). Proctor suggests that the Garnet Hill Fault may be an ancient branch of the San Andreas Fault.

37.6 (0.6) South of the interstate, large cottonwood trees and scarp 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 Ranch was granted in 1845 and encompassed
the entire pass area. Weaver sold a portion of the ranch 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 1853 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 scar 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 Cabazon Fan lobate separated from the Imperial and Painted Hill formations (Murphy, 1986) by the Banning Fault. The Cabazon Fan lobate of Whitewater Hill includes a lens of limestone breccia believed to have been derived from the San Jacinto block (Allin, 1987). 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 Yuca 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 shatter ridges between the powerline and windmills. Devens 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 Bldg. Mt. San Gorgonio is viewed to the left at 10:00; to the right at 2:00 are the Little San Bernadino 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 Mezquita deformed pluton and Pecanbrian 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 left lateral faults. These faults appear to 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 migratory 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 end eroded Morongo Valley Fault east of Big Morongo Canyon. The terrace at the dary 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 granite 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.
57.7 (0.7) Pass Hseopa Road.

58.0 (0.3) Morongo Valley Park. The contact above the shooting range exhibits grey granite 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 Pinto 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 basaltic ultramafics (serpentinite). This fanglomerate overlies and is in fault contact with the quartzite fanglomerate which comprises 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 east. Look ahead for the flashing yellow traffic lights, where we will be turning left.

61.1 (1.6) TURN LEFT on Pioneertown Road, just past flashing yellow pedestrian crossing lights. To the right, this road is called "Door 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 northwest suspected trace of the Pinto Mountain Fault as mapped by Dibblee (1967a).

64.5 (1.8) Water Canyon Fault. Notice that the fluvial 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 km² and may reach a thickness of 60 m. The pile is made up of individual flow units three to seven in thickness, 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 Koozer, 1986; Peterson, 1976; Oberlander, 1972). The basalts overlie and are interbedded with Tertiary arkose deposits and overlie granite 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 alkali volcanics found throughout the Mojave Desert; see Cima Dome, Amboy Crater, Dash Hill, and Pيgah Crater (Neville, 1983; Neville and others, 1985, and see Reynolds, this volume; Lawton, this volume; and Harlett, 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 (Gadde, 1974).

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

67.0 (0.3) Cross Chaparras Wash.
67.3 (0.5) 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 lithiogeochemistry 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 suggest they are time correlative with the upper portion of the Crowder Formation in Cajon Pass (Reynolds, this volume).

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

67.3 (0.3) TURN RIGHT onto Pioneertown Road and resume route northwest. The basalts appear to have flowed over a gendy undulating surface and are thickest to the southeast. 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 mesa 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 columns joining 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 2:30 indicates that the debris may have been stable since middle Pleistocene times.

76.3 (1.9) Quaternary stream deposits of Pipes Wash are exposed to the left and in road cuts. Pipes Wash and Chuparrosa 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 Chuparrel 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 Bobo Springs cuts through Pleistocene sediments.

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

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

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

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

Figure 2. Stop 2. Pleistocene sediments capped by a dense layer of calcium carbonate along Liam Road. R.E. Reynolds photo.
84.6 (1.3) 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) and 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 (1.3) STOP 3. HOMESTEAD VALLEY FAULT. Park off pavement. Linn Road cuts through Pleistocene sediments: red-brown gravel, grey-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 Umberger, 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 (1.3) TURN RIGHT onto Beltfield 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. Cyprian Ridge, the low ridge at 9:30 southeast of Hidalgo Mountain, consists of Pleistocene lacustrine sediments deformed along the West Calico Fault (Dibblee, 1967; Knauss, 1982).

96.0 (3.6) TURN RIGHT (south) on Border Avenue.

Cross the trace of the Emerton Fault near its intersection with Copper Mountain Fault, which runs southeast at 10:30. We are driving over calcified Pleistocene fans, and we have left the Emerton Lake drainage system and have entered the drainage which runs northeast to Deadman Lake and Boulion Wash. Surprise Spring, which is the locality of a Pleistocene dune 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 southeast through the mountain. The view also includes eroded granite 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 (0.4) TURN LEFT onto Pole Line Road. We are traveling parallel to the trace of the Pinto Mountain Fault, Copper Basin playas, to the north (left), receives water from Yuca 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) Aridic 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 at Twentynine Palms; the next reliable gas stop is not until Parker.

114.3 (3.4) Indian Cove Road. Continue on Highway 62.
114.6 (0.3) To the north at 10:00 a.m. a thick section of Pleistocene sediments is visible 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 (0.5) Lancer/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 (0.9) 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 Bouillion 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 Adree 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 (Corbett 1991).

121.2 (>0.0) 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 calcified sediments below desert pavement. Campbell Hill is to the northeast at 2000.

123.3 (0.8) Stop sign at Two Mile Road.

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

128.6 (0.5) Green lacustrine sediments at Copper Basin. R.E. Reynolds photo.

128.7 (0.6) Arkoisic 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 archeological work in this area (Campbell and Campbell, 1939).

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

124.5 (0.1) TURN SHARP RIGHT (jubly) 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 Mosquite Lake Fault.

124.8 (0.1) STOP 5. CAMPBELL HILL. Late Pleistocene sediments at Campbell Hill (Fig. 5) have been uplifted along the north-east side of the Mosquite Lake Fault (Dobler, 1968; Jagello and others, 1992; and see Foster, this volume). Vertebreate remains are typical of the Rancholabrean Land Mammal Age and include ground sloths, dwarf pronghorn, saber-tooth, mammoth, horse, and camel (Jefferson, this volume). Bacheiller (1975) tentatively identified the Bishop Tuff in the section, which is dated at 0.73 Ma.

The Boliom Mountains are due north; Hidalgo Mountain is N30°W, and Gypsum Ridge the low ridge to the right. The Boliom 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 Mosquite Lake is about 1.5 miles further to the north. Before Dale Lake received water from the Mosquite Lake basin, the basin would have been filled to an elevation of 1800'.

127.0 (1.4) We are between the Mosquite Lake Fault and the north branch of the Mosquite Lake Fault. The drainage one mile north of the road leads to Dale Lake.

130.4 (1.4) Enter calcified 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 Mosquite Lake Fault.

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

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

133.6 (1.4) 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' water 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 1.5, below Clarks Pass (see Conolly, 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 Yuca Valley (3400'), Copper Basin (2400'), and Mosquite Lake (1750') during Pleistocene times. Schrock (this volume) reviews the prehistory of this area; Tchakian (this volume) discusses the dune field at Clarks Pass.

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.

147.8 (1.7) A northwest-trending fault is mapped as separating the Sheephole Mountains from dissected, calcified 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 2200', Microwave relay tower.
is on the right. Look to the west at the eroded granite 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). Jackson 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). Celerite (stannous 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, 1955; 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 Gundy, this volume).

167.4 (0.4) The very rare mineral Antarciticite 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 (Q1) 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 across 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 area. 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 RAILCYCLE 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 shoreface 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 RAILCYCLE project, a brief review of the historic, ethnographic, and prehistoric background of the area is presented by Letch (this volume).

The Hope-Now Method mine is approximately 1 mile north along Kelbaker Road. Collectors have recovered a variety of uranium minerals, including rare fluorite, at this prospect. The Iron Hat mine, at 10:5, was mined in the 1940s; the iron ore (hematite and magnetite) occurred in small, shallow lenses in Cambrian limestone (Wright and others, 1953).
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 granite 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 forms, 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 warn, "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 mile road to pipeline mile marker 215

193.0 (0.5) STOP. CADIZ PLEISTOCENE SEDIMENTS at Mile Marker 215 (Fig. 7). The pipeline right of way cuts through white and tan calcite carbonate about 20' 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.

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 (5.9) STOP 8. ARCHER TUFF LOCALITY. Park at the bend in Cadiz Road. The Archer tuff locality, on the south side of road, consists of clastic fluvio-lacustrine sediments and root casts with an abrupt transition to massive carbonates (Fig.

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

Figure 8. Fluvial sediments containing root casts in sharp contact with columnar to massive pedogenic carbonate at Archer, Stop 8. R.E. Reynolds photo.
208.8 (2.5) Site of Archie. Watch for dips; do not take the road heading south. The Old Woman Mountains are visible on the skyline (Heward 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, 1959).

208.6 (1.5) 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 Klippe hills (Wright and others, 1958).

212.6 (3.8) Fossil. 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, -910'; Cadiz basin, due south, is -70' lower (see Reynolds and Knolls, this volume). Ward Valley is to the north, the Turtle Mountains are at 1100, and the West Riverside Mountains at 1200. The Arco Mountains lie at 1000, the Iron Mountains at 300. The turnoff south goes to the Standard Salt Company (do not take).

218.3 (1.5) Site of Milligan.

219.6 (1.3) Standard Salt processing plant (Fig. 10); (see Cadiz, 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 (0.6) Power line road in Ward Valley. The road south to Iron Mountain is closed. 1-40 can be reached to the north. Proceed easterly, towards Bice.

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.
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 67. The Granite Mountains at 11,000, the Iron Mountains at 3,000.

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 Pleistocene times (see Brown and Rosen, this volume).

242.1 (5.0) Blythe. Emergency gas available. Do not take the Rice/Blythe Road heading south. The Turtle Mountains are at 9:00 to the north (see Halett, 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 1941. Other camps were Young, Cucumb, Granite, Iron Mountain, Rio, 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) Nose west-dipping Tertiary volcanics to the north at 3:00.

247.7 (1.5) Pass the site of Groomev. We are at the divide between Rice Valley (to the west) and Vidal Valley, at elevation 1,950’ (Carr, 1981).

248.1 (0.4) Cross the railroad tracks. Castle Rock is at 10:00, between the Turtles and the Mopah Range.
255.2 (7.1) Road cuts dissect calcilified Pleistocene fanglomerate. Pyramid Butte is at 10.00.

256.9 (17) Junction of CA 62 and CA 95 at Vidal Junction. DigoVE 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 (0.0) 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 the 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.

Clastic 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 1).

--- 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, characterized by 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 volcanic rocks and red sandstone of the Turk Mine and Twin Lode Mine (?) formations, which predate the Peach 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/Na values in typical andesites are commonly less than 1%; K/Na 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 (0.1) Green sediments of the Bouse Formation (Mazzei, 1968) on the north side of the road arc 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 Piker. The white bed in the roadcut on the south side of the road before reaching the bridge is the basal sand 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 part 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) Step, TURN NORTH (left) onto AZ Highway 95.

9.5 (0.9) View at 2:00 of white Bouse sediments ringing canyon. The white bed exposed to the east of the road is the basal sand of the Bouse Formation. Cross Osborne Wash.


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 shelled, both due to Miocene compressional faulting and older (Cretaceous and Proterozoic) deformation. As a result, it erodes readily, and typically forms low, rubble-covered hills with poorly integrated drainage—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 redrocks are part of the Copper Basin Formation (Tess 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.

21.3 (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

Figure 14. Shelters dug out under the basal sand of the Bouse Formation. R.E. Reynolds' photo.
east-southeast. Clast types include Mesozoic (?) metasedimentary and metavolcanic rocks; such clasts are found in the Buckskin Mountains but are absent from the Whipple Mountains.

23.2 (0.3) The strata underlying the cliff-forming red beds, part of the Gene Canyon Formation, are dominated by monolithologic breccia beds. Boulder-size material is common. The beds generally are giant-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 side, 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 olive-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 (0.6) 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

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 site 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 red beds.

24.2 (0.3) Stop, TURN SOUTH (right) on CA Hwy 62/AZ 95 and go south.

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

extensional corridor.) This badly-shattered 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 basin 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 crustal. Walk up section along the road.

A basement assemblage dominated by the Granite Porphyry of Parker Dam is unconformably overlain by sandstones, conglomerates, monolithologic breccia, and rare limestones of the Gene Canyon Formation. Note the textural variability of the unit, and the lack of volcanic rocks as clasts. Stratigraphic relations 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 turnout 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.
26.8 (0.6) The coarse-grained deposits viewed at mile 23.3 can be seen eastward across the river. These rocks are part of the same fault block as the Parker Dam region.

27.3 (0.5) The Gene Canyon—Coppper 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 basaltic flows 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.

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 sediments. The Whipple Mountains (Fig. 16) are north.

31.3 (0.4) Transition from red volcanlastic sediments to mostly Protocrocite 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 splits into the Whipple Detachment Fault.

32.9 (0.7) Bulldog 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 foreshore 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) Bouse 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 Osbourne 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 Mead Vista 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 Osbourne 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.
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 wash north of the wash. The western half of Black Peak is part of the Colorado River Indian Trusts 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 Poic Spring Tuff. The red beds are conformably overlain by olive basalt which makes up the top of Black Peak. The basalt stack contains an angular unconformity within it. A breccia unconformably separates the flanking Phonolite 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 southeast (Fig. 17). The bluff to pale pink unit is an Upper Miocene fanglomerate (Osborne Wash; Busing, 1980; Osborne Wash Formation of Davis and others, 1980), this is capped by 1-2 m of yellow sandstone. The yellow sandstone locally intertongues 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 ash of the Bouse Formation basin fill association (Busing, 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. 18c) also note the local abundance of oosparodes and cm-scale gastropods.

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

50.1 (0.3) Return to Shea Road; go west.

55.5 (3.3) Stop, cross the railroad tracks.
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 Jurassic Formation capped by Tertiary-Quaternary Colorado River gravels and Quaternary alluvial/pleistocene 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 mass of the Riverside also consist of Bouse Formation capped by Quaternary gravels.

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

63.9 (0.2) 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 flanking Bouse Formation basin fill strata and the crystalline topography at the base of the Flamingo 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.
75.3 (0.5) Turn left at the pole line, before a bend in the pavement. If Highway 95 has turned from north to NADW, you have gone too far. Follow the road approximately west into low hills ranking 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 TUFAS. 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 greenish basement rocks (black and dark red-brown) with a discontinuous veneer of Bouse Formation shoreline tufa (gray) (Fig. 19). Tufa forms a rim on basement exposures; locally, the base of the tufa includes a monolithic breccia of angular basement-class 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 butting 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.

85.5 (1.8) TURN RIGHT (west) off Highway 95 onto Wilson Road (2 lane guided dirt).

B3.8 (0.9) Road divides; take the left fork.

B3.9 (1.0) TURN LEFT on small, quiet, ungraded trail.

86.6 (0.7) STOP 7: BOUSE FORMATION SHORELINE COMPLEX; GRAVELS OF MULTIPLE AFFINITY. Take the 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 youth of turn-around.

The basement rock north of the drainage is the same greas 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 spot 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 breccia 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 Calzoza 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 class assemblage includes locally derived quartzofeldspathic greis 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 (affiliates) of the various gravels/conglomerates exposed here. Gravel overlie tufa; these are probably Bouse. Gravels are overlain by tufa; these may be Bouse Formation or O'Brien Wash. Less consolidated gravels overlie the Bouse gravels of the "narrow" - are these post-Bouse, or just a younger generation of Bouse gravel? 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. (On turn south, if you want to go to Blythe, 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. Agnew and Reynolds (this volume) discuss Irvingtonian LMA mammal fossils which indicate that some of these sediments were deposited in middle Pleistocene times.
Retrace your route to Highway 95.
87.1 (8.8) 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 (1.6) 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 3 route to Twentynine Palms. At Quartzsite, 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 northeastern of the east-trending faults which characterize the eastern Transverse Ranges in southern California. Dobbele (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 of result of deformations of the San Andreas Fault in the San Gorgonio Pass area (Melti 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 Pleistocene 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 tanglelormite 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.

LATE CENOZOIC SEDIMENTARY ROCKS
Late Cenozoic sedimentary rocks are exposed south of Twenty-nine 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 arkose sandstone (Arkose). It is unconformably overlain by a conglomerate characterized by quartzite clasts (Quartzite Clast

Figure 1. Faults and rock units in the Yucca Valley vicinity.
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 arkose section. This deposit is in a tight fold with a core of metamorphic rock immediately below near horizontal Basalt Clast Fanglomerate. The rock 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 grader upward from thick bedded combinations of greenish-gray claystone, siltstone and fine-grained sandstone (basal facies) to a reddish-brown arkosic sandstone, cobbles, 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 class 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. Basaltic 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. Basaltic clasts containing black glassy megacrysts have a 2 to 3 mm thick medium gray rind or subdued rind, 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, 1985).

**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 cannot 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 Clay Fanglomerate. The uplift appears to have been active during the Pliocene.

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

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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 Clarendon LMA (11.5 Ma) fossil vertebrates in the upper half of the section and by the Hesperornithoid Ma (8.8 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 Pano Mountains. Tertiary sediments are on the eastern flank of the San Bernardino Mountains record periods of uplift and deposition. In the western San Bernardino Mountains at Cajon Pass (Woodburne and Gola, 1972; Woodburne, 1991a), the age of the late middle Miocene Cajon Formation (Meesing and Welden, 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 the Morongo 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 basin (6.86 ± 0.25 Ma, Lowman, 1989).

Suddes (1982,1983), Strathouse (1982), Neville (1983), and Lowman (1989) recognized this age discrepancy and referred the sediments at Pioneertown to the Santa Ana Sandstone. The type area of the Santa Ana Sandstone is 12 miles to the west at Barstow Flats, along the Santa Ana River drainage. Lowman (1989) divides the sediments at Pioneertown into a lower brown member containing fragments of gnesis, 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 SBCHM 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 1.

PIONEERTOWN FAUNA

Hyracotherium sp., the camel, and the sciuroid are known throughout the middle and late Miocene (Savage and Russell, 1983). Cynocephalus sp. species are known from North America from approximately 16.4 Ma (Lindsay, 1991; Woodburne, 1991b; Reynolds, 1991b) to latest Hemphillian LMA (Cracraft

Table 1. Pioneertown Composite Fauna

<table>
<thead>
<tr>
<th>Species</th>
<th>Habitat</th>
</tr>
</thead>
<tbody>
<tr>
<td>Myopipus sp.</td>
<td>rabbit</td>
</tr>
<tr>
<td>Gaudrypatheres dorsalis</td>
<td></td>
</tr>
<tr>
<td>Antilocapridae (H)</td>
<td>elephant</td>
</tr>
<tr>
<td>Camelidae (cani)</td>
<td>small camel</td>
</tr>
<tr>
<td>Equidae</td>
<td>horse</td>
</tr>
<tr>
<td>Sciuroidae</td>
<td>squirrel</td>
</tr>
<tr>
<td>Geomyidae</td>
<td>geomyid rodent</td>
</tr>
<tr>
<td>Proc 사람들이信访</td>
<td>pocket mouse</td>
</tr>
<tr>
<td>Cupulidimus n. sp. (cani)</td>
<td>kangaroo rat</td>
</tr>
<tr>
<td>Cupulidimus, sp. cf. C. aravatensis</td>
<td>kangaroo rat</td>
</tr>
<tr>
<td>Cynocephalus sp.</td>
<td>ord iro rodent</td>
</tr>
</tbody>
</table>


The geomorphology of the primitive elephant first appeared in North America about 15 m.y.a. (Woodburne, 1991b) and persists through the Miocene.

The simian rodent is similar in measurements and morphology to Paraprintopithecus sp. from the Clarendon LMA pit of the Crowder Formation (Reynolds, 1991a) and to an unnamed simian from the early Clarendon LMA Formation (SBCHM collections). The equid from Pioneertown consists of an upper molar fragment. Although worn, the remaining portion is as tall as un worn molar of Myopipus sp. from the late Barstow LMA. The fourth premolar crown is significantly larger in diameter and in height than Myopipus sp. from the Barstow LMA. Both these specimens suggest a LMA younger than Barstovian.

Cupulidimus n. sp. (smial) from Pioneertown is smaller than C. lindseyi from the Barstovian LMA (Barnosky, 1986) and compares very well with the small species from the early Clarendon LMA of Avawatz. The simple morphology of the P. suggests that this species was derived from C. lindseyi. Cupulidimus sp. cf. C. aravatensis is a large dipodomyonidae and compares well morphometrically with C. aravatensis from Avawatz. The early Clarendon LMA Avawatz fauna is dated at 10.9 Ma (Evenden and others, 1974, recalculated after Dalrymple, 1979).
DISCUSSION

The fauna from sediments north of Pionertown appears to be referable to the early Cladoniodrion LMA. Two similar taxa of Cladoniodrion occur at Ayavake with radiometric dates of 10.9 Ma (Brown and others, 1984). The Cladoniodrion LMA spans a period of time from approximately 11.5-8.5 Ma (Woodburne, 1987, 1991). Thus the age of the lower member of the Tertiary sediments north of Pionertown may be as old as 11.5 Ma, while the upper member intertongues with basalts dated between 9.3±0.7 Ma and 6.86±0.26 Ma.

The Pionertown sequence was deposited on a middle Tertiary erosional surface developed on granitic rocks (Oberlander, 1972). Judging by the northwestern 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 Pionertown sequence, the Santa Ana Sandstone (Sadler, 1985; Stratthouse, 1983) was deposited between 7.4 Ma and includes 6.2 Ma basalts (Woodburne, 1979) low in the sequence. Sadler (1985) recognizes basaltic asedimentary rocks that contain middle Miocene (Hemingfordian LMA) Mammals. He notes that this supports the presence of a nonconformity between basalts 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 Pionertown. The uppermost sediments north of Pionertown may overlap in time the oldest, quartzite facies of the Santa Ana Sandstone, but they are lithologically distinct. The stratigraphic status of the sediments north of Pionertown can not be evaluated until possibly similar deposits (such as those recognized by Grimes in Yucca Valley) are studied. Until that time, the sediments north of Pionertown that include early Cladoniodrion LMA vertebrate fossils and which interfinger with dated basalts may be referred to the "Pionertown Sequence" to indicate their temporal and geographic distinction from the Santa Ana Sandstone.

ACKNOWLEDGEMENTS

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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 Yuca Valley and east of the Big Horn 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 overlies 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 as 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. Pipe's Wash flows by 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 range is the north and east the Roden and Bellon Mountains, respectively. The area is named (Umberger, 1992) for its most distinctive geographic feature, Spy Mountain, which lies in Homestead Valley just south of Johnson Valley. An unusual, 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 McWhorter of San Bernardino Valley College; Hill and others, of the California Institute of Technology (CIT); and CDMG: Ebil, Hill and Peckman from CIT; and Hutton and others, from CIT and the USGS.

Manson (1980) 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 California Natural Special Study 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 (1977) 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 tephrochronologic and geomorphic expression. The Pleistocene is represented by older tafoni 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 interfinge with alluvial deposits that nonconformably overlie Mesozoic granite (Mg2) comprising the pediment in the core of Spy Mountain. Some of the homestead region 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 fans deposits are composed of sand and gravel derived from grooves and granitoid rocks that make up Homestead Mountain. These fans have been elevated and currently are being incised.

Old Caliche Deposits

Low-density, 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 Line 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 upfited 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 coarse sediments 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-contorted 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 lessic gravel 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, granite, 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 argillaceous, 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, split-rich sand with coarse gravel and a poorly-sorted, sand-supported matrix consisting of calcareous-cemented, clayey sand. Clasts consist dominantly of garnet aplite, and also include representatives of 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 10°13’16”NE. Source of the tilt may be derived from drift 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 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 is 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 overlaps the trace of the Homestead Valley fault.

Colluvium

Colluvium consists of granitic and metamorphic residual soil derived from the San Bernardino-Joshua Tree (7) 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, tanned 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. Calcite 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 predominately 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, un cemented, 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, 1960-1986) indicates that depth of alluvium ranges anywhere from 28 to 335 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 valley 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.

Aquclic Sand Deposits

Holocene aeolian sand is deposited on the surface of the eastern banks of Pipes Wash, where it intertongues with young alluvium. It overlies Miocene bedrock and older alluvium in Pipes Wash. The stabilized portions of the deposit supports vegetation, while newly arrived sands remain unconsolidated. 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 southwesternly 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 (Brown 1985; Garfinkel, 1972). 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 (Mogm) over Quaternary debris-flow deposits (Qd). One fault plane measured in the field showed the fault to be moderately-laterally-dipping at N30W/55W. Slickensides in the clay gouge of the exposed fault plane trend N12E and plunge 415W.

A second measurement taken approximately 100 m northward of the initial exposure reveals the fault as a vertical planar striking N45W (photo 25), with the Homestead Mountain block uplifted relative to Pipes Wash. Between the two stations, a third reading was N33W/65W.

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 subaerial, eroded old fault scarp 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, lithological signature and reflectivity.

These possible old Homestead Valley fault scarp consists of a loosely consolidated material with gravel and debris. Several talus slopes (colluvium) overlie a possibly uplifted old stream terrace consisting of older alluvium (Qua) that includes clasts of volcanic rocks. Two talus cliffs, believed to have a Pleistocene Basalt provenance (Lowman, 1989), were transported to the area via Pipes Wash. Existence of the basalt cliffs 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 downslope end in Pipes Wash (at the southeast corner of the intersection of Lien Road and Belfield Road), residents reportedly only drilled 10 m to reach groundwater.

SEISMICITY

Seismograph studies conducted by Steenken and others, (1983) subsequent to the 1795 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 (Mjölnink, 1980). Right-lateral strike-slip movement was documented with at least 10 cm near the northern end of the fracture zone, and as much as 25 cm along a 50 km section of the Homestead Valley Fault north of the study area, Ebel and others, (1980) calculated seismic moment to be 7 x 10^14 dynes 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 suggests 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 1795 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 downsloping, 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 acknowledge, however, that the crack could merely be an incipient slump crack.

The linearly 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, connect west, can be seen in the orthophoto quads.

Geomorphology

The Spy Mountain region is a desert landscape with boulder mantles 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 rounder crests. The "V"-ended, rugged topography rendered by phyllic rock suites is typical of domed isles (Oberlander, 1972), including boulder mantled slopes and pediments with guse between them. Cohesional joining 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 desiccation 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, colluvial steps abruptly at what is either a meander scar cut by Pipes Wash, or the buried trace of the Homestead Valley fault as mapped by Dibblee (1967). The poorly exposed fault in this section of Homestead Mountain has elevated older alluvial deposits, which subsequently have been undercut, and which grade into the slightly inclined, young alluvium of Pipes Wash. To the northwest, the trend of the fault is physiographically well-expressed.

The homoclinal shape of Spy Mountain is an anomaly. One possible explanation in the younger, pediment-forming granite (MgS) in the concave core of Spy Mountain. The granite was intruded into the inner and central margins of the ridge-forming biotite granite. This granite slopes gently westward at about 3° into older alluvium and aeolian sand deposits. It is phaneritic and has a tendency to flake; being coarse-grained than the older granite, it is less resistant and has eroded so substantially as to give the center of Spy Mountain as 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 plumes 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 this interpretation, the Spy Mountain region represents the northeastern extreme of an elevated surface planus sloping gradually eastward into the Mojave Desert plain.


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U.S. Department of Agriculture, 1952-53, aerial photographs nos. AXL 10-X-126 through 128; AXL 38K-26 through 29; scale 1 inch = 1,000 feet.


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 Flat 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 Corps 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 (1979) 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 orientation. Mesquite Lake fault trends NW, west of the stripes.](image1)

![Figure 3. Vegetated, linear depression formed by infilling of a ground crack formed in 1976. 45cm-long geologic pick for scale.](image2)

Trenching of the fissures by Wahler Associates (1984) showed that these features are formed by tension with open linear cavities below the filled depression 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 (1979) 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 (1979, 1980) that these cracks were essentially the surface expression of a left-lateral fault. However, the subsurface cavities and lineations suggest tension as the stress mechanism, not compression and shear, and it appears the fault model is untenable. Neal and others (1980) 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 (1985) 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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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 geographic setting within the eastern Mojave Desert include Pinto Basin (Campbell and Campbell, 1935; Scharf, 1935; Jefferson, 1986, 1991a) and Pinto 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 1916 (Basset and Kupfer, 1964), and Bacheller (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

<table>
<thead>
<tr>
<th>Locality</th>
<th>Institutional Locality Number</th>
<th>LACM 4281-4283</th>
<th>SBCM 1.86.4, 1.86.9, 1.86.11-1.86.13</th>
</tr>
</thead>
<tbody>
<tr>
<td>Campbell Hill</td>
<td>Institution for locality number: LACM 3350</td>
<td>Equisetum sp.</td>
<td></td>
</tr>
<tr>
<td>Surprise Springs</td>
<td>Institution for locality number: SBCM 1.86.1-1.86.3</td>
<td>Camptosorus sp.</td>
<td></td>
</tr>
</tbody>
</table>

CAMPBELL HILL

At Campbell Hill (3 2.4 km east and northeast of Twenty-nine Palms), vertebrate remains were recovered 73 m below the eroded top of an approximately 358 m thick stratigraphic section of fluvialite sandstones and gravels. The deposit is informally known as the Campbell Hill Formation (CHF) (Bacheller, 1978), and was mapped as Quaternary Older gravels (Qog) by Dickie (1968). This unit contains lithologically distinctive sedimentary layers derived from the San Bernardino Mountains about 45 km west of Twenty-nine Palms (Dickie, 1968; Bacheller, 1978). The beds have been uplifted along the northwest-southeast trending Mesquite Lake Fault, are gently folded, and dip about 30 degrees to the northeast.

TWENTYNINE PALMS GRAVEL PIT

The Twenty-nine Palms Gravel Pit area (4 km east of Twenty-nine Palms and south of Highway 62) lies immediately west of the main trail 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 Dickie (1968) mapped this area as Qoa (Quaternary older alluvium), it is clear from the detailed work of Bacheller (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 Dickie, 1968) or the Twenty-nine Palms Formation (TPF) (informally named by Bacheller, 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 (Bacheller, 1978). Distinctive sedimentary layers in the TPF are derived from the Little San Bernardino and Pinto Mountains a few km south of Twenty-nine Palms.

SURPRISE SPRINGS

The vertebrate remains from Surprise Springs (approximately 20 km northwest of the Twenty-nine Palms Gravel Pit and 8.5 km west of Deadman Lake) were also recovered from sediments mapped as CHF (Bacheller, 1978) (= in part to Qoa of Dickie, 1967). Near Surprise Springs, the fiat-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 30.
VEGETATION REMAINS

Age and Correlation

The vegetation remains from Campbell Hill, Surprise Springs, and the Twentynine Palms Gravel Pit localities were recovered from fluvial and alluvial deposits that are part laterally correlated, and are probably of comparable, late mid-Pleistocene age (Morrison, 1991). The Bishop Tuff, dated at 750 ka BP (Sarna-Wojcicki and others, 1984), has been tentatively identified 90 m below the top of the Greenish Silt 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 Blain sp. has not been confirmed from Campbell Hill or the Twentynine Palms Gravel Pit localities, Blain sp. from the CHF at Surprise Springs may approach the age of the earliest record in the region (about 280 ka BP at Lake Mono).

Campbell Hill (San Bernardino County Museum, SBICM 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 Rancho La Brea assemblages. Many of the taxa are regionally very rare (Jefferson, 1994b). Although the collections are not large, a total of fourteen mammals have been identified (Table 1). Of the three represented carnivore species, Taxiwasps sp. has been reported from two localities (Koksewad Cave, Schuiling Cave, Fols canyon from three locations (Lake Manix, Mitchell Caverns, Schuiling Cave), and Smiladons sp. from only one other locality (Lake China) in the Mojave Desert region (Table 2). In addition, the extinct prosimian Calapennex sp. is known from only one other desert locality (Schuiling Cave), and the occurrence of Oleodontis sp., cf. O. virginiensis 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 canadensis, Canis familiaris, and Osiris sp. of O. canadensis — probably reflects the paucity 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 Equids sp., Hemischnicus sp., and an abundance of Caninales 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 Oleodontis sp., Antilocapra sp., or Osiris sp. (Jefferson, 1991b).

Within the Mojave Desert region, Oleodontis sp. cf. O. virginiensis is represented by a single, nearly complete left otter from the Twentynine Palms Gravel

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

<table>
<thead>
<tr>
<th>TAXA</th>
<th>d</th>
<th>e</th>
<th>f</th>
<th>g</th>
</tr>
</thead>
<tbody>
<tr>
<td>Megapelyx</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nototharchus</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Glycyphus</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Camelae</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Antilocapra</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oxydalis</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Equus (large)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Equus (small)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Equus sp.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Caninales</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kentodon</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Macrauchenia</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Navajoceros</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Odocoileus</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Capromeryx</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Antilocapra</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pteralopryx</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bison (large)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bison (small)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Osiris sp.</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

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

<table>
<thead>
<tr>
<th>TAXA</th>
<th>c</th>
<th>d</th>
<th>m</th>
<th>CM</th>
<th>P</th>
</tr>
</thead>
<tbody>
<tr>
<td>Equus sp.</td>
<td></td>
<td></td>
<td></td>
<td>1</td>
<td></td>
</tr>
<tr>
<td>Equus sp. (large)</td>
<td>5</td>
<td>4</td>
<td>4</td>
<td>2</td>
<td>1</td>
</tr>
<tr>
<td>Equus sp. (small)</td>
<td>26</td>
<td>9</td>
<td>36</td>
<td>10</td>
<td></td>
</tr>
<tr>
<td>Equus sp. (small)</td>
<td>26</td>
<td>7</td>
<td>2</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>Caninales sp.</td>
<td>56</td>
<td>21</td>
<td>64</td>
<td>21</td>
<td>6</td>
</tr>
<tr>
<td>Antilocapra sp.</td>
<td>12</td>
<td>15</td>
<td>12</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>Bison sp.</td>
<td>11</td>
<td>29</td>
<td>6</td>
<td>12</td>
<td>3</td>
</tr>
</tbody>
</table>

Locality follows from west to east (left to right). Abbreviations: L = Lake China localities (NISP = 963); D = Dome Springs localities (NISP = 148); M = Lake Manix localities (NISP = 766); CST = Twentynine Palms area localities (NISP = 97); V = Vale Springs localities (NISP = 1639); Lake Valley localities.
fit locality, SBCM 1863-6. It is an eurytropical 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 extralimital species known from the eastern Mojave Desert region, such as the packrat Neotoma albigula (Jefferson, 1991a).

Blim sp. is known from only six of 47 (13%) late Pleistocene localities that yield large mammals in the greater Mojave Desert region (Table 2). There are Lake China (Forth, 1978), Dove Springs Wash (Whistler and others, 1991), Lake Manix, Pinto Basin, Piute Valley, and Surprise Springs (Jefferson, 1992b). This distribution may reflect the presence of seasonally favorable higher elevation habitats near localities where Blim 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). Blim 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 southwestern Great Basin also yield Blim sp. remains. Blim sp. is about twice as abundant throughout the Great Basin than in the Mojave Desert, and is present in 31 out of 41 (76%) large mammal assemblages from Nevada.

Most of the other large mammals from localities in the Twenty Nine Palms area occur throughout the region (Jefferson, 1991b). Canis lupus familiaris, Hesperocyon mackenzi, 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 Twenty Nine Palms area appears similar to that in the Lake China fauna (Table 3).

ACKNOWLEDGEMENTS

I thank C. J. Bell, R.E. Reynolds, and R.L. Reynolds of SBCM, J. H. Hutchins of the Museum of Paleontology, University of California, Berkeley, and S. McLeod of the LACM who provided information and/or allowed access to collections under their care. J. V. Kelly kindly shared regional expertise and data. R.E. Reynolds of the SBCM, and J. M. Harris and C.A. Shaw of LACM reviewed the manuscript and provided helpful comments and suggestions.

LITERATURE CITED


Eolian 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 northwest flanks of the Sheephole Mountains and the adjoining Pinto Mountains (Fig. 2). Sand ramps are formed primarily by episodic deposition of eolian 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 eolian deposits are rare in the southwestern deserts, and therefore these exposures are important for analyzing the paleoenvironmental conditions of eolian 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 eolian transport and deposition (William and others, 1991).

GEOMORPHOLOGY OF THE SAND SHEET
At the Dale Lake Sand Sheet and surrounding areas, up to six eolian depositional phases separated by paleosols have been identified, on the basis of geomorphology, 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 Q3. The most well sorted sediments are found in the crest of active transverse and complex dunes in Cadiz Valley, while the most poorly sorted sediments are from Q3. Highest silt-clay percentages (4.8%) are found in the stabilized climbing dunes of the Calumet Mountains, followed by Q2 (3.1%) and Q1 (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 trasluciation 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 evaporates, such as calcium carbonate and sodium carbonate, owing to the fact that most dune fields are near to and downwind from playa basins, potential sources for both deflation and groundwater soluble. Airborne evaporate materials also enhance the Fh conditions of the available moisture, thereby facilitating silica dissolution and reprecipitation (Tchakerian, 1989; Pye & Tosar, 1990). Vegetated linear dunes in Rice Valley, California, and in the Cactus Plain in southwestern 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 Q1 and Q2 show high percentages of solution etchings, large hollows and pits (ranging in diameter from 5 to over 20 microns), blocked 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 Q3 and Q2, is found largely as a result of chemical weathering processes owing to the presence of vadose water and alkaline solutions (including desert dew).
which dissolve quartz (and silica minerals) and then promote reprecipitation during evaporative periods (Pye and Tsoar, 1990). The quartz grains from the stabilized clearing dunes on the Calumet Mountains show more pronounced chemical weathering microfeatures than those from Qe1 and Qe3, 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 35% 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 aeolian deposits can be distinguished. The following micromorphological characteristics associated with mechanical and chemical weathering processes as well as with grains surface morphologies were used for this analysis: breakage blocks, conchooidal 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 aeolian 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 aeolian depositional sites in the east-central Mojave Desert) indicate significant oscillations in aeolian activity during late Quaternary time (Tchakerian, 1989, 1991). The aeolian deposits are believed to have accumulated largely in response to lowering of lake levels and a consequent increase in fine sediment availability in these desiccating lake basins, possibly accompanied by winds that were stronger and more persistent than today (Wells and others, 1987; Tchakerian, 1989).

Six stages of aeolian 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...

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

Figure 3. Sand ramp and dune wall marked X in Fig. 2. Sheepole Mountains in background. Exposed sand ramp here is about 20 m in depth.
transitional episode between Qc2 and Qc4. The alluvial units in Stage V (Qc4, Qe5, Qc6) were most likely deposited during the Holocene. The uppermost unit, Qc6, yielded a calendar-
ecological 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 Qc4, Qc5, and Qc6, indicating their similarities with respect to quartz grain microstructures and, thus, most likely representing 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 Alithermal (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 alluvial 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 cores of the transverse and complex dunes in Cadiz Valley exhibit surface micromorphologies indicative of active alluvial processes, such as high percentages of upturned plates and meandering grooves and ridges (Trubey and others, 1989). The sand dunes contain well preserved primary sedimentary structures and show no soil or calcium carbonate development.
A preliminary study of the alluvial stratigraphy in a sand ramp at the eastern edges of Iron Mountain (about 30 km east of the Lake Sand Sheet) also indicates multiple episodes of alluvial deposition (V. Trubey, unpublished field notes). Up to three alluvial depositional pulses separated by paleosols have been tentatively identified. The sand ramp is veneered by rock talus from the mountains and terminated by several ephemeral streams, and contains dune wadis similar to those found in the Dake Lake Sand Sheet.
A tentative evaluation of the absolute dates of the alluvial sediments at the Dake Lake Sand Sheet

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

<table>
<thead>
<tr>
<th>Grain Size</th>
<th>Mean</th>
<th>Standard Deviation</th>
<th>Skewness</th>
<th>Kurtosis</th>
<th>Percent Silts and Clay</th>
</tr>
</thead>
<tbody>
<tr>
<td>Qc1</td>
<td>2.44</td>
<td>0.33</td>
<td>0.18</td>
<td>1.17</td>
<td>2.00</td>
</tr>
<tr>
<td>Qc2</td>
<td>1.86</td>
<td>0.81</td>
<td>0.14</td>
<td>1.15</td>
<td>1.36</td>
</tr>
<tr>
<td>Qc3</td>
<td>1.75</td>
<td>1.17</td>
<td>0.05</td>
<td>1.96</td>
<td>1.33</td>
</tr>
<tr>
<td>Qc4</td>
<td>2.48</td>
<td>0.57</td>
<td>0.23</td>
<td>1.35</td>
<td>2.35</td>
</tr>
<tr>
<td>Qc5</td>
<td>2.33</td>
<td>0.29</td>
<td>0.07</td>
<td>1.90</td>
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</tr>
<tr>
<td>Qc6</td>
<td>2.29</td>
<td>0.75</td>
<td>0.15</td>
<td>0.95</td>
<td>1.35</td>
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<tr>
<td>Qc7</td>
<td>2.91</td>
<td>0.87</td>
<td>0.22</td>
<td>1.29</td>
<td>0.83</td>
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<tr>
<td>Qc8</td>
<td>3.35</td>
<td>0.28</td>
<td>0.03</td>
<td>0.99</td>
<td>0.59</td>
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<table>
<thead>
<tr>
<th>Grain Size</th>
<th>Very Fine</th>
<th>Coarse</th>
<th>Medull</th>
<th>Fine</th>
<th>Very Fine</th>
</tr>
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<tbody>
<tr>
<td>Qc1</td>
<td>0.48</td>
<td>0.14</td>
<td>0.75</td>
<td>0.00</td>
<td>0.03</td>
</tr>
<tr>
<td>Qc2</td>
<td>0.46</td>
<td>0.13</td>
<td>0.45</td>
<td>0.00</td>
<td>0.08</td>
</tr>
<tr>
<td>Qc3</td>
<td>1.31</td>
<td>0.95</td>
<td>0.66</td>
<td>0.14</td>
<td>0.50</td>
</tr>
<tr>
<td>Qc4</td>
<td>1.27</td>
<td>0.76</td>
<td>0.45</td>
<td>0.25</td>
<td>0.76</td>
</tr>
<tr>
<td>Qc5</td>
<td>1.50</td>
<td>0.45</td>
<td>0.55</td>
<td>0.10</td>
<td>0.25</td>
</tr>
<tr>
<td>Qc6</td>
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<td>0.80</td>
<td>0.20</td>
<td>0.15</td>
<td>0.10</td>
</tr>
<tr>
<td>Qc7</td>
<td>3.05</td>
<td>0.85</td>
<td>0.45</td>
<td>0.68</td>
<td>0.42</td>
</tr>
</tbody>
</table>

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

<table>
<thead>
<tr>
<th>Episode</th>
<th>Grain Size</th>
<th>Sorting Quality</th>
<th>Micromorphological Features</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>V</td>
<td>Active transverse dune</td>
<td>Units Qd1, Qd2 and Qd3. Alignment of surface features with rock fabric.</td>
</tr>
<tr>
<td>II</td>
<td>IV</td>
<td>Median Sorting Quality</td>
<td>Units Qd4, Qd5 and Qd6. Alignment of surface features with rock fabric.</td>
</tr>
<tr>
<td>III</td>
<td>IV</td>
<td>High Sorting Quality</td>
<td>Units Qd7, Qd8 and Qd9. Alignment of surface features with rock fabric.</td>
</tr>
<tr>
<td>IV</td>
<td>V</td>
<td>Very High Sorting Quality</td>
<td>Units Qd10, Qd11 and Qd12. Alignment of surface features with rock fabric.</td>
</tr>
</tbody>
</table>

Figure 4. Composite stratigraphic columns compiled from 4 detailed sections along dune wadis (Fig. 3).
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 Q6, Q10, & Q14 as one group. Clustering of the remaining aeolian units is similarly striking.

by thermoluminescence (TL) and infra-red stimulated luminescence (IRSL) techniques and cation-ratio (CR) and activation mass spectrometry (AMS) radiocarbon dates from rock varnish from stabilised 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 aeolian deposition in the Mojave Desert.

REFERENCES CITED


Cultural Resources at Dale Dry Lake

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 Dry 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—Twenty-nine 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 3,154 feet above sea level; the beach is about 1,188 feet above sea level.

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

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 big horn 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 (Schoff 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 1970; Simpson 1980). Childrens and Minshall 1980). It is
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. Wazzen (1964) 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 scrapers-plans, choppers, hammerstones, manos, and milling stones, with the mortar and pestle introduced during the latter part of the period. In addition, shell scrapers, incised slate, sandstone tablets and pendants, bone awls, and flail stone and Olsolky 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, 8-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 equalized 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 Aiken (1970) and by Wilke and Flenniken (1969). 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 replicate 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 (Yoke, 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 Hatakyaa 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 developments and a continuation of the influence from the Southwest (Warren 1984). Although the influence from the Anasazi declined, that of the Hatakyaa 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.

Protohistoric Occupation

The native American group in the immediate project area during the protohistoric and historic periods was the Serrano (Bean and Smith 1970). European influence on the Serrano was negligible until 1834 when most of the natives were moved to an assistance near Redlands. Prior to that time, according to Strong (1929), they were organized into local family lineages. Each local lineage 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, 1980) 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 rural land until the early 1900’s, 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 1872 (Vredenburgh, Shumway, and Harrill 1883), 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 Harrill 1901:141). A pump station to the south of Dale Dry Lake, shown on 1924 and 1955 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 1906 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 Levin Bush and
Dry Lake was investigated. Interest in Dale Dry Lake began with the procurement of a collection from Mr. Stonegraph in 1939. This was a general collection described under Accession Number 49C125 on as “Dale Dry Lake, approx. 21 miles east of Twenty-Nine Palms” (page 125). The collection included “150 Miscellaneous pieces (flints, arrowpoints, knives, chips, etc.) Possibly Chumashian. 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 Y31, E. F. Walker and his wife toured over collections from sites 428 and 479, 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 viewed 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-erising 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 quite extending out into the lake play, 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

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* X = in context, but not yet given. 1 = Noted as present on p. 2681, but not collected.
as several composite 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 potholes, 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 Vikings Headquaters, Joshua Tree National Monument, are summarized in Table 1. At this time, no attempt was made to compile 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 artifacts are points and "chips" or pieces of debitage. Lithic materials include obsidian, Jasper, chalcedony, chert, basalt, jasper-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 bone combs and perforates. The "crude points" (arrow and dart) include numerous perforates and rejects. In addition to the biface cores, simple platform, unidirectional cores are present. Core tools in the collection include hammerstones and scrapers. Plane tools. One "tombshell" 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 metal 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 collection.

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 (1981), involved only the intersection of Antboy 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 ofolithic 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 ACI, according to the Bureau of Land Management Desert Plan, and have special archaeological significance (US Department of the Interior 1983). 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 jacksnake (Leptus 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 in field observation, to be fresh water clam (Anodonta sp.). Only one potsherd, probably Tuzon 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 42.

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 plays 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, quartzites, 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 northernmost east of Dale Dry Lake. There are 17 mapped concrete 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. Groff (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 secured historic locus was located to the southwestern of Dale Lake in an area of sand dunes. It appears to be connect- ed with the gold mine district of Dale in that an old bottle, dating to between 1900 and 1950, was found at the site (Lorain, 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 Campbell suggests that whenever there was sufficient water in the lake to maintain a population, people used the resources. The presence of Amandita 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 "sinkers" found at site 438 may indicate the use of nets in a lake. It was found with Desert Sidetipped points and ceramics.

The numerous camps found by the Campbell at 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 (Schooth 1986). The survey was undertaken at the request of Dave Groff for the Western America Ore Co. and was performed by the author under the direction of Dr. P. J. Wilke (UCR-ARU Project No. 847). The maps used herein were reproduced using the Archaeological Research Unit computer software.

Lesler 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. Roscoe Pepliot, Curator at the Visiter's 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 Kitzmiller, 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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AAA. 1914. Map of a Portion of Southern California and southwestern Nevada embracing the Acid Regions of Mojave Desert, Colorado Basin, and Death Valley. Blueprint copy on file at the Archeo- logical Information Center, San Bernardino County, San Bernardino County Museum.


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


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 epirogenic structural depression about 200 km long that crosses the Mojave Desert along a west-northwest trend (Thompson, 1965; Bassett and Kupper, 1964; Gardiner, 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 Castle Lake playa lies slightly lower, at 165 m, and represents the regional 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 faults 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 Johns, 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.
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 (east Bristol basin) and southwest (Cadjie Lake basin). Structural boundaries evidently segment the basin surface under the Bristol-Danby trough in this area. The structures may be related to right-lateral faulting or to pull-apart between blocks affected by right-slip (Miller and others, 1982; Dukka and Travis, 1990; Richard and Dukka, 1993). 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 Cadjie 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 basins under Bristol Lake as conspicuously the deepest.


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 (555 km²) in a system of three northwes-southeast trending dry lakes (playas) located in a structural trough between the Bristol Mountains to the north and the Bullion and Shearlock 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.

GEOLAGIC 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 km²). The present total drainage area into Bristol Dry Lake, excluding drainage from Cadiz Dry Lake and Alkali Dry Lake, is just over 2000 km².

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 substantially 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 2-5 years when no precipitation has been recorded.

DEPOSITIONAL ENVIRONMENTS

Handford (1982a) 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 wash system, 3) the salt flat 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 source. 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 once sandy wash with a 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 cobble, gravel, and sand. Extensive beachrock calcite is present in poorly sorted cobble strata of the upper part of the mid fan. The calcrite 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 calcrite zone is widespread throughout the fan. The calcrite zone is approximately 1-3 m below the surface of the fan and it is overlain by a dark red paleosol. The paleosol is discontinuously overlain by a thin 1-3 m uncremented layer of low-angle trough cross-stratified braided
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 palaeosol and calcrite 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. Spring 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 downwind 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 onto the divide into the Cadiz Basin. No unequivocally aeolian
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 into 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 central basin. The details of this zone are presented in Bussen and Warren (1999). 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 sediments (quartz, feldspars, clays, and a large suite of heavy minerals), ranging from sand to clay-size particles, and authigenic rhombo calcite less than 1 mm 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 most 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, millimeter-laminated authigenic calcite is precipitated along with the detrital muds when there is a sufficient water body at the base of the mud flat.

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-rioting ground" and "polyn" (Thompson, 1929; Basset and others, 1969; Gale, 1951). The "polyn" is due to displaceable halite growing just below the surface and pushing the sediment upward. Desiccation cracks and millimeter-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. Displaceable halite crystals form in the sediment creating the "self-rioting 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-rioting 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 displaceable celestite and gypsum crystals are exposed on the surface of the playas.

Near the playa margin facies, the muds contain abundant molds of what were 1.0 mm displaceable halite cubes. These molds are now empty but are filled 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) celestite 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-thick, vertically elongated chevron halite, stippled by millimeter bands of detrital silt or mud. The halite is 99% pure, and both pans consist of approximately 30% chevrons, 40% clear diagenetic halite, and 20% porosity. Pores decrease from 30% in the modern salt pan to 6% in 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 dissolve the tepees flattening the salt pan to an almost levelized 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 plane of halite precipitation. The mud is thixotropic and structureless and there are no displaceable halite cubes in the mud.

Although the salt pan halite is only tens of millimeters thick at the surface, 3 meters below the surface there is a 1 m thick halite bed. In some cores, relatively pure (90-95%) 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%.

**FACES 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-crystalline halite and the brine pan, is dominantly by siliciclastic muds with only ephemeral displacive halite tubes and hopper crystals precipitating in the facies. The separation of the gypsum zone from the subsequently 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 anhydrites) are found in the saline mud flat and basin center, and much of this is later transported from the defaunting 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.

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**Figure 3.** Simplified core logs, Bristol Dry Lake, California. Sediments must contain >50% evaporite minerals before they are considered an "evaporite horizon."
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 °C measurements in basalt indicate that by 150 m depth, the chlorid is approximately 2 million years old (J. Emk., pers. comm., 1988), this dating technique is still experimental and may have large errors associated with it (Philips and others, 1980). Nevertheless, there is an indication that the basin is relatively old. Correlations of the deepest tephra layers indicate that the basin is at least 3.5 ± 2.2 million years old at those levels in the core. 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 core 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 Needleman quadrangle. Using an equation from Telford and others (1976), the total thickness of the basin center is estimated to be approximately 60 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 exist 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 playas 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 “Burns Lake” (now Dale Lake) and Bristol, Cadiz and Danby Lakes. The reader is referred to one review of the original literature by Pipke (1976) for greater elaboration of general playas characteristics, classifications, or definitions. Druse (1961) describes the variety of playa surfaces of the Mojave Desert, and reports on several groups of clay minerals (brucite 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 1932, there had been no comprehensive coverage since the 25 year period 1942 through 1967, such as the works by Tuckett and Simpson (1943), Wright and others (1953), Mumford (1954), Smith (1965), 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 (Brown and Pratt, 1977; 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 occurrence or development potential maps of overall areas have been and are currently being produced with coverage in the Dale, Bristol, Cadiz and Danby areas (Caiz 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 tectonostatigraphic terranes (Albers, 1982; Albers and Fratticelli, 1984) and for mineral deposit models (Cyk 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 resultant 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 Aubrey” (Fig. 1), Lake Le Conte was used for the area now superimposed 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 twofold-out map plates depicting dry lakes omit 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 (e.g., Blanc and Cleveland, 1961a, 1961b; Rose, 1961; Brown and Roser, this volume). Three 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 resistant evaporite types deposited in each basin over time in a chain of interconnected basins, are based on study of the palo-ovens w. drainages (Hardie and Eggerle, 1976; Dever, 1988). Based on the Hardie-Eggerle 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 plays, 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 (1956). This classification consists of five types: dry type, clay-encrusted/nash-encrusted wet type, crystal body type, compound type, and artificial. According to Ver Planck, Dale, Bristol, Cadiz, and Danby Lakes are 4 of the only 5 crystal body type plays in California (Searles Lake, San Bernardino County being the fifth).

Mineral Classification

The southern Mojave Desert lies within platform facies crustal terrace delineated by Albers (1981). As indicated on a mineral resources assessment map (Albers and Fraticelli, 1984), Dale, Bristol, Cadiz, and Danby Lakes all are classified within geologically favorable terraine that contain known saline deposits of potential commercial value as indicated by drill-hole data. There has been other classification of non-marine evaporite deposits and there is a need to perform modeling studies (Shelton and Raup, 1981). Mineral deposit models for the salines are not recognized in many works (Ridge, 1986; Skinner, 1985) including a more recent, major work and follow-up Open File Reports on mineral deposits (Cow and Singir, 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 plays can be described in increasing saline concentration as fresh, brackish, hypersaline, or evaporite, 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 evaporites. In another convention (Lloyd and Heathcote, 1988) the terms in increasing salinity are fresh, brackish, saline, and brine, brine having at least 100,000 total dissolved solids (TDS) or milligrams 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 gram/cm³. According to Ver Planck (1958), Dale brines were reported with a specific gravity of 1.21 to 1.25 gm/cm³. 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 as known Leasing Areas for sodium (a portion near productive wells). Bristol and Danby Lakes are also classified as Known Leasing Areas for sodium. Based on the California Desert Plan Geology-Energy-Mineral Resource Area Assessment, all four lakes are classified as having high potential for sodium mineral resources (Bureau of Land Management, 1980). For other minerals not leaseable, a mineral lands classification map depicts the productive 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

Figure 1. Map of Lake Le Conte and Aubury, showing plays along Bristo- Danby trough. From Bailey (1902).
crystalline sodium compounds for the four playa deposits are reported in Calza and others (1979).

SALINE MINERAL PRODUCTION HISTORY

Playa evaporite sediments have been worked for sodium chloride (rock salt, NaCl), sodium sulphate (salt cake, Na2SO4), and calcium sulphate (selenite-gypsum and gypsum, CaSO4·2H2O). Associated brines have been the source of common salt (NaCl) salt cake (Na2SO4) and liquid and anhydrous-fake calcium-chloride (CaCl2) with potassium chloride (KCl).

There is past production of sodium sulphate from Dale and Danby Lakes, and calcium sulphate from Bristot 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, 1066 and 1007 feet deep, bottomed in salt at Bristol Lake (Calza, 1979b). 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 (Calza, 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 523 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 (Calza, 1979c). A 440-foot deep drill hole revealed salt only to a depth of 2.5 feet in Dale Lake (Calza, 1979b).

Dale Lake

Initial exploration drilling took place between 1920 and 1934 by Irving Bush, Wright and others, 1953). Dale Chemical Company produced salt and sodium sulphate from well brine from 1931 to 1946, according to Ver Planck (1958). Dale Chemical Industries incorporated 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 loan unharvested salt remaining from Dale Chemical Industries (Ver Planck, 1958). Between 1949 and 1986, Southwest Salt Company acquired ownership of the operation and mineral rights. Southwest Salt turned over the property to G. Green Cott and family in the mid-1980s (as a result of Southwest Salt property acquisition by Morton-Thiokol, inc., Morton Salt, in several Mojave Desert dry lakes). An application for a sodium prospecting permit was made by Superior Salt Company (relation to several Cottos) and was granted in 1989 for looking peripheral federal lands adjacent to the private holdings dominating the deposit. No exploration occurred by the November 31, 1991, permit expiration.

Brill Lake

In 1908, Crystal Salt Company made placer claim locations for calcium chloride (Gale, 1951). Salt Company produced salt by stripping and blasting from 1909 to 1913, as Consumes Salt Company in 1916 and 1917, as Pacific Rock Salt Company in 1920, and as California Rock Salt Company from 1921 to 1950. In U.S. vs. California Rock Salt Company (1950, 60 L.D. 430, 442), all but two of 38 contiguous placer claims of California Rock Salt Company were obtained invalid based on the one claim per locator provision of the Act of January 31, 1901 (Saltine Place Act, 31 Stat. 745). California Salt Company operated here from 1942-1951 (Gale, 1951), and became California Salt Company in 1950 (Ver Planck, 1958). Leslie-Cadiz Salt Company organized in 1924 and became Leslie Salt Company in 1936, currently owned by Carrel Corp. Original placer mining claims were located in 1908, with unreported production of calcium chloride until 1951. 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.

A.J. Salt and Chemical Company produced salt in 1921, followed by Solite Products Company from 1924 to 1936. Ownership or operator names changed to Hallix 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 Plant 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 valentine 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 Bardley 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 epiphanies, 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.
Danyb 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 1862; 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 direct wagons drawn by "locomotive trains" (boiler-powered steam tractor). Salt product was hauled in traction wagons to the Santa Fe Railroad Darty station (now A.T. & S.F.). The A.T. & S.F. rail line route from Cadiz station to Parker and Phoenix alongside Dazby Lake (now leased by the Arizonia and California Railway Company) was not yet in existence. Interestingly, Bailey's report mentions the unconfirmed existence of niter, or "cubic" or "chili saltpepper" discovered and located in 1951 in the playa deposits near the salt beds at Dazby Lake (this "niter of soda" is otherwise known as soda niter, sodium nitrite, or NaNO3). Both deposits are shown at the north end of Danyb 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-Rivas holdings. Rock Salt Products Company (L. W. Redder) produced salt by stripping and blasting at the southern part of the playa near Saltmarsh from 1928 to 1928. Arthur Duran and Tom Schofield produced surface mound deposits of mirabilite (sodium sulfate) west of Saltmarsh, shipping 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. Danyb Salt Corp. and MWD continued operations into the 1960s (Moly, 1965). 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 brines are held by W. C. Redder, F. Reiley and Morton-Thiokol (Morton Salt Company).

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LOCATION

Participants in the 1992 Milovice 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 90 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 NNW-SEE. Among the best known vents are Dish Hill, Siberto Crater, Sunshone Cone, and Mr. Pizazz. Amboy is the easternmost of the cones, and one of the youngest. Three small volcanoes are roughly aligned along the northwestward continuation of the Bristol-Darby tuff, a string of basins containing the saline playas of the same name. The basins may have resulted from tectonothermal activity 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. Confirmed with low silica, alkali undersaturation may occur, leading to the crystallization of kawai in place of potassium feldspar. Given the typically high eruption temperature of alkali basalt lavas, and relatively low crystallization temperature of kawai, this phenocryst is not often found. Some alkali basalt lavas, however, contain abundant leucite phenocrysts. The flows at Amboy contain sparse, small feldsparic and leucite-plagioclase phenocrysts. The silica content of Amboy lava is about 47 weight percent (Park, 1979; 1983; Zoller Hirsch, written communication, 1985). This compare with silica values of 50-52 weight percent for other types of basalt.

Regional study of phase equilibria in the basalt system indicates that alkali basalts must been from partial melting at greater depths than other basalts. Because of the great thickness of continental crusts, basalts erupted in interior continental settings tend to be alkali in composition. Many alkali basalt flows include mare-derived peridotite xenoliths. No such xenoliths been found in the Amboy lava; however, they are abundant at nearby Dish Hill (Whitshire and others, 1985; Whitshire and Nielsen-Pike, 1986).

REGIONAL VOLCANISM

Alkaline eruptions mark the earliest stage of continental rifting, accompanying thermal arching, and the incipient founding of fault blocks along the axis of the arc. This founding marks the site of a future rift valley. In the immature southern East African Rift System, alkaline volcanism is common. With progressive extension of the crust and founding of rift valley floors, the focus of partial melting rises to shallower depths, and ultimately tectonic basaltic characteristic of typical seafloor are erupted.

Apart from geophysical coincidence, it is not clear how the volcanism is related to the Bristol-Darby trough. One might suppose that the trough and its extension toward Baray as an intracraton rift system. But vents locally line up in north-south, or NNE-SSW-trending groups, cutting at high angles across trough-bounding fault 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- to 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 estimate.

A Quaternary alkalic volcanic province very similar to the Amboy-Dish Hill-Paguate group exists in China, where extensional strike slip faulting has also contributed to the development of pull-apart basins (Liu Jiaoqi, 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 Red Grande rift valleys.

AGE

Most young-looking cinder cones in the central Mojave are Pleistocene in age, with fissure-track and K-Ar ages falling around two million years (Whitshire and others, 1985). Basalt flows interbedded with lacustrine deposits deep within the basin are dated are probably associated with even older volcanic events in the Bristol-Darby tectonic province. Basalt is exceptionally young. While some fluvio-lacustrine sediments associated with Bristol Lake cake up against the lava flow, the basalt may be the earliest deposits. The youngest sedimentary layers. If the last significant deposition of sediment occurred in Bristol Lake shortly before the end of the last Ice Age, Amboy Crater would be less than 10,000 years old. If the lake level was high during the glaciation, the volcanic field could be even younger—less than 6,000 years (Park, 1983). Sedimentary 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 in pahoehoe throughout, suggests 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 vescularity of some samples, and the abundance of tumuli (mound-like "gaz 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 raised 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 extended 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 flat-topped 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 overrides wet sediment. The tableland itself may have grown as a perched lava pond, or what 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 rampos. 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 conspicuous 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 Crater 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 cones
resulted. The rims of these intra-cinder 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 zigzags this breaching flow. Climbing to the crater rim, one may find a profusion of lava bombs—most spodic, 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 blast which once occurred here.

A short walk around the volcano 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 contrasts? 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 that the channels resulted from rib-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 IPuka, 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 relieved by spatter agglutinate deposited during an early stage of the eruption. This material is relatively resistant to weathering, and does not dislodge easily during periods of high rainfall. The eastern flank is much less consolidated, however. This material is relatively resistant to weathering, and does not dislodge easily during periods of high rainfall. The eastern flank is much less consolidated, however. This material is relatively resistant to weathering, and does not dislodge easily during periods of high rainfall. The eastern flank is much less consolidated, however. This material is relatively resistant to weathering, and does not dislodge easily during periods of high rainfall. The eastern flank is much less consolidated, however. This material is relatively resistant to weathering, and does not dislodge easily during periods of high rainfall.

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

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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 RAILCYCLE 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 RAILCYCLE 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 continued to operate the line up to the present (Myrick 1963:366,780).

Along this stretch of the railroad, several camps were named in alphabetical order, starting with Amboy and continuing eastward to Bristo, Cadiz, Danby, Essex, Ferrer, Goffs, Homer, Bax, 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 1925 to Bolo (Gulde 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 Burger's 1903 Automobile and Mines' Road Map of Southern California, on Mendellin's 1905 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 Entering the And

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, Shoshone Desert Watering Places, and on the State Railroad Commission's 1926 Official Railroad Map of California.

Several 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 Kistiak 1964:75; Bryant 1974:332,316). 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 1966 USGS 15' Cadiz topographic quadrangle, surveyed in 1954. The toponym "Bolo" had by then been shifted to designate the small hill adjacent to this spot.

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 old 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 old 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 1939 (Boucek 1990:84). The caribbean end 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 southeastally to Cadiz on the railroad bed around the south end of the Marine Mountains. That road had been abandoned for travel purposes (although it continued in use until recenty as a pole-line access road) by 1914, when it was supplanted by an old road originally known as National Old Trails Road and later as U.S. Route 66. The old 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. Seasonal camps such as Bristol/Bombay/Bengal/Boa 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 (Leach 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 Seranno 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 Seranno 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, geese, 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, green 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 (Kroeter 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). Kroeter (1972:38) recorded one stop in the Mojave Salt Song as Mejafu-wa’vawaw-, 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 RAILCYCLE project was guided by a number of research questions designed to address paleoecological 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 epoch? (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 spatially 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 sometime 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. (1959:10-50) 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 flooding."

Recent work by Hardford (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 reflect back 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 to test this possibility resulted in the finding that the "shoreline" feature actually was the result of playa deflation (cf. Blackwelder 1935), 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 RAILCYCLE project site. All but one of these are related to the procurement of lithic resources from the alluvial fan portions of the project site. Nodes of paper, rhyolite, and basalt occurring as flint 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 usable material and were prepared by trimming away the cortical surfaces and removing the resulting core for further reduction into stone tools elsewhere. In these cases, all that remains are the discarded cortical flakes and an occasional hammerstone.

A prehistoric campsite dated on typological grounds to
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 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 Lelio in California (594 m asl) and discharging southward into the Bristo, Cadiz and Daisy Lake basins (the Colorado River) would be over 12,000 km² in size. However, there is a distinct lack of surface and subterranean 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, Bristo, Cadiz and Daisy Dry Lake basins indicate that neither of these connections has occurred during the last 3.7 my. B.P. No distinct paleoshorelines, wave-cut terraces, gravel bars, or paleo mounds have been located at elevations corresponding to the above hydrologic systems in Death Valley, Silver-Soda or Daisy Lake basins. In the Bristo, Cadiz, Silverman 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 last Pleistocene.

INTRODUCTION

For almost a century, regional geology 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): 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 River system draining the San Bernardino Mountains of southern California. Under the present and 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 (e.g., 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 Pleistocene, 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). Deposited 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 Mann Site-experienced several prominent lake stands between ~350 ka. and ~14 ka. B.P., when Afton Canyon was cut and the basin breached (Jefferson, 1985; Meck, 1989). The exact timing and nature of the first overflow of Lake Mann into the Soda-Silver basins remains somewhat unclear; however, detailed reconstructions of late 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...
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 north and east where the basin contains over 670 m of fill (Dickey and others, 1979; Negreti 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) (Moussaig and others, 1979; Dickey and others, 1979; Weits 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 msl, over 300 meters below the current overlow height at Ludlow (Fig. 1) (Wells and others, 1989). Lithologic logs for the three deepest deposits in the Soda Basin are presented in Figure 2. USGS Core 1/2 (236 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 clay characteristic of playa, alluvial fan, and eolian depositional facies (Moussaig and others, 1987). Analysis of these deposits indicate that they are lacking in lacustrine facies assemblages except for the late Lake Mojave deposits of Brown and others, 1979). 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-bloc mountains separated by desert basins (Hunt, 1975). Elevations range from a maximum of over 4400 msl at Mount Whitney, CA (Owens River Drainage Basin) to ~87 msl at Badwater in Death Valley. This prominent relief produces significant variations in temperature, precipitation and evapotranspiration within the study area. Under the current and 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.

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-grayish clastic sediments below the latest Wisconsin 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-n-shallow lake existed in this location during the early-mid-Pleistocene. This "Ancestral Soda Lake" was temporally more persistent, although spatially more restricted 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

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**Figure 1.** Pluvial lakes and playas of the Death Valley-California River region. Insets of Silver-Soda & Bristal dry lakes show location of boreholes and drillcores discussed in text (after Blackwelder, 1954).
Figure 2. Simplified core and borehole logs, Soda Dry Lake, California. Explanation of lithologic symbols given in lower left of figure. Sediments must contain >60% evaporite minerals before they are considered an "evaporite horizon." Fauna columns indicate that lacustrine fauna are present in core. Original data from USGS Soda Core 1/2 and 3 (Havens et al., 1977; original data from SDA Drillhole 3P (Dickson et al., 1979).

(Extraction starts here.)

**Bristol Lake Basin**

A detailed history of the Bristol Lake basin is presented in this guidebook (Rosen, 1992). Six deep cores have been drilled in Bristol Dry Lake (Fig. 1, insets): USGS Bristol Cores 1 and 2, BR-1, BR-2, CAES1, and CAES2 to depths of 306.5, 307, 153, 247, 529, and 537 m, respectively (Basnett and others, 1969; 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 mound-wide lacustrine event or a through-flowing fluvial system in the Bristol basin.
Danby Lake Basin and the Bouse Embayment

Unlike the Bralorne 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 142.2 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 (Lucchitta, 1972) sometime during the mid- to late Pleistocene. Bouse sediments are found in isolated outcrops in the Rhyolite-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 Pleistocene 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 gyttjaiferous 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, 1999) which are interpreted to have been deposited during closed basin lacustrine-pla ya 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, 1999) 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 is the uppermost 50 m of sediment. Rossen (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

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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 "evaporite horizon." Fauna column indicates that lacustrine fauna are present in core. Original data from USGS Cadiz Core 1 and USGS Danby Core 1 and 2 (Bassett et al., 1999).
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 saline seepage 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). In addition, sediments analyzed from between 81.4 and 83 m below the playa surface contain rare representatives of a single foraminifera species, Ammodiscus brevior (P.B. Smith, 1957), 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 Dusty Basin. Evidence for this type of migration is suggested by the presence of the foraminifera Euhelix in a 2 m thick package of sediments (43 m depth) from Panamint Core 1 drilled in Panamint Valley (Fig. 1) (Smith and Pratt, 1957). Extensive studies of the Owens River catchment 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-Dusty basins has not occurred during the last 3.7 ma B.P.

Evidence for an extensive "Death Valley Lake system" (e.g., Hale, 1980) is also lacking. Under the current geographic setting, a lake capable of producing overflow at Ludlow would be over 12,000 km² 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 msl elevation currently needed for overflow, nor have they been found at elevations below this level. Husby and Major (1948), Blackwelder (1954), Miller (1981), and Hale (1985) have argued that the absence of surficial fluvial-lacustrine features is due to flora 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 brine pond conditions since before 3.7 ma B.P. and although the Pleocene Gulf of California Ruser embayment incorporated the Dusty and possibly the Cadiz basins, it was not connected with the Bristol Basin.

Several plausible explanations would account for the similarity of fish faunas 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 coincidently 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, 1954; 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.

**REFERENCES CITED**


* We would like to thank the following individuals and organizations who provided field and monetary support for this research: The New Mexico Water Resources Institute, the University of New Mexico Student Research Allocation Committee (SRAC), the staff of the Desert Research Institute at Soda Springs. In addition, P.J. Meffertinger, G.J. Smith, A. Building, and K. Breats provided helpful insight on the geologic history of the region.
Pleistocene Faunas in the Bristol-Danby Trough

Robert F. Reynolds and Richard L. Reynolds, Division of Earth Sciences, San Bernardino County Museum, Redlands CA 92374

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 Tonbe 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 Fenn Valley (Reynolds, 1991) are found in sediments east of the Bristol Lake and north of Cadiz play, south of Chumash, and north and south of the railroad siding of Cadiz. These distal fluvial deposits were derived primarily from the Fenn Valley drainage and secondarily from the Clipped and Orange Blossom washes. Fenn Valley drains Lantau 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 knolls and layers of pedogenic carbonate. The section is poorly exposed, but the vertical distribution of paleontologic localities suggests that it may be 80 to 100 feet thick. The fauna recovered from the Cadiz localities is given in Table 1.

Faunal Assemblage

The faunal assemblage at Cadiz differs from other Inyoan and Rancholabrean assemblages in southern California. No single taxon stands out as distinctly representing the Inyoan Land Mammal Age (LMA) or Rancholabrean LMA. Boar, an indicator Rancholabrean taxon, and Manisovodon and Mammuthus, common in most Mojave Desert Pleistocene localities, are absent (see Jefferson, 1989 and 1991). Condorine is relatively common in Inyoan 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 (C.T. Jefferson, Page Museum, pers. comm. to BER, 1991; E.J.

Table 1. Cadiz Assemblage

<table>
<thead>
<tr>
<th>Taxon</th>
<th>Abundance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Poidaeus intermedius</td>
<td>small</td>
</tr>
<tr>
<td>Physa sp.</td>
<td>small</td>
</tr>
<tr>
<td>Nyctiphrus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Mysopoma sp</td>
<td>small</td>
</tr>
<tr>
<td>Renuella sp</td>
<td>small</td>
</tr>
<tr>
<td>Triglootis sp</td>
<td>small</td>
</tr>
<tr>
<td>Geocitellus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotonidae sp</td>
<td>small</td>
</tr>
<tr>
<td>Microtus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Cynomys sp.</td>
<td>small</td>
</tr>
<tr>
<td>Urocitellus sp</td>
<td>small</td>
</tr>
<tr>
<td>Peromyscus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Gomphocerus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotona sp.</td>
<td>small</td>
</tr>
<tr>
<td>Arvicanthus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Spicilegus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Marmota sp.</td>
<td>small</td>
</tr>
<tr>
<td>Agricola sp.</td>
<td>small</td>
</tr>
<tr>
<td>Lepus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Jackrabbit sp.</td>
<td>small</td>
</tr>
<tr>
<td>Sylvilagus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Cottontail sp.</td>
<td>small</td>
</tr>
<tr>
<td>Lagomys sp.</td>
<td>small</td>
</tr>
<tr>
<td>Abert's squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Faulty squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotraulus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Spermophilus (Erethizon) sp</td>
<td>small</td>
</tr>
<tr>
<td>S. bicolor sp.</td>
<td>small</td>
</tr>
<tr>
<td>Squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei</td>
<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei sp.</td>
<td>small</td>
</tr>
<tr>
<td>Perognathus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Perognathus sp. (lrg)</td>
<td>small</td>
</tr>
<tr>
<td>Lepus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotona sp.</td>
<td>small</td>
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</tr>
<tr>
<td>Urocitellus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Peromyscus sp. (lrg)</td>
<td>small</td>
</tr>
<tr>
<td>Peromyscus sp. (med)</td>
<td>small</td>
</tr>
<tr>
<td>Abert's squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Faulty squirrel sp.</td>
<td>small</td>
</tr>
<tr>
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<td>small</td>
</tr>
<tr>
<td>Spermophilus (Erethizon) sp</td>
<td>small</td>
</tr>
<tr>
<td>S. bicolor sp.</td>
<td>small</td>
</tr>
<tr>
<td>Squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei</td>
<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei sp.</td>
<td>small</td>
</tr>
<tr>
<td>Perognathus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Perognathus sp. (lrg)</td>
<td>small</td>
</tr>
<tr>
<td>Lepus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotona sp.</td>
<td>small</td>
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<tr>
<td>Ochotona sp. (lrg)</td>
<td>small</td>
</tr>
<tr>
<td>Urocitellus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Peromyscus sp. (lrg)</td>
<td>small</td>
</tr>
<tr>
<td>Peromyscus sp. (med)</td>
<td>small</td>
</tr>
<tr>
<td>Abert's squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Faulty squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotraulus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Spermophilus (Erethizon) sp</td>
<td>small</td>
</tr>
<tr>
<td>S. bicolor sp.</td>
<td>small</td>
</tr>
<tr>
<td>Squirrel sp.</td>
<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei</td>
<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei sp.</td>
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</tr>
<tr>
<td>Perognathus sp.</td>
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</tr>
<tr>
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<td>small</td>
</tr>
<tr>
<td>Lepus sp.</td>
<td>small</td>
</tr>
<tr>
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<td>small</td>
</tr>
<tr>
<td>Ochotona sp. (lrg)</td>
<td>small</td>
</tr>
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<td>small</td>
</tr>
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<td>small</td>
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</tr>
<tr>
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<td>small</td>
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</tr>
<tr>
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<td>small</td>
</tr>
<tr>
<td>Quoymyys bottei</td>
<td>small</td>
</tr>
<tr>
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</tr>
<tr>
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</tr>
<tr>
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<td>small</td>
</tr>
<tr>
<td>Lepus sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotona sp.</td>
<td>small</td>
</tr>
<tr>
<td>Ochotona sp. (lrg)</td>
<td>small</td>
</tr>
<tr>
<td>Urocitellus sp.</td>
<td>small</td>
</tr>
</tbody>
</table>
Lindsay, Univ. of Arizona, pers. comm. to RER, 1992), although Rancholabrean records are common in Florida and Texas (Fairbanks, 1968; Holman, 1969, 1970; Milstead, 1986; Moodie and VanDevender, 1979).

The grackle (Quiscalus major) resembles a form found in the late Pleistocene deposits of Marinette (Reynolds and Reynolds, 1940; Reynolds and others, 1992). The medium-sized leporid is similar morphometrically to specimens from deposits considered to be Pleistocene LMA in the west central Mojave Desert (Reynolds, 1989).

The small proterornithine Carpopteryx and Tetragnyx are both known from the late Pleistocene LMA in southern California, but Tetragnyx is rare in late Rancholabrean deposits (Kent and Anderson, 1968; Savage and Russel, 1968). The medium-sized solen is an unclassified bat not known from the Ranch La Brea deposits (Jefferson, pers. comm. 1992).

In general, the fauna from Cadiz suggests an early Rancholabrean or late Pleistocene LMA. The presence of Psalidion cerasus suggests the presence of standing water; the crickets Neotoma and Peromyscus (small and large), and the Sphenictes 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 - 42.5**

**Location and Description**

The Archer localities are approximately 8 miles southeast of Cadiz on the southern flank of the Ship Mountains, at an elevation similar to that of sediments at Cadiz. The Archer

<table>
<thead>
<tr>
<th>Plantae</th>
<th>root casts</th>
</tr>
</thead>
<tbody>
<tr>
<td>Salvia sp.</td>
<td>sand small</td>
</tr>
<tr>
<td>Phytophora sp.</td>
<td>hornted beaked</td>
</tr>
<tr>
<td>F. plethoraphis</td>
<td>flared hornted beaked</td>
</tr>
<tr>
<td>Boidea</td>
<td>box</td>
</tr>
<tr>
<td>Calotomodon</td>
<td>colubrid snakes</td>
</tr>
<tr>
<td>Aves sp.</td>
<td>bird</td>
</tr>
<tr>
<td>Aves sp.</td>
<td>bird</td>
</tr>
<tr>
<td>Passerina cyanne atrine</td>
<td>common bunting</td>
</tr>
<tr>
<td>Passerellus xanthippus</td>
<td>Swainson sparrow</td>
</tr>
<tr>
<td>Colaptes mexicanus</td>
<td>finch</td>
</tr>
<tr>
<td>Vesperiseralis</td>
<td>small but</td>
</tr>
<tr>
<td>Lupins sp.</td>
<td>jack rabbit</td>
</tr>
<tr>
<td>L. californica</td>
<td>black-tailed jack rabbit</td>
</tr>
<tr>
<td>Sylvisia sp.</td>
<td>cottontail</td>
</tr>
<tr>
<td>Thosinus borowiei</td>
<td>Bottes pocket gopher</td>
</tr>
<tr>
<td>Perognathus sp. (ligg)</td>
<td>large pocket mouse</td>
</tr>
<tr>
<td>Procyonius sp.</td>
<td>pocket mouse</td>
</tr>
<tr>
<td>Dipodomys sp.</td>
<td>kangaroo rat</td>
</tr>
<tr>
<td>D. ordi</td>
<td>Ord's kangaroo rat</td>
</tr>
<tr>
<td>Peromyscus sp.</td>
<td>deer mouse</td>
</tr>
<tr>
<td>Neotoma sp.</td>
<td>wood rat</td>
</tr>
<tr>
<td>N. of. N. lepida</td>
<td>desert wood rat</td>
</tr>
<tr>
<td>Rathineumonimagaeblerisi</td>
<td>harvest mouse</td>
</tr>
</tbody>
</table>

Sediments are ichnologically distinct, however, perhaps due to their proximity to the Ship Mountains. The section consists of approximately 4 feet of sands and angular gravel casts which fine upward into an iron oxided-brown horizon with mottles and carbonate kernels 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 deposits, and the overlying carbonate layer indicates hydrologic regimes that were not active in recent times.

**SALTMARSH ASSEMBLAGE**

**SBCM 1: 44.1 - 44.8**

**Location and Description**

The Salt Marsh locality is in Wind Valley, on the northeast margin of Dusty Lake, approximately 8 miles northwest of Highway 62 and 9 miles southwest of the site of Milligan. Sediments consist of tan playa silts overlain by a soil containing carbonate kernels 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 Camels (sp. and small horse are extinct taxa typical of Rancholabrean LMA (fauna 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>
<thead>
<tr>
<th>Plantae</th>
<th>root casts</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lupins sp.</td>
<td>jack rabbit</td>
</tr>
<tr>
<td>L. californica</td>
<td>black-tailed jack rabbit</td>
</tr>
<tr>
<td>O. oxpekenus</td>
<td>kangaroo rat</td>
</tr>
<tr>
<td>D. ordin</td>
<td>Ord's kangaroo rat</td>
</tr>
<tr>
<td>Peromyscus sp.</td>
<td>deer mouse</td>
</tr>
<tr>
<td>Neotoma sp.</td>
<td>wood rat</td>
</tr>
<tr>
<td>N. of. N. lepida</td>
<td>desert wood rat</td>
</tr>
<tr>
<td>Rathineumonimagaeblerisi</td>
<td>harvest mouse</td>
</tr>
</tbody>
</table>

*Saltmarsh Assemblage*
The relative elevations of fossil localities in the Bristol-Dunby trough compared to present-day playa surfaces are shown in Figure 3. The Salmarsh locality is at an elevation of 600' and is not more than 20 feet above the surface of Dunby playa. Dunby Lake is separated from Cadiz and Bristol playas by the metamorphic rocks of the Kilkbee Hills. Salmarsh 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.

ACKNOWLEDGEMENTS
The authors thank the following for their assistance in this work: Christopher J. Bell (SBGM) for identification of the Bryoisopoda; Dr. John Ford (UPL) for satellite imagery; Dr. Steve Wells (U.C. Riverside) for helpful comments; and all the field and laboratory crew in the SBGM Department of Earth Sciences for their participation in the collection and preparation of these specimens over the last 7 years.

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Geology and Saline Resources of Danby Playa, Southeastern California

J. F. Calzia, U.S. Geological Survey, MS 901, Menlo Park CA 94025

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 yr 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.

GEOLGY

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 (Gribb on Map 1), yet low area is often covered with a white saline efflorescence and/or 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 southeast; 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, 1933). 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, 1933), 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, clary cilt, silt, and sandy clay, and massive clay. gypsum and halite are present throughout this sequence (Bassett and others, 1959; Calzia, 1991). The lacustrine deposits are 280 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 heterogeneous 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, 1954).

Smith (1960, 1970) described formations associated with marine and brackish water mollusks, barnacles, and ostracods in a thick (500 ft) interval of sand, calcareous silt, and laminated blue and green clay 40 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) Boure Formation. Northwest-trending gravity profiles across the playa suggest that the relatively low density lacustrine deposits and the underlying Boure 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$$\cdot$2H$_2$O), halite (NaCl), and chloride-dominated brines. Gypsum is present as disseminated crystals and as several thick (more than 200 ft thick) beds in the lacustrine deposits; halite is found in clay-rich beds in the lacustrine deposits and as coarse-grained crystals in large, discontinuous crystalline salt beds (Bassett and others, 1959; Ver Planck, 1957, 1958). Relatively rare basanite (CaSO$_4$•5H$_2$O) and mirabilite (Na$_2$SO$_4$•10H$_2$O) 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 5 to 10 ft thick (Fig. 1). Assuming a density of 2.16 g/cm$^3$, this salt body contains 13.75x10$^9$ tons of salt at an average grade of 92.04% 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$^9$ tons of salt at an average grade of 73.84% NaCl; the northwest salt body contains 1.5-3x10$^9$ 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 lower calcium chloride than evaporated brines from Brawa and Cadis 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); contour interval is 2 ft.
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(a) As CaCO3.
(b) Total dissolved solids at 180°C.
the salinity content decreases to the southeast across Danby Plaza. 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 (Calif., 1971). For example, brines in the northwest area contain approximately 25 percent NaCl and generally less than 2 percent NaSO\(_4\) (Table 1); brine from a test well along the northeastern side of the playa contains less than 5 percent total dissolved solids (Ver Planck, 1959). This spatial relation suggests that the brines are produced by the interaction of crystalline salt deposits with groundwater impounded 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 Plaza 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 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. Gundy, geologist, BLM California Desert District, personal communication, 1992).

REFERENCES CITED


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 northeastern 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 Jinks, 1986; Knoll, 1986; Nelson, 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 palaeocurrent 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 braided plains emanating from the Old Woman Mountains highland (Knoll, 1986).

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 (Nelson and others, 1990). The basin unconformity is defined by coarse sandstones and conglomerates resting on an erosional surface developed on Cretaceous granodiorite. Paleozoic metamorphic rocks, or Precambrian gneiss (Fig. 3). A variation in thickness of the pre-Peach Springs Tuff section of 600 m attests to considerable palaeotopography on the original depositional surface. An approximately 20 m thick volcanic section composed mainly of alkali, olivine-bearing basalt flows, subordinate interbedded volcaniclastic sandstones and lahars, and rare anodesite 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 17.6 ± 0.7 Ma Wild Horse Mesa Tuff of the Providence Mountains to the north (McCurry and others, 1989).

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
The only reliable age constraint for this unit is provided by the 18.5 ± 0.2 Ma Peach Springs Tuff, which lies 115 m upslope. The unit lies 10.5 m above Paleozoic meta-sedimentary basement rocks. Knoll (1986) estimated that Initial 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 Plute 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 1.

Table 1. Little Plute Mountains Assemblage

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<tr>
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<td>Michaelia sp. cf. M. atactica</td>
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</tr>
<tr>
<td>Maryophoca sp. fami</td>
<td>horse</td>
</tr>
</tbody>
</table>

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

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

Figure 3. NE - SW cross section through the southern Little Plute Mountains.
Figure 4. Sketches of tracks of "footprint gully." (a) antopectolit (caudifur); (b) caudidant?

Milieuose agostica is known to occur in Hemifordian LMA deposits in the Caday Mountains (Miller, 1987) and in the Crowder Formation (Reynolds, 1991a). The genus was described by Frick and Taylor (1971) who indicate it is known from early Hemifordian to late Clarendonian times. *Hemfordia antiqua* is known from the Hemifordian and Barstovian deposits in Cajon Pass and at Barstow (Davidson, 1923; Miller, 1980; Reynolds, 1991b). *Pyriformus* furcatus is known from Hemifordian and Barstovian deposits in and around the Mojave Desert (Reynolds, 1991b; Lindsay, 1977; Whistler, 1991a, 1991b). Paleontoloy *Yakusakia* is known from the Cajon and Crowder formations (Reynolds, 1991a) and from Barstow (Lindsay, 1977, and its usefulness as a Hemifordian LMA indicator is discussed by Reynolds (1991b).

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

CONCLUSIONS

The vertebrate fossils and tracksways exposed in sedimentary deposits of the Little Putee Mountains provide a rare glimpse of Miluoie, Hemifordian LMA faunal assemblages in the eastern Mojave Desert. The presence of grazing animals suggests a palaeoenvironment with grasslandly shrub 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 World War II military airport 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 Haakasilma, Chemehuevi, Mojave and other native peoples into the present historical era. These tribal groups hunted bighorn sheep and other game, establishing camps at Mohap 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 Vital Valley, a remnant 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 unproven. 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 tableing, and multiplex air photogrammetry (Turtle Mountain Quadrangle, 1:62,500). By this time the area lay under the administration of the U.S. Bureau of Land Management.

While prospecting has 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 perite glass from vivic 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 has 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.)
The earliest large-scale geological reconnaissance was done in the late 1950s by railroad company geologists (Cooksey, 1960a, b; Conradi, 1960) and by R.B. Saul (1963), gathering information for the California State Geologic Map. Embers (1967) undertook the first truly detailed geologic mapping, returning to Chetserman'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 Edition Company (Woodward McNell and Associates, 1974). A nuclear power plant had been proposed for development in Vidal Valley, between the Mopah Range and the main bedrock mass 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 volcanism—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 (e.g., Howard and others, 1982; Nelson 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, located in the Turtle Mountains, detailed studies were also continuing in the nearby Whipple and Chermehuvi 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 nonconformably overlying detached basement, all extended by rotation along northeastward-dipping thrust normal faults. The work in the Turtle Mountains made it clear the Whipple Fault continued westward beneath this range, which, in fact, could be constructed as lying in the headwall terrane of the fault. Industry seismic profiling with CALCROUST 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 (Chaya and Frost, 1989). The entire Turtle Mountains therefore is part of a detached allochthon of great regional extent (fig. 2).

GEOLGY
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 hillocks 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-Proterozoic granitic apophyses and veins, and Cretaceous granitic plutons. Retension 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. Pollutions 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 siliceous 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 migmatic fabric in the terrane is an indication of the extreme level reached by metamorphism, and suggests that many intrusive bodies are locally derived. Paleodetritus 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 sand and 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 dikes intrusions, which may have fed some of the
black flows making up the surrounding flat-topped ridges and mesas. Farther east, on the approach to Vidal junction, dark cuestas of subalkali to alkali basalt flows may be seen marking the eastern edge of the Moplah 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 Moplah Range southern Turtle Mountains took place between 22.5 and 14.5 Ma (Hazlett, 1990). Detachment faulting occurred during much, if not all of the 65 million years of deformation, resulting in development of extensive angular unconformities within the volcanic section. Plugs and dikes are aligned parallel to and concentrated within the most intensely faulted areas, showing an intimate coalescence between extensive faulting and magmatism. Composition of volcanics ranges unimediately from mafic-alcaline alkali 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 southwestern Basin and Range; an early maic-to-alkali period of activity, often explosive, preceded a later-stage burst of effusive maic activity. An interval of unexposed duration separated the two pulses of activity, marked structurally by a prominent unconformity. Preservation of vent structures indicates that volcanism took the form of eruptions from clustered cinder cones and domes, not unlike what is presently seen in the much younger Coso Volcanic field near Chula Lake.

The faulted volcanic terrane making up the Moplah segment across Highway 95, north of Vidal Junction, into the western Whipple Mountains. Here volcanic strata crop up 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.

**REFERENCES CITED**


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 a beautifully displaying dramatic landforms and striking color contrasts between different rock types. Studies of sedimentary strata in this region indicate that the present-day tectonics 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 (Cisne, 1980) in areas that experienced large amounts of Cenozoic crustal extension. Core complexes are characterized by: (1) dimal 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 hydromicas; and (4) upper plate rocks that are highly deformed by closely-spaced normal faults (Davis and Lister, 1990).

The Whipple detachment fault has an overall northwestward dip, and extension occurred as the lower plate moved up and out to the southwest, away from the Colorado Plateau.

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 tectonic 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 episodic. 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-extensional basins are now exposed in the southern and eastern Whipple Mountains (Fig. 2). These rocks are dominated by braided stream and sheetflow deposits. In the eastern part of the area, these rocks are associated with monothetic breccia beds deposited by rock avalanches and mass flows. Lutetian 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).
STRATIGRAPHY

Gene Canyon Formation

The Gene Canyon Formation is found only in the eastern part of the study area and is subdivided into three members, which from oldest to youngest are the Girs Wash Member, the Detrital Wash Member, and the Gene Wash Member.

The Girs Wash Member is characterized by dramatic lateral and vertical facies changes, lack of volcanoclastic rocks as primary deposits at the base, and by the presence of melilitoholoclastic 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 sheetflow deposits with subordinate matrix-supported mass flow deposits and small lenses of limestone. Cobble- to small boulder-bearing, sand- to matrix-supported strata, interpreted as mass flow breccias, constitute the Buckskin Mountain State Park section (Fig. 3b) with subordinate sheetflood sandstones and conglomerates interpreted as sheetflood deposits. Clast types include Plahurovskyi and Cretaceous (?) granulites and gneisses, and metasedimentary rocks including quartzite, phyllite, and white marble.

Table 1. Summary of the defining characteristics of the stratigraphic units used in this study.

<table>
<thead>
<tr>
<th>MEMBER</th>
<th>TYPE</th>
<th>POSITION</th>
<th>CHARACTERISTICS</th>
<th>CORRELATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Girs Wash</td>
<td>Member</td>
<td>Volcanic flows and tuff with interbedded sandstone and conglomerate</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
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 streambed deposits. In contrast, laterally persistent beds of thinly bedded fine-grained sandstone and siltstone, interpreted as playa deposits, and interbedded streamflow deposits delineate in the Buckskin Mountain State Park section (Fig. 3). Clasts within the Desilt Wash Member were derived predominantly 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 Giers Wash Member is distinguished from underlying units by the presence of thin, strongly altered, nuface to intermediate lava flows with interbedded coarse-grained sedimentary strata. A rhynolitic ash-flow tuff that tentatively has been identified as the Peach Springs Tuff caps the unit (Hobbs, 1990). The lava flows are sparsely porphyryic, containing about three to five percent small ovalized 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, Nough 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 streambed 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 potash metasomatism. The formation ranges from about 100 to 500 m thick, and sedimentary lobes and beds 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 calcite micrite, which contains abundant algal material. The siliciclastic
content of the limestone is variable but generally high. Mottled conglomerates are composed of subangular to subrounded mudstone and carbonate intraclasts. Secondary silica is a common and characteristic component of the Twin Lobe Mine Formation, resulting from selective replacement of calcite by microcrystalline quartz and chaledony 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 Gove 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 lithologically diverse. Crude stratification and large-scale, low-angle cross-bedding are the most common sedimentary structures. These sandstone and conglomerate beds are interpreted as streamflow 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 Wrangell fault. Tertiary volcanic rocks including the Peach Springs Tuff, Tertiary limestone, and quartzite. Tertiary volcanic rocks are the most abundant clast type at 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 Lobe Mine Formation. The unit is characterized by sandstone, ranging from well-bedded, fine-grained sandstone and siltstone interpreted as interbedded sandstone and siltstone (Walker and Mutti, 1977) to well-bedded, medium sandstone with rippled surfaces. In general, these deposits exhibit an overall cross-bedding and cross-lamination sequence. Locally, interbedded sandstones and conglomerates interpreted as sheetflood deposits overlies the Twin Lobe Mine Formation.

**Osborne Wash Formation and Bureau Formation**

The tuff, synextensional scoria in the eastern and southern Whipple Mountains, is unconformably overlain by alkali-olivine basalts and interbedded fanglomerates and sheetflood deposits belonging to the Osborne Wash Formation, and siltstones and muds of the Bureau 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 Rb-Sr isochron of 125.5 + 3 Ma (Stein) was obtained from a lacustrine tuff located near the base of the Tertiary section in the Aubrey Hills, just north of the study area (Benison et al., 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 younger age limit for the Copper Basin Formation. The accepted age of this widespread tuff is 18.3 + 0.6 Ma (Aliarth, 1990; Leidner et al., 1990). A sequence of basin broads 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; Nickelsen and Burdick, 1980).

**LANDFORM DEVELOPMENT**

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

1. The bulk of the range consists of a distinct topographic high, primarily composed of mylonitic gneisses in the lower plate. Synkinematic 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 structural uplift due to non-uniform tectonic denudation (Spencer, 1982).

2. Prominent northeast-trending ridges flank the range, held up by resistant Early to Middle Miocene andesite flows. (Turf Mine Formation) and strongly cemented sandstones and conglomerates (Copper Basin Formation). The strata are tilted 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 north-northeast trending high-angle normal faults to form half-graben basins occurred during these times during the detachment event; the ridges observed today resulted from the final tilting episode at about 14 Ma (Nielsen and Burdick, 1989; Benison, 1991).

3. The area between the ridges consists of low, irregular hills separated by poorly interconnected washes. Three rubble hills formed on the highly fractured upper-plat do and gneisses.

4. A plateau capped by nearly horizontal tuff-Tertiary basin 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 subduction of detachment faulting (Buising and Burdick, 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 alluvial deposits. Three rubble hills formed on the highly fractured upper plate and gneisses.

6. Tectonic activity of the extensional and contractional regimes is a factor in the development of the Colorado River's present course (Buising, 1991).
ACKNOWLEDGEMENTS

This paper is a modification of Beratan (in press); the landform history was presented at the 1991 Mojave Desert Quaternary Research Symposium meeting (Redlands). Most of this work was done as part of a Ph.D. dissertation at the University of Southern California; further work was done at the Jet Propulsion Laboratory/California Institute of Technology. Thanks to Dr. Gregory A. Davis for suggesting the project, and to Dr. Robert H. Osborne for his advisement. Special thanks to Dr. Jane E. Nielsen for her guidance and support; thanks to her also for her revisions to figures 2 and 3.

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

Anna V. Buliung, Department of Geological Sciences, California State University, Hayward CA 94542-3086

Mio-Pliocene sediments of the lower Colorado River area represent the northernmost well-documented extent of the pre-Gulf of California, a technically 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 my., prior to the onset of spreading- and transform-related subsidence in that area. Fanglomerate and volcanic rocks informally referred to as the Osborn Wash strata interdigitate with the conformably overlying Bouse Formation. The Bouse Formation includes carbonate and coarse terrigenous-clastic basin margin deposits that interdigitate laterally with basin fill material; basin fill strata comprise a basal marlstone-carbonate unit overlain by fine- to medium-grained terrigenous-clastic detrital deposits. The upper portion of the Bouse interdigitates with overlying cobble conglomerates referred to as the Colorado River gravel units. 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 extention in two directions in the sediment pile overlying the incising basin floor. The Bouse Formation and bracketing units record three stages in the evolution of the northern proto-Gulf/lower Colorado River area: (1) dissection of preexisting, detachment fault-controlled topography and localized, inter-drainage fluvial 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 My); (3) progradation of ancestral Colorado River delta into the northern end of the proto-Gulf basins (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 Osborn 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 downdropping, 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.
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: • WPO1—Yuma Mammoth (Blake, 1990; Hoy, 1927)
• USGS-LCRP 4-23-2—Elyeckan Tank Locality (Metzger, personal communication 1964; this locality was published at a later date in Metzger and others, 1972 and Bell and others, 1970)
• JSNI Bill Williams Fork (Newberry, in press; 1861; Hoy, 1927)
• JSNI Elephant Hill (Newberry, in loc., 1866; Hoy, 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 molars from Topock, Arizona (Mc. McShan, personal communication 1991).

In 1971, a nearly complete cranial of a mammoth was discovered by two residents of the Colorado River Indian Tribe Reservation near Baker, 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 heavily eroded. Excavated in fine clays overlying a conglomerate, the individual was preserved in a position indicative of settling on its back in shallow water. Abundant impressions 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 a 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 Metaxyphas meridionalis, based on dental measurements. (There is overlap in most dental criteria for some of the

Figure 1. Recorded mammoth localities, Colorado River Trough. Numbers indicate descending sequence in Table I.
Table I. Recorded Mammoth Localities for the Colorado River Trough

<table>
<thead>
<tr>
<th>Location</th>
<th>Site Number</th>
<th>Reference</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>El Guillo, Sonora</td>
<td>J-ACM 118453</td>
<td>Shaw, 1983</td>
<td>Mammothus primus bx remait petite fragments, 1</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>astragalus, 1 humerus, 1 femur frag., 1 vertebra,</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1 vertebra, 11 vertebrae, 12 ribs, 1 sacrum.</td>
</tr>
<tr>
<td>Yuma Co., AZ</td>
<td>WBP1</td>
<td>Blake, 1900</td>
<td>Yuma, AZ, Mastodon w/ teeth, lower molars, bone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(Saunders #20, 1970)</td>
<td>fragments, 8 vertebrae, 4 rib fragments.</td>
</tr>
<tr>
<td>La Paz Co., AZ</td>
<td>USGS 1-LCP 4-34-5</td>
<td>(Saunders #37, 1970)</td>
<td>Valley fill along Colorado River.</td>
</tr>
<tr>
<td>Mohave Co., AZ</td>
<td>JSN-1</td>
<td>Saunders #38, 1970</td>
<td>Elephant Tank Locality. 380' elev.</td>
</tr>
<tr>
<td>Mohave Co., AZ</td>
<td>JSN-2</td>
<td>Saunders #38, 1970</td>
<td>Elephant Tank, 480' elev. Single tooth 12 miles</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>southwest of Los Angeles. 1882.</td>
</tr>
<tr>
<td>Mohave Co., AZ</td>
<td>SBCM 3-3-3</td>
<td>Needles Museum, 1974</td>
<td>Mohave Canyon N.</td>
</tr>
<tr>
<td>Mohave Co., AZ</td>
<td>SBCM 2-6-21</td>
<td>Needles Museum, 1974</td>
<td>Needles, California.</td>
</tr>
<tr>
<td>San Bernardino</td>
<td></td>
<td></td>
<td>Petroglyphs.</td>
</tr>
<tr>
<td>Co., CA</td>
<td></td>
<td></td>
<td>Petroglyphs.</td>
</tr>
<tr>
<td>Riverside Co., CA</td>
<td>CRF skull</td>
<td>Colorado River, California. west of Pinto, AZ.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Nearby complete skull.</td>
</tr>
<tr>
<td>Clark Co., NV</td>
<td>NAU-QSP 9156</td>
<td>Lake Mead NRA, NV.</td>
<td></td>
</tr>
</tbody>
</table>

DEPOSITIONAL OCCURRENCE

The localities that provide data on the geologic depositional units containing the mammoth remains can be traced generally assigned to the Chimpheuvi Formation (the Chemehuevi Gravel of Lee, 1914; Longwell, 1936). These deposits appear to be relatively continuous from Lake Mead to Imperial Dam, north of Yuma (Metzger and others, 1972). They have been attributed to two different depositional origins: lacustrine and fluvial (Longwell, 1936). Longwell (1936, 1946, 1954, 1960, 1965) favors a ponding (lacustrine) environment of deposition; however, he can account for no suitable base level (natural dam). Metzger and others (1972) 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, "staging 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 Cheemuhevi Formation, as a sedimentary unit consisting of clay, silt, and sand, "laid down in the old lake." He describes its blackish clay and silt that share contact with the underlying Muddy Creek Formation in the Virgin and Muddy Valleys. Near Calvile, Longwell (1936) describes the Cheemuhevi Formation as "two distinct facies: 1) massive to thin-beded 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 caliche ranges from 30,000 to 90,000 BP. Johnson and Miller (1970) discuss the post-Cheemuhevi depositional sequence.

Colorado River sediments near Parker, Arizona, which postdate the Rouge 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 Cheemuhevi Formation at Needle and Blythe.

**CHRONOLOGY**

Few radiometric dates are available for the Cheemuhevi Formation or its enclosed fossils. Bell and others (1978) cite a uranium-thorium date on the Ehrenberg tusk from the Cheemuhevi Formation (unit D of Metzger and others, 1973) as 102,000 ± 200 BP. To our knowledge, no other proboscidean material from the Colorado River Corridor has been subjected to a radiometric age analysis.

Longwell (1978) notes that the mammal horizon occurs in an upper sandy unit near Calvile. Indicative of a Rancholabrean Land Mammal Age for an upper unit of the Cheemuhevi 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 formation.

Based on the sparse information outlined above, the Cheemuhevi 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 Cheemuhevi Formation is pre-Wisconsin (100,000-200,000 BP). However, the presence of M. meridionalis suggests that deposition of the lower portion of the Cheemuhevi 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 Sonoita to Lake Mead has tripled. All southern species for North America are represented: at least three M. meridionalis specimens are known from the Santa Cruz and Arizona portions of the corridor; one M. imperator is known, from El Golfo, Sonora; several M. columbi specimens are known; and there are 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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McKee, Mary, 1974. Huge holes and teeth found in valley. Footprints, 2:16a-b.


The San Bernardino County Museum Association publishes articles and monographs on subjects pertaining to the cultural and natural history of San Bernardino County and surrounding regions. We welcome submissions of such manuscripts.

Subject Matter: articles and monographs pertaining to San Bernardino County, Inland Southern California, and surrounding regions, in history, anthropology, archaeology, paleontology, mineralogy, zoology, botany, ichnology, and related disciplines. Manuscripts considered for Quarterly publication should be written toward the well-educated non-specialist. Technical research will also be considered for publication. All manuscripts should reflect original work which further knowledge in their fields.

Format: Two clear copies of the manuscript must be submitted to the Editorial Board with a letter of transmittal requesting that the manuscript be considered for publication and that it is not presently under consideration elsewhere. Manuscripts should be typewritten, double-spaced, on one side only of 8.5x11" paper. Ample margins should be allowed for editing comments. The first page should contain the title and author(s) name, address, and telephone number. The author’s last name and page number should appear at the top of each following page. Include COPIES of figures, tables, and photographs. Do not send original photographs or figures with your initial submission.

Style: Authors should follow the standards for footnotes, citations, headings, and other conventions as applicable to their discipline. The Editorial Board suggests the following:

- Anthropology/Archaeology: Society of American Archaeology (American Antiquity)
- History: American Historical Association (American Historical Review)
- Geology: Geological Society of America (GSA Bulletin)
- Paleontology: Society of Vertebrate Paleontology (Journal)
- Biological Sciences: American Institute of Biological Sciences (e.g., Journal of Entomology)

Authors should be aware of and avoid inappropriate gender-biased language. The Editor is available for consultation on matters of style, format, and procedures.

Review: Manuscripts will be considered by the Editorial Board of the Museum Association Publications Committee, and will be reviewed by outside experts. Manuscripts may be accepted, provisionally accepted, or be found unsuitable for publication by the Association. Provisional acceptance may include suggestions for revisions. Very lengthy or profusely-illustrated monographs that are otherwise acceptable for publication may require outside funding to help defray publishing costs. Manuscripts will be copy edited after acceptance.

Attachments: Original or equivalent photographs will be required for publication. Photographs should be black-and-white, glossy finish, of good quality and contrast. Figures and drawings should be in India ink or equivalent on white paper or film; P.M.T. are acceptable. Captions should be submitted on separate pages, double-spaced, and referred to their accompanying figures. Photographs should be marked lightly in pencil on the back border with the author’s name and figure number. Authors are encouraged to submit accepted manuscripts on DOS-compatible disks in addition to paper copy.

Responsibilities: The author has the primary responsibility for the correctness and reasonableness of his or her information, arguments, and presentation. In submitting a manuscript for consideration, the author assures the Editorial Board that the manuscript is original work and does not infringe upon the rights of previous authors or publishers.

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